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ANDREW S. GOUDIE

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Foreword

As president of the International Association of Geomorphologists (IAG), a body that seeks to provide a forum for the promotion of geomorphology internationally, I am delighted that, in association with Routledge, it has been possible to produce this great encyclopedia. It is written by contributors from some thirty countries, all of whom have generously agreed that their royalties should go to the IAG. This will add substantially to the financial resources of the Association. The IAG is grateful to the editorial team, and in particular to Andrew Goudie, for the work they have done to bring it to fruition and I am sure that it will be an invaluable resource for the international geomorphological community.

Mario Panizza
Modena, Italy
October 2002

Preface

The term 'geomorphology' arose in the Geological Survey in the USA in the 1880s and was possibly coined by those two great pioneers, J.W. Powell and WJ McGee.

In 1891 McGee wrote: 'The phenomena of degradation form the subject of geomorphology, the novel branch of geology.' He plainly regarded geomorphology as being that part of geology which enabled the practitioner to reconstruct Earth history by looking at the evidence for past erosion, writing:

> A new period in the development of geologic science has dawned within a decade. In at least two American centres and one abroad it has come to be recognised that the later history of world growth may be read from the configuration of the hills as well as from the sediments and fossils of ancient oceans… The field of science is thereby broadened by the addition of a coordinate province – by the birth of a new geology which is destined to rank with the old. This is geomorphic geology, or geomorphology.

Of course, many scientists had studied the development of erosional landforms (see the magisterial history of Chorley *et al*. 1964) before the term was thus defined and since that time its meaning has become broader. Many geomorphologists believe that the purpose of geomorphology goes beyond reconstructing Earth history and that the core of the subject is the comprehension of the form of the ground surface and the processes which mould it. In recent years there has been a tendency for geomorphologists to become more deeply involved with understanding the processes of erosion, weathering, transport and deposition, with measuring the rates at which such processes operate, and with quantitative analysis of the forms of the ground surface (morphometry) and of the materials of which they are composed. Geomorphology now has many component branches and involves the study of a huge range of phenomena.

In 1968 Rhodes W. Fairbridge edited a large and invaluable encyclopedia of geomorphology that explored this diversity. However, geomorphology has changed greatly since that time, not least because of the plate tectonics paradigm, the revolution in our knowledge of the Quaternary Era brought about by new dating and environmental reconstruction techniques, the development of modelling and systems thinking, appreciation of the importance of organisms, application of geomorphology to the study of engineering problems and global change, a greater appreciation of the nature of geomorphological processes, and availability of a whole range of new technologies for analysis of data and materials, the development of satellite-borne remote sensing, and the exploration of space.

Over that time, due to the inspiration of the people to whom this book is dedicated, Geomorphology has for the first time organized itself internationally so that the geomorphological traditions that have grown up in different countries (see Walker and Grabau 1993) can interact as never before. It was therefore felt at the International Geomorphological Congress in Tokyo in August 2002 that the International Association of Geomorphologists (itself officially founded in 1989) should seek to publish a new and truly international Encyclopedia of Geomorphology that could survey the nature of the discipline at the turn of a new millennium. I am indebted to my Consultant Editors and the contributors from some thirty or so countries, who have made this endeavour possible.

Andrew S. Goudie
Oxford

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Thematic entry list

Aeolian

Adhesion
Aeolation
Aeolian geomorphology
Aeolian processes
Aeolianite
Aligned drainage
Barchan
Beach-dune interaction
Bedform
Bounding surface
Deflation
Desert geomorphology
Draa
Dune, aeolian
Dune, coastal
Dune mobility
Dune, snow
Dust storm
Glaciaeolian
Interdune
Loess
Lunette
Nebkha
Niveo-aeolian activity
Pan
Parna
Ripple
Saltation
Sand ramp
Sand sea and dunefield
Sandsheet
Sastrugi
Singing sand
Stone pavement
Ventifact
Wind erosion
Wind tunnels in geomorphology
Yardang

Biogeomorphology

Beach rock
Biogeomorphology
Biokarst
Boring organism
Brousse tigrée
Coral reef
Corniche
Crusting of soil
Desert varnish
Forest geomorphology
Large woody debris
Mangrove swamp
Microatoll
Mima mound
Mire
Mud flat and muddy coast
Nebkha
Organic weathering
Oyster reef
Peat erosion
Reef
Riparian geomorphology
Saltmarsh
Serpulid reef
Stromatolite (stromatolith)
Termites and termitaria
Tree fall
Turf exfoliation
Vermetid reef and boiler
Zoogeomorphology

Coastal and marine

Atoll
Bar, coastal
Barrier and barrier island
Base level
Beach
Beach cusp
Beach nourishment
Beach ridge
Beach rock
Beach–dune interaction
Beach sediment transport
Blowhole
Blue hole
Boring organism
Bruun rule
Calanque
Cay
Chenier ridge
Cliff, coastal
Coastal classification
Coastal geomorphology
Continental shelf
Coral reef
Corniche
Current
Cuspate foreland
Dune, coastal
Equilibrium shoreline
Estuary
Eustasy
Fjord
Fringing reef
Glacimarine
Groyne
Guyot
Integrated coastal management

Lagoon, coastal
Log spiral beach
Longshore (littoral) drift
Managed retreat
Mangrove swamp
Microatoll
Mud flat and muddy coast
Mudlump
Notch, coastal
Overwashing
Oyster reef
Paralic
Postglacial transgression
Raised beach
Ramp
Rasa and constructed rasa
Reef
Ria
Ridge and runnel topography
Rip current
River delta
Rockpool
Sabkha
Saltmarsh
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Shingle coast
Shore platform
Skerry
Spit
Stack
Steric effect
Storm surge
Strandflat
Submarine landslide geomorphology
Submarine valley
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Tidal creek
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Tsunami
Turbidity current
Vermetid reef and boiler
Visor, plinth and gutter

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Allometry
Astrobleme
Base level
Boundary layer
Bubnoff unit
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Cataclinal
Catastrophism
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Climato-genetic geomorphology
Complex response
Complexity in geomorphology
Computational fluid dynamics
Cycle of erosion
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Dynamic geomorphology
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Ergodic hypothesis
Erodibility
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Erosivity
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Experimental geomorphology
Extraterrestrial geomorphology
Force and resistance concept
Formative event
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Geodiversity
Geoindicator
Geomorphic evolution
Geomorphology
Geomorphometry
Global geomorphology
Grade, concept of
Graded time
Horton's Laws
Hydrological geomorphology
Inheritance
Land system
Landscape sensitivity
Laws, geomorphological
Least action principle
Magnitude–frequency concept
Mathematics
Megageomorphology
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Models
Morphogenetic region
Morphometric properties
Mountain geomorphology
Neocatastrophism
Neotectonics
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Paraglacial
Peneplain
Physical integrity of rivers
Physiography
Planation surface
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Rates of operation
Rejuvenation
Relaxation time
Relief
Relief generation
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Rock mass strength
Roughness
Ruggedness
Sediment budget
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Sediment delivery ratio
Self-organized criticality
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Systems in geomorphology
Threshold, geomorphic
Tropical geomorphology
Uniformitarianism

Fluvial

Abrasion
Accretion
Aggradation
Aligned drainage
Alluvial fan
Alluvium
Anabranching and anastomosing river
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Armoured mud ball
Armouring
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Badland
Bajada
Bank erosion
Bankfull discharge
Bar, river
Base level

Bedform
Bedload
Bedrock channel
Beheaded valley
Blind valley
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Buried valley
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Cavitation
Channel, alluvial
Channelization
Comminution
Confluence, channel and river junction
Contributing area
Cross profile, valley
Cut-and-fill
Dambo
Debris torrent
Dell
Donga
Downstream fining
Drainage basin
Drainage density
Drainage pattern
Dry valley
Dune, fluvial
Estuary
Eversion
Fire
First order stream
Flash flood
Flood
Floodout
Floodplain
Flow regulation systems
Fluvial armour
Fluvial erosion quantification
Fluvial geomorphology
Glacideltaic
Glacifluvial
Gorge and ravine
Gravel-bed river
Ground water
Gully
Headward erosion
Hillslope-channel coupling
Hillslope hollow
Horton's Laws
Hydraulic geometry
Hydrological geomorphology
Hyperconcentrated flow
Initiation of motion
Inland delta
Interfluve
Knickpoint
Large woody debris
Levee
Long profile, river
Maximum flow efficiency
Meandering
Megafan
Mekgacha
Meltwater and meltwater channel
Mining impacts on rivers
Mobile bed
Mound spring
Outburst flood
Overflow channel
Overland flow
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Palaeoflood
Palaeohydrology
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Pediment
Physical integrity of rivers
Piezometric
Pipe and piping
Point bar
Pool and riffle
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Rejuvenation
Reynolds number
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Watershed
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Bergschrund
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Roche moutonnée
Rock glacier
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Factor of safety
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Flow regulation systems
Geoindicator
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Integrated coastal management
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Landslide
Landslide dam
Liquefaction
Managed retreat
Mass movement
Mining impacts on rivers
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Quickclay
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Rocky desertification
Soil conservation
Soil erosion
Stream restoration
Sturzstrom
Subsidence
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Biokarst
Blue hole
Cave
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Corrosion
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Slopewash
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Desiccation cracks and polygons
Diamictite
Drape, silt and mud
Duricrust
Effective stress
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Demoiselle
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(Uranium-Thorium)/Helium analysis
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Mud volcano
Neotectonics
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Orogenesis
Passive margin
Plate tectonics
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Rift valley and rifting
Ring complex or structure
Seafloor spreading
Seismotectonic geomorphology
Shield
Tectonic activity indices
Tectonic geomorphology
Volcanic karst
Volcano
Wilson cycle

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Bowen's reaction series
Calcrete
Caliche (sodium nitrate)
Case hardening
Cavernous weathering

Chelation and cheluviation
Chemical denudation
Chemical weathering
Chronosequence
Clay-with-flint
Corrosion
Deep weathering
Desert varnish
Diagenesis
Dissolution
Duricrust
Eluvium and eluviation
Etching, etchplain and etchplanation
Exfoliation
Ferrallitization
Ferricrete
Fire
Freeze-thaw cycle
Frost and frost weathering
Goldich weathering series
Granular disintegration
Grus
Gypcrete
Gypsum karst
Honeycomb weathering
Hoodoo
Hydration
Hydrolysis
Illuviation
Insolation weathering
Kaolinization
Leaching
Liesegang ring
Lithification
Mechanical weathering
Organic weathering
Oxidation
Pressure release
Reduction
Regolith
Rind, weathering
Rock coating
Salt weathering
Saprolite
Silcrete
Slaking
Solubility
Spalling
Spheroidal weathering
Sulphation
Tafoni
Unloading
Water-layer weathering
Weathering
Weathering and climate change
Weathering front
Weathering-limited and transport-limited
Weathering pit
Wetting and drying weathering

A

ABRASION

The mechanical wearing down, scraping, or grinding away of a rock surface by friction, ensuing from collision between particles during their transport in wind, ice, running water, waves or gravity. The effectiveness of abrasion depends upon the concentration, hardness and kinetic energy of the impacting particles, alongside the resistance of the bedrock surface. Abrasion may scour, polish, scratch or smooth existing rock faces. Abrasion ramps are seaward sloping platforms (typically 1° gradient) formed at the base of cliffs in intertidal environments due to continued wave abrasion.

Further reading

Hamblin, W.K. and Christiansen, E.H. (2001) *Earth's Dynamic Systems*, 9th edition, Upper Saddle River, NJ: Prentice Hall.

STEVE WARD

ACCRETION

The gradual enlargement of an area of land through the natural accumulation of sediment, washed up from a river, lake or sea. Sediment accretion is the basic process of wetland formation, as continuous flooding and subsequent receding river flows emplace sediment which then provides the soil base for wetlands.

Accretion also refers to the theory that continents have increased their surface area during geological history by the addition of marine sediments at their boundaries via tectonic collision with other oceanic or continental plates.

Further reading

Pye, K. (1994) *Sediment Transport and Depositional Processes*, Oxford: Blackwell Scientific.

STEVE WARD

ACTIVE AND CAPABLE FAULT

Currently no universally accepted definition has been agreed upon for 'active fault', nor have the principles and criteria for the identification of active faults and their ranking been worked out. As a result, the various definitions of fault activity terms are the source of some confusion and discussion, both in literature and in practice (see FAULT AND FAULT SCARP).

An important review on this topic was presented by Slemmons and McKinney (1977) who, after examining numerous papers, suggested the following definitions. An 'active fault' is a fault that has slipped during the present seismotectonic regime and is therefore likely to show renewed displacement in the future. Fault activity may be indicated by historical, geological, seismological, geodetic or other geophysical evidence. The definitions for 'capable faults', which were specified for siting nuclear reactors, restrict this term to faults that have been displaced once during the past 35,000 years, or movements of a recurring nature within the past 500,000 years or faults which have been active during the Late Quaternary.

The term active fault in Japan was defined as a fault which has moved repeatedly in recent geological times and could resume activity in the future. Subsequently, this term has been used for faults which have moved during the Quaternary. The analysis of topographic features has provided the most important clues in the work of recognizing active faults (RGAFJ 1980).

A particular definition was given by Panizza and Castaldini (1987) who distinguish two categories: (1) active fault: proven displacement of rocks and/or significant forms; (2) fault held to be active: on the basis of supporting geomorphological or other evidence, but showing no visible displacement of rock or other significant forms. Rocks and/or 'significant' landforms are those included in the neotectonic period considered. The distinction between 'active' and 'held to be active' faults is finalized to constrain in a more precise and less subjective way the concept of fault activity.

The 'World Map of Major Active Faults' shows five fault age categories (historical to <1.6 Ma). Slip rate, which is used as a proxy for fault activity, is classified in four categories ranging from $<0.2\,\mathrm{mm\,year^{-1}}$ to $>5\,\mathrm{mm\,year^{-1}}$. The maps are accompanied by a database which describes evidence for Quaternary faulting, geomorphic expression and paleoseismic parameters (Trifonov and Machette 1993).

In some glossaries, an active fault is defined as 'A fault along which there is recurrent movement, which is usually indicated by small, periodic displacements or seismic activity' (Bates and Jackson 1987), or as 'A fault likely to move at the present day' (Ollier 1988).

A paper on the most commonly used terms associated with seismogenetic faults in the United States was published by Machette (2000). The author notes that the three following terms are used in a variety of ways and for different reasons or applications:

- Active fault: one demonstrating current movement or action (what is meant by 'current'? Contemporary, historical, Holocene or Quaternary?).
- Capable fault: one having the capability for movements.
- A potentially active fault: one capable of being or becoming active (this definition is very similar to that of capable fault).

On the Internet various definitions pinpointing the indeterminateness of the term can be found, such as:

1 The definition of active fault is not straightforward. In some cases, the maximum age that can be determined by means of Carbon 14 analyses (35,000–50,000 years) is used as a time span for such measurements: if a fault can be shown not to have been active within this time span then it is not active (http://www.geol.binghamton.edu/class/geo205/html/faults.html).
2 Active faults are structures along which displacements are expected to occur. By definition, since a shallow earthquake is a process that produces displacement across a fault, all shallow earthquakes occur on active faults (http:// www.eas.slu.edu/People/CJAmmon/HTML/Classes/introQuakes/Notes/faults.html).
3 'A fault that is likely to undergo displacement by another earthquake sometime in the future.' Faults are commonly considered to be active if they have moved one or more times in the last 10,000 years (http://earthquake.usgs.gov/image_glossary).
4 An active fault is one that has moved at least once in geologically recent times. In the Californian definition it means a movement occurring within the last 11,000 years, rather than the longer period of 125,000 years used on New Zealand maps (http://www.gsnz.org.nz/gsprfa.htm).

In short, on the concepts of 'active fault' and 'capable fault' the following remarks can be made:

1 the terms are used to indicate faults which have been subject to movement in recent geological time or which might move at present or in the future;
2 their age limits vary depending on the authors;
3 active faults are often associated with strong earthquakes.

Identification of active and capable faults can be based on direct and/or indirect criteria: historical, geological, geomorphological, geomorphic, seismological, geodetic, geochemical, geophysical and volcanic.

Finally, apart from the terminological aspects, some of the major active faults around the world include: the North Anatolian fault in Turkey, the Dead Sea Valley between Israel and Jordan, the Philippine fault, the San Andreas fault in California, the Red River fault in China and the South Island alpine fault in New Zealand.

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DORIANO CASTALDINI AND
DORINA CAMELIA ILIES

ACTIVE LAYER

Ground above PERMAFROST which thaws in summer and freezes again in winter. In the northern hemisphere, it reaches its full depth each year in late August or September. The active layer is critical to the ecology of permafrost terrain, as it provides a rooting zone for plants and is a seasonal aquifer. An ice-rich zone, commonly below the base of the active layer, is responsible for the sensitivity of permafrost terrain to disturbance. Deepening of the active layer and melting of the ground ice leads to subsidence in flat terrain, and landslides with accelerated erosion on slopes (Mackay 1970).

The active layer thaws once air temperature is above 0 °C and the snow cover has melted. The total depth depends on the length and surface temperature of the thawing season, the ice content of the ground, the thermal conductivity of soil materials, and the temperature of near-surface permafrost. The active layer is thickest in bedrock, where there is little ice to melt. In unconsolidated sediments, the thickness is greatest in dry, sandy soils or gravel, where the depth may be enhanced by heat advected in groundwater, and thinnest in peat. Local variation in soil materials may be reflected in active-layer depth, as in hummocky terrain, where the base of the active layer forms a mirror image of the ground surface, with depth greatest beneath the mineral-soil centres and least beneath the organic-rich circumference of the hummocks.

At the end of the thaw season, freezing of the active layer usually begins from the bottom upwards. Upfreezing commonly accounts for up to 10 per cent of the thickness. During upfreezing, moisture is drawn downward into permafrost from the base of the active layer, leading to development of the ice-rich zone. Simultaneously, soil water is drawn upwards from the rest of the active layer, to freeze near the ground surface. As a result, the centre of the active layer tends to be dry when frozen. Stones and structures embedded in the active layer may be pulled upwards as the ground freezes. Characteristically these objects are supported from below during thawing the following summer, leading to their progressive jacking out of the ground.

Freezing and thawing of the active layer modifies the annual propagation of surface temperature into permafrost. Cooling of permafrost in autumn is delayed by freezing, which may take several months, depending primarily on the water content and snow cover. Mean annual temperature decreases with depth in the active layer, due to the seasonal difference in soil thermal properties wrought by freezing and thawing. The difference in mean annual temperature, or thermal offset, between the ground surface and the top of permafrost may be over 2 °C, increasing with water content and depth of active layer (Romanovsky and Osterkamp 1995).

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C.R. BURN

ACTIVE MARGIN

In plate tectonic theory ocean crust is created by SEAFLOOR SPREADING, and old crust is consumed at subduction sites. A continental margin where subduction occurs is called an active continental margin. Active margins occupy essentially the borders of the Pacific: the west Pacific borders are

ISLAND ARC type; the western margins of the Americas are the other type, to be described here.

The spreading of the Atlantic causes America to move west, where it overrides the seafloor, which is subducted. Plate collision is thought to fold and uplift the continental edge to form mountains and their internal structures, and also create a deep trench offshore where sediments are deposited. These may be scraped off the subducted slab to form an accretionary prism, or subducted where they may produce granites, and andesitic magma which erupts as volcanoes.

The Pacific border of the Americas falls into three main units with different MORPHOTECTONICS: South America, Central America and North America.

The Andes run along the entire western side of South America, divided for most of their length into Eastern and Western Cordilleras, with a graben between called the Inter-Andean Depression. Bedrock is folded and faulted Palaeozoic and Mesozoic rocks, with granite intrusions. The region was largely eroded to a plain before ignimbrites spread over large areas, and planation was complete in the Neogene. The area was uplifted as linear fault blocks in the Plio-Pleistocene, or earlier in some places. The large strato-volcanoes are of Quaternary age, erupted onto the planation surface. Major thrust faults diverge from the centre of the Andes in a symmetrical way, hard to explain by one-sided subduction.

Offshore a deep trench extends as far north as Mexico. Trenches have many graben and normal faults indicating extension. Sediments are usually horizontal, and some trenches are almost empty. Mesozoic plutons constitute the world's greatest granite batholith which runs the length of the Andes, covering 15 per cent of the Andes surface. The alignment parallel to the coast suggests some control on the location and possibly the origin of the Andes, but the plutons took over 70 million years to rise and intrusion ceased about 30 million years ago, long before the uplift of the Andes (Gansser 1973).

The many great volcanoes found along the Andes (with some gaps, and some double lines) are Quaternary. They are on the top of horsts, usually close to the Inter-Andean Depression.

Central America can be regarded as the Middle America arc. The trench has no accretionary prism, and sediments are horizontal.

A basement of metamorphic rocks and granites is exposed in northern Honduras. This is block faulted, and split by the Honduras Depression consisting of north–south graben that opened in the early Pliocene. The chain of volcanoes close to the south coast consists of five straight-line segments. These young cones are built on a basement of older volcanics. The same basement forms the Nicaraguan volcanic upland, separated from the young volcanoes by a major fault scarp. To the north these volcanics overlap the Honduras Massif.

The Isthmian link to Panama is not the young volcanic chain or even rocks of the Nicaraguan volcanic upland, but consists of even older volcanics. Block faulting is common.

Western North America is largely a collage of exotic terranes (Howell 1989). There are abundant strike-slip faults (such as the San Andreas fault) with movement of hundreds of kilometres. There is no offshore trench, but an offshore topography of basins and swells, possibly related to strike-slip fault blocks. These differences perhaps occur because the mid-ocean ridge runs aground near the Mexico/USA border. To the north the transform faults associated with seafloor spreading affect the continental margin as they run nearly parallel to it.

The Pacific border region of the USA consists of two main ranges: in the west are the Coast Ranges, in the east to the north are the Cascades and to the south the Sierra Nevada. The Coast Range seems to have formed as a large but rather simple arch. The Cascade Range is mainly a huge pile of volcanic rocks, with many famous strato-volcanoes such as Mount Shasta and Mount St Helens. The Sierra Nevada is a huge tilt block, mainly uplifted in the Quaternary. The Coast Ranges of Canada are a continuation of the Cascades of the United States, and also consist of a simple arch.

Planation surfaces are common on the North American cordillera (Ollier and Pain 2000). The mountains of North America were uplifted in the Neogene, mostly within the past 5 million years, though subduction has presumably been going on for the life of the Pacific, at least 200 Ma.

Plate tectonic theory has been applied not only to coastal ranges, but to mountains 1,500 km inland (Miller and Gans 1997). The Rocky Mountains consist of elongated blocks aligned in all directions, including east–west (Uinta Mountains).The blocks have Precambrian cores, and divergent thrust faults on both sides. They are too far inland to be explained by subduction, separated from the Pacific by the extensional

Basin and Range Province, and the uplift occurred in the last few million years.

As Gansser (1973) explained, plate tectonic theories that use the Andes as a model adopt simplified assumptions which neglect the fact that only the recent morphogenic uplift made the apparently uniform Andes, masking a very complicated geological history. The same seems true of North and Central America.

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SEE ALSO: mountain geomorphology; plate tectonics

CLIFF OLLIER

ACTUALISM

Actualism is a concept based on the premise that present causes of environmental change are sufficient to explain events of the past. Causes of changes in the past differ not in kind, but often in energy, from those now in operation. The French term *actualisme* and the German terms *aktualismus* or *aktualitatsprinsip* are commonly used in Europe in opposition to catastrophism. Hooykaas (1970) makes a distinction between actualistic methodology and actualistic historical description. Tidal variation over geological time provides an instructive example. Actualist methodology leads to the conclusion that the Moon and Earth were very much closer and that gravitational attraction was therefore greater before 3.5 billion years BP. Huge tidal ranges require a catastrophist historical description.

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SEE ALSO: catastrophism; uniformitarianism

OLAV SLAYMAKER

ADHESION

Adhesion refers to the adhering of wind-blown sand to a wet or damp surface. Adhesion is most common in damp or wet INTERDUNEs between active dunes, but also occurs on SANDSHEETs, beaches, riverbanks and damp portions of dunes. Adhesion ripples and plane bed are the most common surface features that result from adhesion, and each forms a distinctive sedimentary structure with deposition. A related feature is formed by adhesion of sediment to salt during periods of high humidity.

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GARY KOCUREK

AEOLATION

The moulding of desert landscapes by the erosional action of wind (see WIND EROSION OF SOIL). At the start of the twentieth century there was a phase of what has been termed 'extravagant aeolation' (Cooke and Warren 1973). This had its roots in the work undertaken in Africa by French and German geomorphologists, such as Walther and Passarge, but was put forward in its most exuberant form in the USA by Keyes (1912), who believed that material weakened by thermoclasty (INSOLATION WEATHERING) would be evacuated by wind and deposited as dust sheets on desert margins. He argued that the end result of such activity would be the formation of great plains, mountain ranges without foothills, and towering eminences. As he remarked (p. 551):

> Under conditions of aridity plain meets mountain sharply. The bevelled rock-floor of many intermont plains throughout the dry regions is explicable on no known activity of water action in such situations. Existence of isolated plateau

plains rising abruptly out of the general plains surface far from any sight of running water is an anomaly met with only in the desert.

Not all American geomorphologists took such a firm view as Keyes, and Tolman (1909), for instance, in his study of the Arizona BOLSON region, considered the role of both fluvial and aeolian processes to be important and recognized that STONE PAVEMENTS 'fortified' large tracts of the arid region of the south-west of the USA against wind attack.

Aeolianist views declined in popularity so that from about 1920 onwards the belief that entire landscapes were shaped by wind became less acceptable. The reasons for the decline of aeolianist views were many.

First, the great PEDIMENT landscapes of the American deserts were seen, following the work of McGee (1897) and others, as being attributable to planation by sheetflood activity. The second reason for the decline of aeolianist views was that many desert landscapes were thought to have been moulded by fluvial processes that had been more powerful and widespread during the pluvial phases that were held to be a feature of the Pleistocene. Third, doubt was expressed about the power of thermoclasty as a process capable of preparing desert surfaces for subsequent aeolian attack. Such doubt largely arose because of laboratory simulations. Fourth, it was widely held that lag gravels (stone pavements) and salt and clay crusts would limit the extent to which aeolian processes could cause excavation of surfaces below the water table. Fifth, it became apparent that many of the world's great LOESS deposits, in North America, China and the erstwhile USSR, were the product of deflation from glacial areas rather than from deserts. Glacial grinding was thought to be the most efficient way of producing silt-sized quartz particles. Sixth, it was recognized that not all deserts had either adequate supplies of abrasive sand or of frequent high-velocity winds for wind erosion to be achieved with any degree of facility. Finally, features that were conceded to have an aeolian origin (e.g. YARDANGS, VENTIFACTS and pedestal rocks) were thought to be but minor, bizarre embellishments of otherwise fluvial environments, whilst other possibly aeolian features (notably stone pavements and closed depressions) were also explicable by other means. STONE PAVEMENTS, for example, could be the product of the removal of fine sediments by sheetflood activity or they could result from vertical sorting processes associated with wetting and drying, dust inputs, salt hydration or freezing and thawing. Deflational removal of fines to leave a lag was just one possible formation mechanism. In the same way, closed depressions could be attributed to wind excavation, but might also be explained by tectonic, solutional or zoogenic processes.

Nevertheless, the power of wind erosion cannot be dismissed. Closed depressions (PANS) and wind moulded landforms (yardangs) are important landforms in some arid areas and wind erosion plays a significant role in their development.

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A.S. GOUDIE

AEOLIAN GEOMORPHOLOGY

Aeolian geomorphology is the study of the effect of the wind on Earth surface processes and landforms. It encompasses studies of the fundamental physical mechanisms and movement of materials at the scale of a single grain, studies of the development of landforms such as dunes (see DUNE, AEOLIAN) and YARDANGS, and studies of the wider effect of the wind at the regional scale of SAND SEA AND DUNEFIELD, SANDSHEETS and LOESS deposition. It is also concerned with applied aspects of aeolian activity. In recent years this has led to a particular focus on the erosion by wind of soils in agricultural lands, especially in the semi-arid lands (see WIND EROSION OF SOIL and DEFLATION). Aeolian geomorphology also includes the study of the palaeoenvironmental significance of aeolian features, for landscapes can be just as sensitive to changes in the activity of the wind as they are to water and ice. As with other elements of geomorphology, technological changes in recent years have led to considerable advances in the understanding of aeolian geomorphology.

At the scale of the movement of individual sand and dust particles, the benchmark work was undertaken by Bagnold and summarized in his

Physics of Blown Sand and Desert Dunes (1941). Since Bagnold's work very considerable progress has been made in the use of wind tunnels and field instruments (see WIND TUNNELS IN GEOMORPHOLOGY). Although the fundamental physics of aeolian sand transport is much as Bagnold described it, very considerable detail has been added in recent years (see also AEOLIAN PROCESSES and SALTATION).

At the scale of individual landforms considerable progress has also been made. As a result of the improvement of technologies for data capture there has been a spate of studies of wind flow and sand flux on single dunes (e.g. Tsoar 1983; Walker 1999) reviewed by Wiggs (2001). Often these are coupled with improving surveying techniques that enable accurate measurement of change (e.g. Stokes *et al.* 1999). In addition, ground penetrating radar (GPR) is now being routinely used to ascertain the internal sedimentary structure of dunes as an important indication of the evolutionary history of dunes (e.g. Bristow *et al.* 2000). Studies have also investigated erosional features such as YARDANGS and VENTIFACTS (e.g. Laity 1994).

At the regional scale the development of remote sensing has enabled a better grasp of the relationships between landforms. The advance of remote sensing investigations in dryland areas has been of particular importance because desert areas are often difficult to access. The pioneering work of McKee and co-workers (McKee 1979) has been followed by numerous applications of remote sensing in aeolian studies. Imagery has been used to map dune patterns (e.g. Al-Dabi *et al.* 1997), detect small changes in dune morphology using high resolution synthetic aperture radar (SAR) imagery (e.g. Blumberg 1998), map and detect dust emissions from dryland pan systems (e.g. Eckardt *et al.* 2001) and detect mineral assemblages (e.g. White *et al.* 1997).

Aeolian features also hold considerable palaeoenvironmental information because aeolian activity is sensitive to changes in environmental controls such as wind energy and moisture availability. The extent of dunefields at the last glacial maximum was used as a surrogate indicator of global aridity by Sarnthein (1978) but a basic on/off classification of aeolian activity is now seen as too simplistic (Livingstone and Thomas 1993). Kocurek and Lancaster (1999), for instance, have sought to incorporate variability of sediment availability along with wind energy in discussions of past aeolian activity in the Mojave Desert.

A profound impact on aeolian studies has been the development of luminescence dating techniques (see DATING METHODS). Many aeolian deposits lack organic matter and so have not been susceptible to radiocarbon dating. Since the early 1980s luminescence dating primarily of quartz grains has enabled dating of aeolian deposits such as dunes and loess, and luminescence dates in aeolian studies are now commonplace (e.g. Stokes *et al.* 1997).

Improvement of dating techniques has led to considerable interest in the palaeoenvironmental information stored in LOESS (terrestrial deposits of aeolian dust). The best documented of these are the deposits of the Chinese loess plateau. Here mineral magnetism has been used as a proxy for weathering of PALAEOSOLS, patterns of magnetic reversals have been used to date deposits covering the past 2.5 million years and loess particle size has been used as an indicator of palaeo wind speeds. Techniques developed on the Chinese deposits have been extended to loess deposits elsewhere and knowledge of the extent and nature of world loess deposits has steadily increased (e.g. Derbyshire 2001).

Aeolian geomorphology has moved on considerably since the claims of Keyes in the early part of the twentieth century (see AEOLATION). The task that faces aeolian geomorphologists is to move from studies of individual landforms formed predominantly by aeolian activity to consider the wider role of the wind alongside water and ice in forming landscapes (e.g. Bullard and Livingstone 2002).

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IAN LIVINGSTONE AND GILES F.S. WIGGS

AEOLIAN PROCESSES

Wind is the movement of the mixture of gases that constitute the air. It is a fluid like water and obeys the same fundamental physical mechanisms as water. However, there are clear distinctions between the effects of water and wind at the Earth's surface. Air is 100 times less dense than water and consequently is only able to carry small clastic material. However, wind is not constrained by channels in the way that much water action is and consequently its influence can be much wider spread. Paradoxically, it is this wide spread of activity that means that aeolian activity sometimes goes unrecognized. A few millimetres of erosion or deposition of material over a large area is much less obvious than the erosion of rills and gullies or deposition of bars in channels even though the total amount of material moved may be similar.

Controls on aeolian processes

While aeolian activity is often associated with hot deserts, it is not restricted to these areas, although these are among the regions with the most favourable conditions for aeolian activity. Primary requirements for aeolian activity are: sufficient wind energy; material of a size that can be transported; and surface conditions that make that material available to the wind. Aeolian activity is therefore controlled by transport capacity, sediment supply and sediment availability (Kocurek and Lancaster 1999).

- *Transport capacity* Most places on the Earth's surface experience sufficient wind energy for aeolian processes to operate, so wind energy is rarely a limiting factor. The high levels of aeolian activity in low-latitude deserts do not occur because they are windier than other places. In fact the windiest places on Earth are close to the poles and around coastlines.
- *Sediment supply* Because wind is not as dense or viscous as water it is much more selective about the size of material that it can carry. The size of material most readily entrained by the wind is fine sand (see below). Wind rarely carries material above sand size, although transport of gravel-sized particles (>2 mm) has been reported from the dry valleys in Antarctica where wind speeds are very high and the extremely cold air is dense. The size-selectivity of wind means that surface materials must usually be sand- or dust-sized to be entrained. Often this requires that the materials are pre-sorted by other fluvial, glacial or marine processes. Some of the best sources of aeolian material are alluvial fans, glacial outwash plains and beaches (Bullard and Livingstone 2002).

- *Sediment availability* Provided with sufficient wind energy and material of the right size, the remaining control on aeolian processes is the surface conditions. Deserts have high levels of aeolian activity because soil, vegetation or moisture do not seal the surface. Conversely, aeolian activity is more rare in mid-latitudes, not because of lack of wind or material of the right size, but because surface conditions prevent the wind from entraining material.

Processes of wind erosion

Erosion of materials at the Earth's surface by the wind occurs as a result of two processes: deflation and abrasion (both of these mechanisms also occur in flowing water although the equivalent term 'fluid stressing' is used instead of deflation in fluvial geomorphology). DEFLATION is very simply the entrainment of material by the wind. Surfaces on which dust- or sand-sized material are exposed are particularly susceptible and in some places agricultural land with sandy or dusty soils where farmers expose the soil by ploughing is subject to considerable deflation by the wind (see WIND EROSION OF SOIL). Abrasion is caused by bombardment by particles being transported by the wind, most usually by SALTATION (see below). The impact of these transported grains can cause considerable sculpting of natural and built features. YARDANGs and VENTIFACTs are the geomorphological features most affected by abrasion.

Sediment entrainment

Aeolian sediment entrainment on a stable non-eroding surface occurs when the shear stress of the wind (a function of wind speed, turbulent energy and surface roughness) overcomes forces of particle cohesion, packing and weight. The principal erosive forces include lift, form drag and surface drag. The first two of these forces both result from air pressure differences around an individual particle. Higher velocity winds are associated with lower air pressure, so where wind flow is accelerated over a particle lying on the surface there is also a decrease in pressure above the particle resulting in a lift force. Similarly, form drag results from the high wind pressure on the upwind side of the particle contrasting with the decreased pressure in the downwind region. These two pressure forces combine with the surface drag resulting directly from the shearing stress of the wind to shake the particle loose before spinning it up into the airstream.

The relationship between erosivity and entrainment can effectively be simplified to two parameters, critical wind shear (u_{*ct}) and particle diameter (d) (Bagnold 1941):

$$u_{*ct} = A\sqrt{\frac{(\sigma - \rho)}{\rho}\, g \cdot d}$$

where: σ = particle density, g = acceleration due to gravity, A = constant dependent upon the grain Reynolds number (≈ 0.1).

Generally, larger grains require a greater wind shear to dislodge them. However, as shown in Figure 1, this relationship is reversed for particles smaller than about 0.06 mm (dust-sized) where increased electrostatic and molecular cohesion require larger erosive forces for entrainment. Figure 1 also demonstrates that the grain sizes most susceptible to entrainment have diameters between 0.06 and 0.40 mm, sand-sized particles. It is this susceptibility of sand to entrainment that allows the accumulation of extensive dunefields (see SAND SEA AND DUNEFIELD) and SANDSHEETs in dryland regions.

A further process important in the entrainment of sand grains is the bombardment of the surface by grains that are already in transport. Once a few grains have been entrained by the wind they may be transported by the process of saltation,

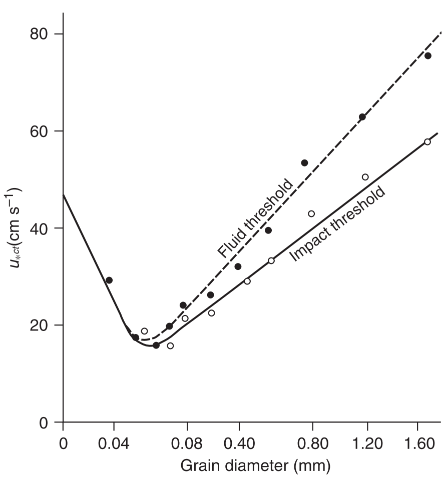

Figure 1 The relationship between particle size and the threshold shear velocity required for entrainment (after Chepil 1945)

Source: Thomas, D.S.G. (1997) *Arid Zone Geomorphology*, 2nd edition. © John Wiley & Sons Limited. Reproduced with premission bouncing along the surface with ballistic trajectories (see below). Each saltating grain gathers momentum from the wind and then imparts it to the sand surface on impact. This impact can 'splash-out' up to ten other grains that may also become entrained by the wind. A few saltating grains can quickly induce mass transport of sediment in a cascading system (Nickling 1988). Two thresholds of entrainment may therefore be identified: one (the fluid threshold) relates only to the drag and lift forces of the wind, the second (the impact threshold) is lower and combines wind forces with additional forces provided by impacting grains already in transport (see Figure 1). Once a sediment surface has begun to be eroded by wind forces at the fluid threshold, sediment transport is maintained at the lower impact threshold because energy is also available from the saltating grains. Wind shear stress may therefore reduce once entrainment has begun, but sediment transport will continue until the wind drops below the new impact threshold.

Transport mechanisms

The grain size of entrained particles also determines the mode of transport undertaken. Although dust-sized particles are not the easiest to entrain, they have very low settling velocities in comparison to potential wind lift and turbulent velocities, so can be transported in *suspension*. Particles suspended in the atmosphere may be held aloft for several days and hence travel long distances. An example of this is the deposition of Saharan dust in the south-eastern USA (Prospero 1999). Often this transport in suspension is in barely visible dust haze, but sometimes there are more concentrated dust plumes which are clearly seen on spectacular satellite images, and still less frequently suspended aeolian material is concentrated as dust storms which can lead to 'blackout' conditions.

Particles up to about 1.0 mm are commonly transported in SALTATION. Figure 2 shows the typical trajectory of saltating particles with a progressively increasing forward velocity from entrainment to impact as the particle draws momentum from the wind.

The actual trajectory of a particle depends on the height of its bounce. Wind velocity increases at a logarithmic rate away from the surface and so a particle that bounces higher into the wind will be able to draw greater momentum from it and so travel further and faster. The length of jump is thought to be about 12–15 times the height of bounce, or further if the particle spins

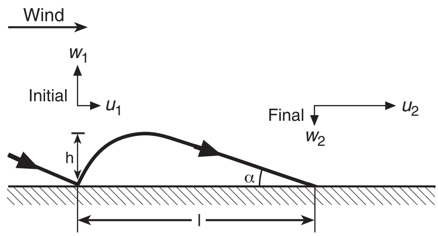

Figure 2 The ballistic trajectory of a saltating sand grain. *w* and *u* represent vertical and horizontal velocities, respectively (after Bagnold 1941)

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and induces an additional lift force (called the Magnus effect, White and Schultz 1977). The height of the saltation layer is dependent both on the wind velocity and also on the hardness of the surface over which the particles are saltating. Sand that is saltating over a rock or pebbly surface loses much less of its momentum on impact and so tends to bounce higher, reaching up to 3.0 m. An average saltation height, however, is about 0.2 m. The amount of sand in transport declines exponentially with height and so up to 80 per cent of all saltation activity takes place within 0.02 m of the surface (Butterfield 1991).

Grains which are ejected into the airflow as a result of the impact of a saltating grain may also enter the saltation system. However, some of these ejected grains may not have sufficient velocity fully to enter saltation and hence take only a single jump in a downwind direction. This process is termed *reptation* and, whilst much further research is required into its operation, it may be very significant in near-surface aeolian transport (Anderson *et al.* 1991).

The final mode of sand transport is *creep* and this describes the downwind rolling of larger sand particles (usually >0.5 mm). Such a process results both from the drag of wind on the surface of the particles and also the high velocity impact of saltating grains. It is thought that the process of creep may account for up to one-quarter of the bedload (saltation plus creep) transport rate.

Sand flux

The mass flux of sand (q) transported during an erosion event is often calculated as a cubic function of wind shear velocity (u_*). Most relationships are derived from theoretical analyses or wind tunnel experiments and a popular expression is that of Lettau and Lettau (1978):

$$q = C\left(\frac{d}{D}\right)^{0.5}(u_* - u_{*ct})u_*\frac{2\rho}{g}$$

where: C = constant (4.2), d = grain diameter, D = standard grain diameter (0.25 mm), u_{*ct} = shear velocity threshold of grain entrainment, ρ = air density, g = acceleration due to gravity.

There has been little empirical testing of relationships like the one above and that which has been accomplished shows considerable variation between observed and predicted rates. Such variation is to be expected when the complex nature of the saltation system is considered and the fact that the predictive expressions available rarely account for variations in terrain, vegetation, surface moisture or wind turbulence. Furthermore, the accurate measurement of sand flux in the natural environment is very difficult, with the published efficiencies of sand traps varying between 20 and 70 per cent (Jones and Willetts 1979).

Deposition

Just as material is entrained when shear stress overcomes inhibiting forces, so material is deposited when shear stress is no longer greater than these forces. This manifests itself both at the large scale where, if regional wind patterns lead to a decrease of wind speed, SANDSHEETs or dunefields (see SAND SEA AND DUNEFIELD) are formed, but also at the much smaller scale where surface irregularities may be responsible for the deposition of sand patches. Finer-grained material carried in suspension is often deposited as LOESS.

Although a lack of vegetation is usually important in the aeolian entrainment of material, paradoxically its presence can be important in trapping dust and sand in depositional features. Dust is only deposited as loess where it is prevented from re-entrainment, often by vegetation, and vegetation can also be important in stabilizing coastal dune ridges.

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GILES F.S. WIGGS AND IAN LIVINGSTONE

AEOLIANITE

Aeolianite is a cemented sandstone that has formed as a result of the processes of entrainment, transportation and deposition by wind. This rock type has various names including eolianite (US), miliolite (India and the Middle East), dunerock (South Africa), kurkar (Israel) and grès dunnaire (Mediterranean). Aeolianites of Quaternary age are most commonly found between 20° and 40° either side of the equator although examples have been found as far north as about 60°. Most aeolianites in the geological record are Quaternary in age and these tend to range between about 0.5 m to about 100 m in thickness.

Most aeolianites are rich in carbonate although silica-dominated forms also exist. Carbonate-rich aeolianites are largely associated with coastal sources of sediment and are thus close to modern or palaeo-shorelines. Semi-arid to sub-humid tropical shorelines are the most suitable locations for aeolianites as the oceans are productive in the formation of shelly biogenic grains or ooliths; strong onshore currents breakdown and move the sediment onshore; and climatic conditions are conducive for subsequent DIAGENESIS. In arid environments where there are strong onshore winds and a high sediment supply, dunes may be transported several hundred kilometres inland.

Along coastlines aeolianites often form elongate shore parallel or oblique bodies deposited as transverse ridges. The dunes are often stacked up against one another and may coalesce. The sediment size of aeolianites is typically coarse silt to sand-sized. Clays are rare because those that are entrained by the wind tend to be removed by suspension. Insufficient wind energy to transport grain sizes coarser than 2 mm generally limits the upper size of the clasts.

Aeolianites have distinctive bedding structures such as cross-bedding and laminations, which represent the progradation and growth of the dunes. Steeply dipping units up to 30°–34° reflect the former dune slipface. As a result of erosion the palaeodune bedforms are commonly lost and the dune type and direction of sand movement has to be deduced largely from the internal structures.

Lithification under freshwater vadose, mixed and/or phreatic conditions may occur. The main diagenetic processes result in alteration of unstable aragonite and high-Mg calcite clasts to low-Mg calcite clasts and cement. The balance between dissolution of carbonate grains by leaching and the production of cement is the prime control on the degree of dune induration. The main sources for the cement come from biogenic skeletal remains (e.g. molluscs, foraminifera, echinoderms, algae and coral fragments), ooliths, biota, sea spray, dust, bedrock and ground water.

Alteration of aeolianites by diagenetic processes in the vadose environment is the most common, occurring in three ways: (1) by loss of Mg^{2+} from the crystal lattices of high-Mg calcite; (2) by dissolution of aragonite, the loss of some strontium and reprecipitation as low-Mg calcite; and (3) by calcitization of aragonite *in situ*. A wide range of controls results in significant variability in aeolianite diagenesis in terms of both causal factors and diagenetic product (Gardner and McLaren 1994). Such controls include climate, sea level and time at the macro-scale; sea spray, plants and texture at the meso-scale and at the micro-scale the amount, rate of movement and chemistry of pore waters.

Major unconformities in aeolianites are often marked by PALAEOSOLs that develop as a result of solution and weathering. Commonly these soils are *terra rossa* and red latosols that have developed *in situ* but may contain inputs from wind-blown dust. Weathering may subsequently result in a solution of carbonate products and karstification. In semi-arid environments surface crusts and thin laminar CALCRETEs often form as a result of solution and rapid reprecipitation.

Radiometric dating of aeolianites is notoriously difficult. The effects of diagenesis mean that there are chances of contamination from secondary calcite. Occasionally unaltered shells are found which have allowed radiocarbon dating (e.g. McLaren and Gardner 2000). In addition, uranium series dating, amino acid racemization, luminescence and electron spin resonance dating have been used with varying success (see Brooke 2001).

Lithification increases the aeolianites' resistance to erosion and enhances their preservation potential in the geological record. The cement types (such as meniscus, rim, pore filling and needle fibre), amount and distribution, along with geochemistry, can aid interpretations concerning palaeoenvironments such as identifying palaeo-water tables, palaeo-erosion surfaces or degree of exposure to marine environments.

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SEE ALSO: karst

SUE McLAREN

AGGRADATION

The long-term accumulation of sediment in a channel and the readjustment of the stream profile where there is a vertical growth of the land surface in response. Some possible agents of this process are running water, waves, glaciers and wind. Aggradation can occur at a variety of spatial scales and temporal scales (gradual or PUNCTUATED AGGRADATION), and may take place under constrained or unconstrained conditions. As aggradation is a long-term process, short-term fluctuations in sediment transport have no relevance.

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STEVE WARD

ALAS

The periglacial landform term alas, which is of Yakutian origin, was first introduced by the Russian worker P.A. Soloviev in the 1960s. It refers to the substantial circular and oval depressions with steep sides and flat floors, sometimes occupied with lakes, which characterize the geomorphology of the higher river terraces in central Yakutia (65°N, 125°E). Alases typically have diameters from 0.1 km up to 15 km and depths in the 3–40 m range. In morphological expression a tract of alas depressions on a river terrace surface is not dissimilar to a suite of KETTLE AND KETTLE HOLE forming a pitted glacial outwash plain. Genetically, an alas is a type of THERMOKARST feature, i.e. a subsidence landform arising from the degradation and settlement of ice-rich frozen ground. An essential prerequisite for alas development is terrain with a high ground ice content. The natural vegetation cover is taiga (coniferous forest) although the alas floors are often grass covered.

Coalescence processes amongst individual alas depressions can lead to the development of alas valleys. These valleys are characterized by a variable width with an alternation of narrow sections marking the former location of watersheds and wide sections together with branches with no outlets. The longitudinal profiles are not necessarily graded, reflecting the fact that they are thermokarst landforms rather than normal river cut features. In the Yakutian lowlands drained by the Lena River the spread of alas-related depressions has affected some 40 per cent of the higher river terrace surfaces. Surprisingly, the alas valleys form pockets of cultivated land where hardy strains of some grains along with some root crops are produced during the relatively warm summers.

Some prominence has been given in periglacial texts to a hypothetical reconstructed sequence of alas development (Soloviev 1973) although it needs to be emphasized that its applicability outside Yakutia has yet to be established. An important factor is that Yakutia is unglaciated yet nevertheless sustained permafrost throughout much of the Quaternary. Accordingly its ground ice history is complex with, for example, massive syngenetic ice wedges attaining sizes well in excess of those formed epigenetically.

The initial stage in alas development is a disturbance to the ground surface thermal regime's equilibrium state, such as can arise from the destruction of the natural vegetation as the result of climatic change, a forest fire or human activities. The upset thermal balance invariably leads to the degradation of the ice wedge tops beneath their surface polygonal troughs and the resultant growth of an enhanced hummocky surface morphology. Once ponded water accumulates between the mounds, further ice wedge decay is inevitable as in summer the water quickly warms and heat is transferred to the ice beneath. Thaw settlement in conjunction with progressive amalgamation of the ponds accelerates the melting process and leads to the creation of a thermokarst lake at the bottom of a major flat-bottomed depression. With time stability may be attained and lakes occupying old alas depressions may disappear through either sedimentation or drainage. Either process causes the sub-lake taliks to shrink leading to the growth of one or more PINGOs beneath the now dry hollow floor.

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SEE ALSO: ice wedge and related structures; permafrost; pingo; thermokarst

PETER WORSLEY

ALIGNED DRAINAGE

A parallelism of drainage lines. In some cases parallel or aligned drainage (rather than more normal dendritic patterns) covers great areas. As W.L. Russell (1929: 249) wrote:

> One of the most remarkable features of the northwestern Great Plains is the well-defined northwest–southeast alignment of the valleys

and ridges. The prevailing direction of the valleys and ridges is nearly identical over such great areas that it is evident that the causes or forces which produced the parallelism must have operated on a grand scale.

Using available maps, Russell showed that the alignment was developed in parts of western South Dakota, western Nebraska, western North Dakota, western Montana and eastern Wyoming. He suggested that this alignment was not caused by any structural control (though in other areas this is a perfectly valid hypothesis) but was associated with the former presence of sand dunes and associated interdunal channelling of erosion, particularly in the susceptible Pierre Shale. The aeolian hypothesis was endorsed by Flint (1955) who pointed out that alignment could be produced either by the former existence of linear dunes or by deflation of susceptible materials such as the Pierre Shale. Aligned drainage also occurs on the High Plains of Texas.

Similarly large expanses of aligned drainage occur in parts of Africa and are associated with the former greater extents of the Kalahari and Sahara deserts. In southern Angola, for example, even in areas where the current mean annual precipitation is as high as 1,200 mm, aligned stream channels (many of which are tens of km in length) run from east to west, as do the old dunes of the Mega-Kalahari (Thomas and Shaw 1991). In west Africa aligned drainage, related to the Pleistocene expansion of the Sahara, occurs as far south as southern Nigeria and Cameroon (Nichol 1998).

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A.S. GOUDIE

ALLOMETRY

Allometry is the measurement of proportional changes in parts of an organism and correlated with variation in size of the total organism (Gould 1966). Church and Mark (1980) have provided the most comprehensive discussion of the geomorphological applications of allometry.

Allometric relations are usually described by power laws, such as $y = ax^b$ where x is an index of system scale, y is an attribute of the system, a is a constant and the exponent b is the ratio of x and y.

Four distinctions need to be made:

1 *Allometry and isometry* If x and y have the same dimensions, isometry obtains when $b = 1$. Under this condition, there is no change in the relative proportions of x and y with increasing scale and the system is described as self-similar. When $b \neq 1$, the relation is allometric, implying a scale-related distortion of geometry. For example, in Bull's (1964) analyses of the areas of alluvial fans compared with their contributing drainage areas, $b = 0.9$ and allometry obtains. This is an indication that larger drainage basins have a relatively greater tendency to store sediment than smaller basins.
2 *Negative and positive allometry* If $b > 1$, the relation is positively allometric; if $b < 1$, the relation is negatively allometric. However, care must be taken to check that the dimensions of x and y are the same. If, for example, y is a length (L) and x is an area (L^2), then a value of b of 0.5 would indicate isometry and values of $b >$ or < 0.5 would indicate positive and negative allometry respectively.
3 *Dynamic and static allometry* In biology it is relatively easy to compare organisms at various stages of growth (dynamic allometry). Typically, landforms are compared at one moment in time with little control over their absolute ages (static allometry). There are serious limitations to static allometry, not least of which is the spatial heterogeneity of geological materials and the difficulty of defining drainage basins with similar growth histories.
4 *Simple and compound allometry* If the b value of an allometric relation changes as system scale changes compound allometry obtains. There is increasing evidence that compound allometry results from dominant process change between slope-dominated and channel-dominated basins.

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SEE ALSO: fractals

OLAV SLAYMAKER

ALLUVIAL FAN

Alluvial fans are depositional landforms created where steep high-power channels enter a zone of reduced STREAM POWER. Typically they range in scale from axial lengths of tens of metres to tens of kilometres. They are usually cone-shaped forms with surface slopes radiating away from an apex, located at the point where the feeder channel enters the fan. This form can be modified by the presence of confining neighbouring fans or valley walls. In addition, the burial of the fan apex area can cause backfilling into the mountain catchment. Alluvial fans are subaerial features, however if they extend into water they are known as fan deltas.

Many of the classic studies of alluvial fans, which established the basic properties, were carried out in the basin-and-range terrain of the deserts of the American south-west (Blackwelder 1928; Blissenbach 1954; Hooke 1967), culminating in Bull's (1977) review paper. Since then there have been many studies of other (mostly) dry-region fans, with emphasis on relations between sedimentary processes and morphology (Wells and Harvey 1987; Blair and McPherson 1994) and on fan dynamics (Harvey 1997), in addition to studies of fans in humid regions (see Rachocki and Church 1990).

Fan occurrence

Alluvial fans occur in two characteristic situations: at mountain fronts and at tributary junctions. In both cases, high sediment loads encounter zones of reduced stream power, with accommodation space for deposition. These conditions are controlled by long-term landform evolution, including the tectonic setting and erosional history. Mountain fronts may be fault-controlled or erosional, in which case the fans may bury an older PEDIMENT surface. Tributary-junction settings are controlled by the long-term dissectional history. A common fan setting occurs in glaciated mountain terrain, where steep tributary valleys join wide formerly glaciated valleys.

Much of the literature emphasizes the importance of alluvial fans in desert mountain areas. In such areas, FLASH FLOODs transport abundant coarse sediment, and the depositional setting created by regional tectonics may be enhanced by the tendency for desert floods to lose power downstream. However, neither active tectonics nor aridity are prerequisites for fan formation. Fans can occur in mountain areas in all climatic settings and in tectonically stable areas, provided there is juxtaposition of high coarse-sediment transport and a sudden downstream loss in transporting power. But, as outlined below, as fan morphology tends to respond to climatically controlled water and sediment supply, climatic change can induce a change in fan processes and morphology.

Fan processes

Processes on fans include four groups. Primary processes deliver sediment to the fan, principally by DEBRIS FLOWs, or fluvial processes (by channelized and/or sheet flows). These processes are expressed by the sediments comprising the fan and by the surface morphology. Debris flows are massive, usually matrix-supported coarse sediments with clasts up to boulder size. Depositional features may include lobate and levee forms. Fluvial sediments in fan environments are usually moderately sorted gravels and cobbles, stratified or lensed in channelized or sheet bodies. Depositional features may include a range of channel forms or shallow bar and swale topography.

Fans have been classified on the basis of the primary processes into debris-flow and fluvially dominant fans. These processes are catchment controlled and depend on the water:sediment mix fed to the fan during flood events. Debris flows operate as sediment-rich flows, but under conditions of greater dilution become transitional or HYPERCONCENTRATED FLOWS then fluvial flows. Debris flows are most common where sediment concentrations are high, e.g. from small, steep catchments (Kostaschuk *et al.* 1986). Fluvial processes are more common from large, less steep catchments. The old-fashioned distinction between 'dry' and 'wet' fans, interpreted on the basis of climate, is outmoded – the primary processes are controlled mainly by catchment characteristics.

Secondary processes rework the sediment on the fan by fluvial, or in arid areas by AEOLIAN

PROCESSES. Third, stabilization processes involve surface modification by soil formation and vegetation colonization. Such processes may influence the hydrology of the fan surface, but are important in fan studies as they allow the relative ages of fan segments to be assessed (see McFadden *et al.* 1989). In arid and semi-arid areas such processes include surface modification by desert

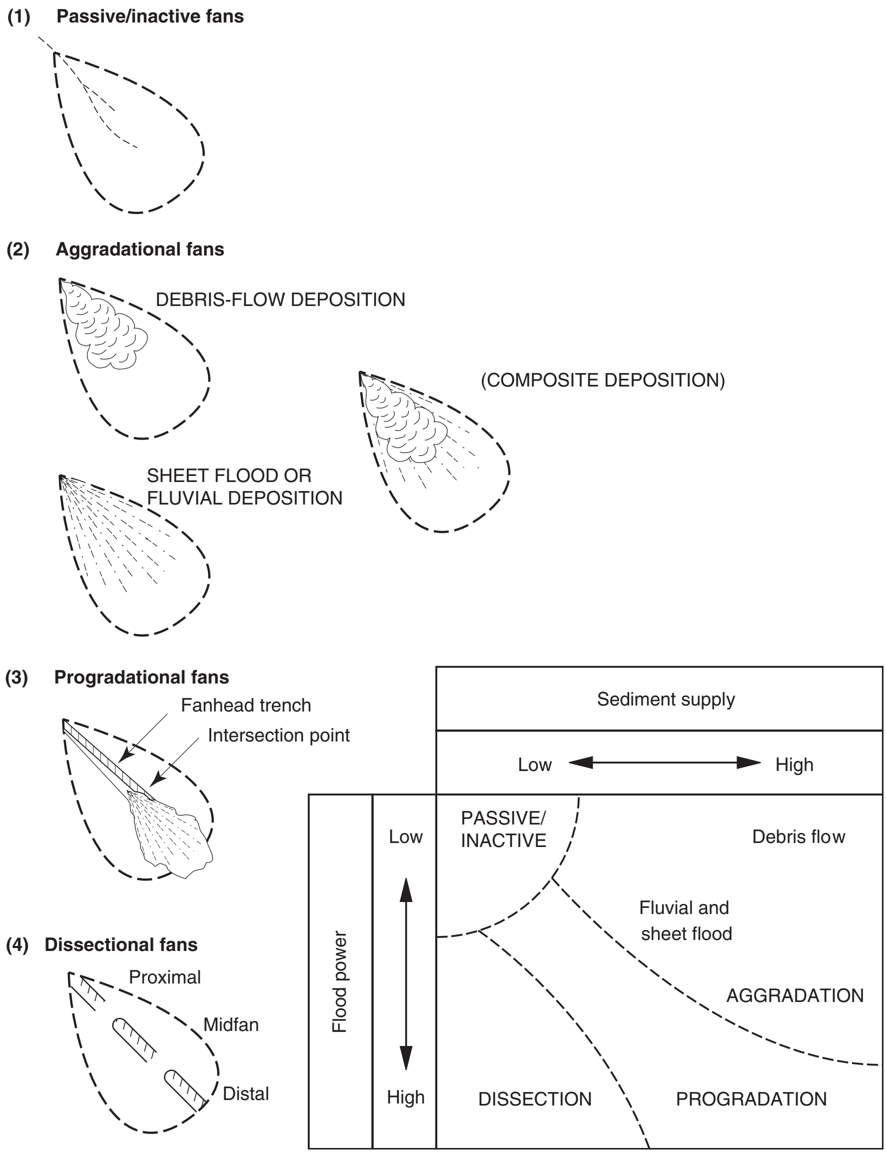

Figure 3 Alluvial fan styles: response to flood power and sediment supply (modified from Harvey 2002c)

pavement (see STONE PAVEMENT) formation and DESERT VARNISH development, and pedogenic processes leading for example to carbonate accumulation and CALCRETE formation. In humid areas lichen colonization and colonization by higher plants may be important as well as soil formation. Finally, dissection processes may erode the fan surface. Dissection may simply increase with fan surface age, or be accelerated by climatic or base-level change.

Fan morphology

Within the context of the topographic setting, fan morphology reflects fan processes and evolution. The relationships between erosion and deposition on the fan can be described as fan style (Figure 3), which in turn depends on the relationship between flood power and sediment supply. Under conditions of low power and little sediment supply the fan may be inactive. Under conditions of excess sediment supply the fan will aggrade, by debris-flow or fluvial processes dependent on the water:sediment mix fed to the fan. Such AGGRADATION will occur from the fan apex downfan. Commonly, both power and sediment supply are moderate. The feeder channel incises into the fan surface to form a fanhead trench, which emerges onto the fan surface at a midfan intersection point (Plate 1), beyond which deposition occurs. Such fans are described as 'telescopic' and may extend by progradation. A zone of coalescent deposition from adjacent prograding fans is known as a BAJADA. If power

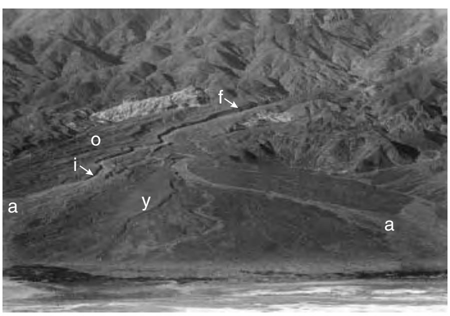

Plate 1 Characteristic alluvial fan morphology: Death Valley, California (photo: A.M. Harvey)

Notes: f = fanhead trench; i = intersection point; o = older fan surfaces; y = younger fan surfaces; a = active depositional segment

is excessive, either through high runoff or sediment starvation, erosion may dominate. Erosion may be concentrated within the fanhead area, in midfan, or in the case of base-level induced erosion, at the fan toe.

On many fans, BASE LEVELS are stable, at least over moderate timescales, and fan processes are primarily proximally controlled – by water and sediment supply from the catchment. A climatic (or other environmental change, e.g. related to human activity) causing changes in water and sediment supply may result in a change in fan style towards greater erosion or deposition.

Two aspects of fan morphometry have been demonstrated to reflect fan context, processes and evolution. General relationships of fan area and fan gradient to drainage areas have the forms:

$$A_f = p\ A_d{}^q \qquad (1)$$
$$G_f = a\ A_d{}^{-b} \qquad (2)$$

(where A is area, G is gradient, f of the fan, d of the drainage basin, pqab are constants). For the fan-area relationship, exponent q generally ranges between 0.7 and 1.1, and the value of the constant p reflects fan age, degree of confinement, basin area, geology and climate. For the fan-gradient relationship exponent b generally ranges between −0.15 and −0.35, and the value of constant a primarily reflects sedimentary processes (Harvey 1997). Debris-flow fans are steeper than fluvial fans. Fan-surface and fan-channel profile relationships (Figure 4) reflect erosion and deposition histories, and the interaction between proximal climate- and sediment-led controls and distal base-level controls.

Fan dynamics

Three sets of factors affect the geomorphology of alluvial fans: (1) context and locational factors, particularly tectonics and geomorphic history; (2) water and sediment delivery to the fan, controlled in the context of catchment geology size and relief, largely by climatic factors; (3) factors affecting the fan environment itself, especially base level.

Interactions between tectonic, climatic and base-level factors form a major thrust of alluvial fan research. Tectonics and gross geomorphology may control the fan setting, but the consensus is that, at least for Quaternary fans, climate appears to have the primary role in causing changes in fan

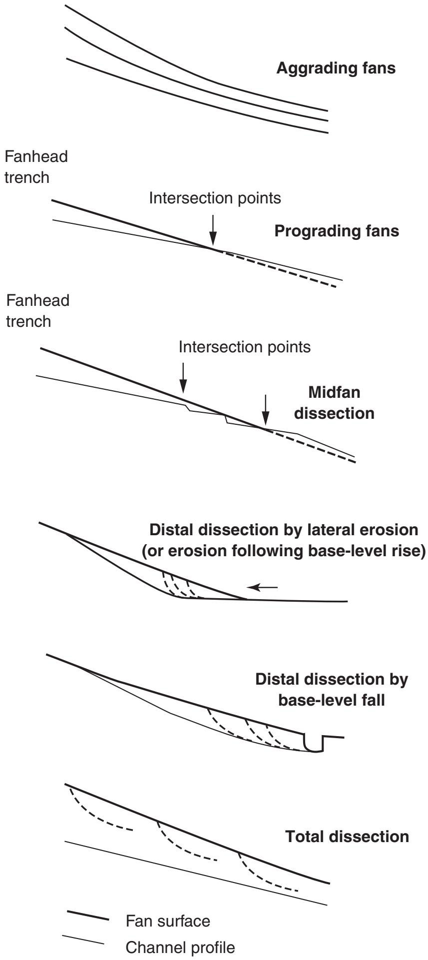

Figure 4 Fan surface and fan channel relationships

behaviour and dynamics (Frostick and Reid 1989; Ritter *et al.* 1995), modified by base-level conditions (Harvey 2002a).

Significant changes in fan processes in response to Quaternary climatic changes have been identified in many areas, including dry regions (e.g. Wells *et al.* 1987; Bull 1991), and humid temperate regions, especially in a PARAGLACIAL context in mountain areas glaciated during the Pleistocene (Ryder 1971).

Alluvial fans are important features within mountain fluvial systems. They act as sediment stores, modifying the transmission of coarse sediments through the fluvial system. They have a profound effect on the buffering/coupling relationships of fluvial systems (Harvey 1997, 2002b). Similarly they preserve a sensitive sedimentary record of environmental change within the mountain source areas.

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ADRIAN HARVEY

ALLUVIUM

Alluvium (neuter of the Latin adjective *alluvius*, meaning washed against) is the term used for the sediments that are deposited by flowing water in river valleys and deltas. Alluvial sediment originates ultimately from the breakdown (weathering) of pre-existing rocks on land. This sediment is then transported downslope by mass wasting, overland water flow, river flow and floodplain flow, and deposited in areas of the river valley where the water flow decelerated. Alluvium is deposited in distinctive landforms (e.g. channel bars, channel fills, levees, crevasse splays, flood basins, fans and deltas). Alluvial sediments are normally stratified gravels, sands, silts and clays, and the texture and stratification of the sediments are determined by the associated landform and the mode of deposition and subsequent erosion. Alluvium has been deposited on the Earth's surface for as long as rivers have existed. The sedimentary characteristics of modern alluvium are used to interpret the origin of ancient alluvium. Alluvium of modern rivers and floodplains commonly is fertile agriculturally, and a source for ground water, sand and gravel. Ancient alluvium commonly contains economically important resources such as water, gas, oil, coal and placer minerals. Reviews of the origin and nature of alluvial deposits are given by Bridge (2002), Carling and Dawson (1996) and Miall (1996).

Nature of transport and deposition of alluvium

Water flow in alluvial river channels and floodplains is turbulent. Turbulent water flow results in relatively coarse sediment (sand and gravel) being transported near the sediment bed as bedload, and finer-grained sediment (sand, silt and clay) being transported within the flow as suspended load. Depending on the sediment transport rate, the bed is normally moulded into various types of bedform, such as ripples, dunes and antidunes. Near-plane beds also occur in sands when the sediment transport rate is relatively high and in gravels at low sediment transport rates. The geometry of ripples is dependent on bed-sediment size. The geometry of dunes and antidunes is related to flow depth. Bars are larger bedforms that occur in channels, and their geometry is controlled mainly by channel width. Ripples, dunes and antidunes may be superimposed on bars.

Deposition of alluvium occurs mainly due to spatial (but also temporal) decrease in water flow velocity (actually bed shear stress) and sediment transport rate. This deposition occurs over a large range of spatial scales in areas of flow expansion and deceleration such as the lee sides of bedforms, at the edge of channels as water moves onto the floodplain, abandoned channels, zones of tectonic subsidence, and where water flows into lakes and the sea (forming deltas). Most deposition occurs during floods when flow velocity, bed shear stress and sediment transport rate are large. If sediment transport rate is large, spatial decrease in sediment transport rate will cause relatively high deposition rate. However, much deposited sediment is subsequently re-eroded during the same or subsequent floods. The coarsest sediments are deposited from bedload in places where bed shear stress is large (typically channels), whereas the finest grained sediments are deposited from suspended load (typically in abandoned channels, flood basins and lakes). Intermediate sediment sizes are deposited from bedload and suspended load. It is common for the size of channel-bed sediment to decrease down valley, primarily due to downstream decrease in channel slope and bed shear stress. It is also common for bed-sediment size to decrease laterally from the channel to the distal edge of the floodplain, also due to decreasing bed shear stress.

Alluvial landforms

Alluvial river channels contain various types of bars, the geometry and evolution of which determine the plan form of the channel. Simple (unit) bars occur in all alluvial channels. In meandering channels, the unit bars combine to give compound point bars on the inside of channel bends. In braided channels, the unit bars combine to give mid-channel compound bars (braid bars) in addition to point bars. As the supply of water and/or sediment increases, alluvial channels change from meandering to braided. Straight channels are rare and occur when the stream is not powerful enough to erode its banks. Channels change position by bank erosion and bar deposition, or by channel diversions. Channels can be diverted within their channel belts by cutoff, and channel belts can be diverted to different positions within their floodplains (avulsion). Channels abandoned by cutoff or avulsion become blocked with bars and eventually become elongate lakes.

Floodplains are the areas adjacent to channels that are inundated with water during seasonal floods. LEVEEs are wedge-shaped accumulations of sediment that form floodplain ridges adjacent to channels. Crevasse channels cut levees in places and pass downstream into lobate sediment accumulations called crevasse splays. Some levees are composed of laterally adjacent crevasse splays. Crevasse splays are fan shaped in plan and contain a system of distributive and or anastomosing channels. The active and abandoned channels, levees and crevasse splays constitute the alluvial ridge that stands above the adjacent flood basin. The alluvial ridge exists because deposition rate is greatest in and around the main channel. The flood basin contains floodplain channels, both ephemeral and permanent lakes, and abandoned channel belts (alluvial ridges).

Alluvial fans and deltas are areas of alluvial deposition that are distinctive because of their plan shapes and distributive and/or anastomosing channels bordered by floodplains. Deltas build into standing bodies of water. If fans build into standing bodies of water, they are referred to as fan deltas. The term terminal fan has been used for fans in arid areas where water flow percolates into the ground before reaching beyond the fan margins. Alluvial fans occur in all climates where a confined channel passes from an area of high slope to an unconfined area of lower slope. The abrupt change of slope results in a downstream decrease in bed shear stress and sediment transport rate, which leads to deposition. ALLUVIAL FANs commonly occur adjacent to fault scarps, and the preservation of fan deposits is enhanced by the subsidence of the hanging wall. Usually, one channel is active on a fan surface at any time, but avulsion is a common process and many wholly or partially abandoned channels occur on fan surfaces. Where fan surfaces are steep (and relatively coarse grained), sediment gravity flows are common depositional processes in addition to water flows.

A RIVER DELTA is a mound of sediment deposited where a river channel enters a body of water (such as a lake or sea) and supplies more sediment than can be carried away by currents in the water body. At the river mouth, the previously confined flow expands and decelerates, depositing its sediment load. The coarse bedload is deposited close to the mouth (as a mouth bar), whereas the finer sediment in suspension is carried further into the water body before being deposited. Currents in the body of water (perhaps associated with tides, wind waves, geostrophic flows or turbidity currents) may subsequently rework and move the deposited sediment. The morphology and sediments of deltas reflect the balance between these different stages of delta formation.

River terraces (see TERRACE, RIVER) are remnants of floodplains, fans or delta plains that have become elevated relative to the modern river and floodplain, as a result of widespread channel incision. Different episodes of incision and deposition can result in a series of terraces of different height, and valley fills with a complicated internal structure.

Alluvial deposits

Depending on the availability of different sediment sizes, channel deposits are usually mainly gravels and sands. Floodplain deposits are mainly sands, silts and clays. Different scales of stratification in alluvial deposits depend on the scale of topographic feature associated with the deposit: ripples form small-scale cross strata (set thickness <30 mm); dunes form medium-scale cross strata (set thickness 30 mm to metres); unit bars form simple sets of large-scale inclined strata (set thickness normally decimetres to metres); compound bars form compound sets of large-scale inclined strata (set thickness metres to tens of metres). Channel fills are composed of bar deposits overlain by lacustrine silts and clays. Channel belts are composed of superimposed bars and channel fills, and are commonly metres to tens of metres thick and hundreds to thousands of metres wide. Levees, crevasse splays and lacustrine deltas may be metres thick and hundreds to thousands of metres long and wide, composed mainly of sands and silts. Floodplain-channel fills typically are up to metres deep and tens to hundreds of metres across. Silty and clayey deposits of flood basins and lakes commonly occur in metre-thick sequences. Floodplain deposits are normally subjected to pedogenesis, and soil horizons are ubiquitous in alluvium.

The nature and degree of soil development varies in time and space as a function of floodplain deposition rate, parent materials, groundwater composition, climate and vegetation.

The proportion of channel-belt deposits (coarse sediments) relative to floodplain deposits (fine sediments) in a valley fill depends on factors such as the frequency of channel-belt diversions (avulsions), the width of the channel belt relative to the floodplain width, the overall deposition rate, and tectonic subsidence or uplift within the valley. High proportions of channel deposits in valley fills typically occur on the upstream parts of alluvial fans where deposition rate and avulsion frequency are locally high, in parts of valleys that are narrow relative to the channel-belt width (e.g. incised valleys), and in areas of the valley where tectonic subsidence has attracted avulsing channel belts. Low proportions of channel deposits in valley fills occur typically where floodplain (valley) width is large relative to channel-belt width (e.g. on delta plains), and in tectonically uplifted parts of floodplains.

As avulsion frequency, relative widths of channel belts and floodplains, overall deposition rate, and subsidence or uplift rate are controlled by climate, eustatic sea-level change, and tectonism, the nature of the valley fill will also be controlled by these factors. Furthermore, spatial and temporal variations in the effects of climate, eustatic sea-level change and tectonism on deposition and erosion of alluvium result in spatial variations in its texture and internal structure. These spatial variations are commonly cyclic.

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SEE ALSO: aggradation; anabranching and anastomosing river; antidune; avulsion; bank erosion; bar, river; bedform; bedload; braided river; channel, alluvial; current, downstream fining; erosion; dune, fluvial; flood; flood plain; point bar; suspended load

JOHN S. BRIDGE

AMPHITHEATRE

Some early studies of curved valley heads of non-glacial origin attributed them to erosion under arid climates, but they are also widespread in humid areas. They occur mainly where a valley extends headward through gently inclined sedimentary rocks, or through dissected volcanic domes, such as those of Hawaii. Unless angular morphology is maintained by strong rectangular fracturing, curved planimetry develops as a strong CAPROCK is undercut either by seepage or by mass failure.

An amphitheatre can be likened to an arch lying on its side, because lateral stresses hold blocks in place on the curved rock face. This is especially so where the dominant stresses in a rock mass are essentially horizontal, and keep the rock face in compression. Amphitheatres thus tend to be more stable than straight cliff lines. The development of the curvature seems to be linked to the three-dimensional distribution of stresses on the rock face. Experimental studies for open-cut mining show that slopes are most stable where the radius of curvature approximates the height on the back wall, but that stability decreases markedly as the radius of curvature increases to about four times the height. Similar relationships occur in many natural amphitheatres. South of Sydney, Australia, 90 per cent of amphitheatres have a radius-to-height ratio below 5:1, with approximately 20 per cent of them below 2:1. The dimensions of amphitheatres are far from random, and are indicative of an equilibrium between form and stress distribution.

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R.W. YOUNG

ANABRANCHING AND ANASTOMOSING RIVER

An *anabranching* alluvial river is a system of multiple channels characterized by vegetated or otherwise stable alluvial islands that divide flows at discharges up to bankfull (Plate 2). The islands may be developed from within-channel deposition, excised by channel AVULSION from extant floodplain, or formed by prograding distributary-channel accretion on splays or deltas. A specific

Plate 2 An aerial view of muddy anabranching channels at South Galway on Cooper Creek in western Queensland, Australia. Standing water is pale grey and the recently wetted channel boundary is darker and about 20–30 m wide. The islands were not quite over-topped by this bankfull flow

subset of distinctive low-energy anabranching systems associated with mostly fine-grained or organic sedimentation are defined as *anastomosing* rivers (Smith and Smith 1980; Knighton and Nanson 1993; Makaske 2001). Neither of these terms now applies to BRAIDED RIVERS where divided flow is strongly stage dependent around bars that are unconsolidated, ephemeral, poorly vegetated and overtopped at less than bankfull. However, some confusion remains because an individual low-flow channel in a braided system is sometimes referred to as an anabranch. The islands in an anabranching river are about the same elevation as the adjacent floodplain, persist for decades to centuries, have relatively resistant banks, and support mature vegetation.

Anabranching bedrock rivers can occur where the individual channels follow joint and fracture patterns. However, bankfull flow is unclearly defined making such rivers difficult to compare to their alluvial counterparts. Channels are often sediment free with pools, cataracts and waterfalls. Van Niekerk *et al.* (1999) found that bedrock anabranching channels on the Sabie River in South Africa have a significantly greater potential to transport sediment than do all the other channel types along that river. At present, relatively little is known about bedrock anabranching systems.

Anabranching is not a mutually exclusive category for it occurs in association with other patterns whereby individual anabranches braid, meander (see MEANDERING) or are straight, and it occupies a wide range of environments, from low to high energy, and in arctic, alpine, temperate, humid tropical and arid climatic settings. Anabranching rivers are more common than has been recognized previously; a total of more than 90 per cent by length of the alluvial reaches of the world's five largest rivers anabranch and it is a particularly widespread river pattern in inland Australia for both large and small rivers. In Europe many rivers used to anabranch but most of these have now been modified to provide more convenient single-thread systems in densely populated and heavily utilized valleys.

Determining the fundamental cause of anabranching remains elusive but it is understood that in some cases, the advantage of anabranching over a single wide channel is that islands concentrate stream flow and maximize bed-sediment transport per unit of stream power, thereby maintaining equilibrium conditions. This occurs particularly where there is little or no opportunity to increase channel gradient (Nanson and Huang 1999) or where vegetation increases channel roughness (Tooth and Nanson 2000). In other words, some anabranching rivers appear to exhibit MAXIMUM FLOW EFFICIENCY and LEAST ACTION PRINCIPLE (Huang and Nanson 2000). However, there are also cases where anabranching is associated with non-equilibrium sediment transport and inefficient flow, exhibiting extensive overbank flooding, the dispersal of sediment over extensive floodplains (Plate 2), and rapid vertical accretion (Makaske 2001; Abbado *et al.* 2003). As with meandering and braiding rivers, it is apparent that anabranching systems can exhibit equilibrium or non-equilibrium behaviour.

Classification

Six types of anabranching river have been recognized by Nanson and Knighton (1996) on the basis of stream energy, sediment size and morphological characteristics: Types 1–3 are lower energy and Types 4–6 are higher energy systems. Figure 5 illustrates the planform expressions for various types of anabranching river. Type 1 consists of *cohesive sediment* rivers (commonly termed anastomosing rivers) with low w/d ratio channels that exhibit little or no lateral migration. Type 2 consists of *sand-dominated island forming* rivers and Type 3 consists of *mixed load laterally active*

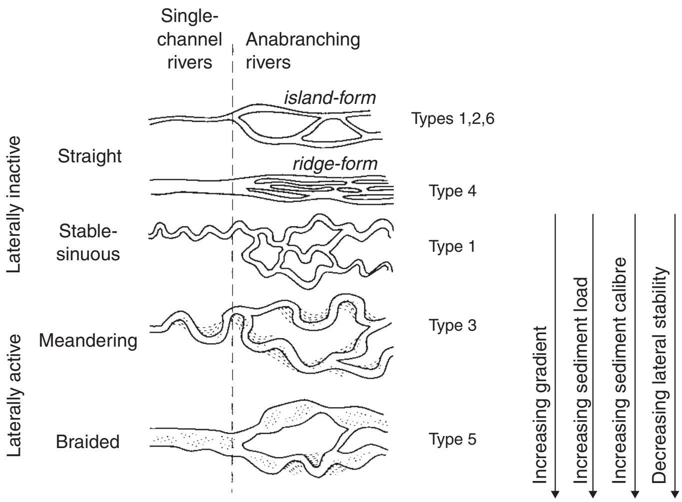

Figure 5 A classification of river channel patterns including single channel and anabranching planforms. Laterally inactive channels consist of straight and sinuous forms whereas laterally active channels consist of meandering and braided forms. The anabranching types are described in the text (after Nanson and Knighton 1996)

meandering rivers. Type 4 consists of *sand-dominated ridge form* rivers characterized by long, parallel channel-dividing ridges. Type 5 consists of *gravel-dominated laterally active* rivers that interface between meandering and braiding in mountainous regions. These have been described as wandering gravel-bed rivers (Church 1983). Type 6 consists of *gravel-dominated stable* rivers that occur as non-migrating channels in small, relatively steep basins.

Anastomosing rivers

Anastomosing rivers are an economically important subgroup of anabranching rivers and consequently have been studied in detail by sedimentologists because of their fine-grained nature and tendency to accumulate a substantial organic (coal) stratigraphy. Anastomosing commonly occurs in the lower fine-textured reaches of rivers, or in depositional basins, where vertical accretion can be rapid and hence their preservation potential is high. Crevasse splays and thick natural levees may be common. Makaske (2001) describes them as forming by avulsion and the islands as having flood basins, but these characteristics are not so apparent in some arid environments (Knighton and Nanson 1993). Modern examples were first described in detail in the alpine and humid environment of the Rocky Mountains of western Canada (e.g. Smith 1973; Smith and Smith 1980) but have subsequently been described in a wide variety of settings including arid environments (e.g. Knighton and Nanson 1993; Gibling *et al.* 1998; Makaske 2001).

Anastomosing river stratigraphy

In rapidly accreting humid settings, peats can accumulate in floodplain lakes and swamps to form coal, and sandy palaeochannels may act as reservoirs for hydrocarbons. However, not all anabranching rivers are rapidly vertically accreting and in arid environments they do not accumulate organics. Makaske (2001) found no standard sedimentary succession for anastomosing rivers, although he described them in three different settings and showed some common characteristics. The Columbia River is an example of the style of stratigraphy in a rapidly vertically accreting humid montane setting with organic-clastic accumulation (Figure 6). Such anastomosing rivers (and delta distributary

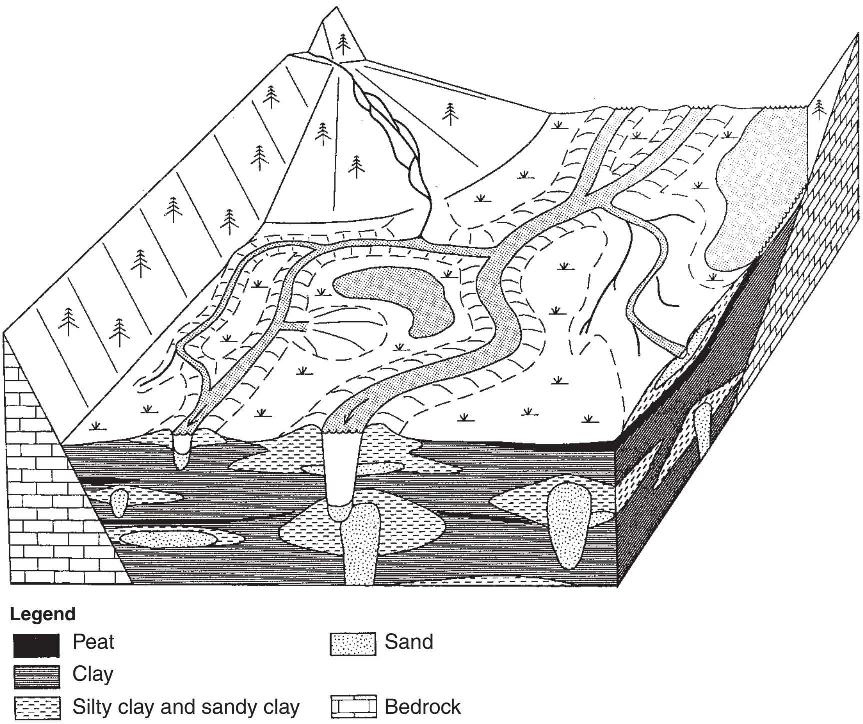

Figure 6 Textural facies model of the upper Columbia River (British Columbia, Canada), a rapidly aggrading anastomosing system in a temperate humid montane setting. Scale is approximately 2 km in width and alluvial thickness ~10 m (after Makaske 2001)

channels) tend to have fixed channels that aggrade with only limited lateral migration, thus generating ribbons or narrow sheets. Many of these deposits lie in rapidly subsiding settings, especially foreland and extensional basins characterized by large sediment flux and low gradients. The avulsion of a major channel into wetlands may generate a splay complex with suites of small, transient anastomosing channels, with the eventual establishment of a stable, single-channel course. In arid environments, alluvial and aeolian deposits can be juxtaposed, whereas in vertically accreting humid environments channel fills are flanked by silty levee deposits, lacustrine clay and coal. However, because it is difficult to show that the palaeochannels formed a synchronous anastomosing network at a single point in time (Makaske 2001), assessing the truly anabranching origin of such stratigraphies may be sometimes an educated guess.

Vegetation

Vegetation plays a major role in the development and maintenance of anabranching rivers. Indeed, it is very likely that truly anabranching rivers did not exist prior to the Devonian Period when the evolution of land plants and their associated role in the weathering of clays and the stabilization of the land surface became important. The establishment and maintenance of channels and islands with stable, often near vertical, banks means that the channels, instead of widening as a simple function of shear stress and limited alluvial strength, maintain narrow, deep and flow-efficient channels. Smith (1976) demonstrated the enormous increase in erosional resistance that plant roots can offer riverbanks. In some dryland rivers anabranching has been shown to increase in intensity below tributary junctions due to irrigation of the often dry channel floor and the greater flow and sediment-transport resistance offered by trees growing on the channel bed (Tooth and Nanson 1999). Such anabranching, resulting from the progressive evolution of within channel bars to ridges, organizes the flow into well-defined multiple channels, narrower in total than the adjacent single-thread reaches (Wende and Nanson 1999; Tooth and Nanson 2000). In certain dryland environments where bankline trees are less dense, then cohesive mud plays an important role in producing stable multiple-channel systems (Gibling *et al.* 1998).

Conclusion

Anabranching characterizes a disparate group of alluvial systems from low-energy organic or fine sediment-textured, to high-energy gravel transporting rivers, and even occurs in bedrock systems. It is a widely represented – even the dominant – style along the world's largest alluvial rivers. Alluvial anabranching rivers can be equilibrium systems that maintain their sediment flux by confining bankfull flows, or non-equilibrium systems that very effectively distribute and deposit excess sediment over extensive depositional surfaces. Anabranching is commonly associated with flood-dominated flow regimes and well-vegetated, erosion-resistant banks. As such they sometimes exhibit mechanisms to block or constrict channels and induce channel avulsion. Some develop as erosional systems that scour channels into floodplains or jointed bedrock, while others build long-lived, stable islands or ridges within existing channels. On deltas they can build floodplains vertically around initially subaqueous channels. Anabranching rivers are commonly laterally stable but individual channels can meander, braid or be straight, and as such they represent a diverse river style.

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SEE ALSO: avulsion; bedrock channel; braided river; floodplain; meandering

GERALD C. NANSON AND MARTIN GIBLING

ANTHROPOGEOMORPHOLOGY

Anthropogeomorphology is the study of the human role in creating landforms and modifying the operation of geomorphological processes such as weathering, erosion, transport and deposition (see, for example, Brown 1970; Nir 1983;

Goudie 1993). Some landforms are produced by direct anthropogenic processes. These tend to be relatively obvious in form and are frequently created deliberately and knowingly. They include landforms produced by: construction (e.g. spoil tips, bunds, embankments), excavation (e.g. road cuttings, and open-cast and strip mines, etc.), hydrological interference (e.g. reservoirs, ditches, channelized river reaches and canals) and farming (e.g. terraces; see Plate 3).

Landforms produced by indirect anthropogenic processes are often less easy to recognize, not least because they tend to involve not the operation of new processes, but the acceleration of natural processes. They are the result of environmental changes brought about inadvertently by human actions. By removing or modifying land cover – through cutting, bulldozing, burning and grazing – humans have accelerated rates of erosion and sedimentation (see SOIL EROSION). Sometimes the results will be spectacular, for example when major gully systems rapidly develop (see ARROYO, DONGA). By other indirect means humans may create subsidence features (Johnson 1991), cause lake desiccation (Gill 1996), trigger mass movements like landslides, and influence the operation of phenomena like earthquakes through the impoundment of large reservoirs (Meade 1991). Rates of rock weathering may be modified because of the acidification of precipitation caused by accelerated sulphate emissions (see SULPHATION) or because of accelerated salinization in areas of irrigation (Goudie and Viles 1998).

Plate 3 Strip lynchets in Dorset, southern England, are a manifestation of the impact that agricultural activities can have on the geomorphology of slopes. Many of the lynchets are the result of ploughing in medieval times

There are situations where, through a lack of understanding of the operation of geomorphological systems, humans may deliberately and directly alter landforms and processes and thereby set in train a series of events which were not anticipated or desired. There are, for example, many records of attempts to reduce coast erosion by important and expensive hard engineering solutions, which, far from solving erosion problems, only exacerbated them (Bird 1979).

As so often with environmental change, it is seldom easy to disentangle changes that are anthropogenic from those that are natural (Brookfield 1999). There has, for example, been a long-continued debate about the origin of deeply incised gullies, called ARROYOs, which developed in the southwestern United States over a relatively short period in the late nineteenth century. Some workers have championed human actions (e.g. overgrazing) as the cause of this erosion spasm, while others have championed the importance of natural environmental changes, noting that arroyo cutting had occurred repeatedly before the arrival of Europeans in the area. Among the natural changes that could promote the phenomenon are a trend towards aridity (which depletes the cover of protective vegetation) or increased frequencies of high-intensity storms (which generates erosive runoff).

Another example of the complexity of causation is posed by a consideration of the potential causes of loss of land to the sea in coastal Louisiana (Walker *et al.* 1987), something that appears to be proceeding at a rapid rate at the present time. Among the factors that need to be considered are the natural ones of sea-level change, subsidence, progressive compaction of sediments, changes in the locations of deltaic depocentres, hurricane attack and degradation by marsh fauna. Equally, however, one has to consider a range of human actions, including the role that dams and levees have played in reducing the amount and texture of sediment reaching the coast, the role of canal and highway construction and subsidence caused by fluid withdrawals.

In many cases, however, as with the USA Dust Bowl in the 1930s, it is a conjunction in time of human actions (the busting of the sod) with a climatic perturbation (a great drought) that produces change.

The possibility that the build-up of greenhouse gases in the atmosphere may cause enhanced global warming in coming decades has many implications for anthropogeomorphology (see

GLOBAL GEOMORPHOLOGY). Increased sea-surface temperatures may change the geographical spread, frequency and wind speeds of hurricanes – highly important geomorphological agents, particularly in terms of river channels and mass movements. Warmer temperatures will cause sea ice to melt and may lead to the retreat of alpine glaciers and the melting of permafrost. Vegetation belts will change latitudinally and altitudinally and this will also influence the operation of geomorphological processes. Changes in temperature, precipitation amounts, and the timing and form of precipitation (e.g. whether it is rain or snow) will have a whole suite of important hydrological consequences. Some parts of the world may become moister (e.g. high latitudes and some parts of the tropics) while other parts (e.g. some of the world's drylands) may become drier. The latter would suffer from declines in river flow, lake desiccation, reactivation of sand dunes and increasing dust storm frequencies.

However, among the most important potential future anthropogeomorphological changes are those associated with sea-level change caused by the steric effect and by the melting of land ice. Low-lying coastal areas (e.g. saltmarshes, mangrove swamps, sabkhas, deltas, atolls) would tend to be particularly susceptible. Moreover, rising sea levels could promote beach erosion, as is suggested by the BRUUN RULE.

Some landscapes – 'geomorphological hot spots' (Goudie 1996) – will be especially sensitive because they are located in areas where it is forecast that climate will change to an above average degree. This is the case, for instance, in the high latitudes of Canada or Russia, where the degree of warming may be 3–4 times greater than the global average. It may also be the case with respect to some areas where particularly substantial changes in precipitation may occur. For example, various scenarios portray the High Plains of the USA as becoming markedly more arid. Other landscapes will be especially sensitive because certain landscape-forming processes are closely controlled by climatic conditions. If such landscapes are close to particular climatic thresholds then quite modest amounts of climatic change can switch them from one state to another.

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A.S. GOUDIE

ANTIDUNE

A symmetrical fluvial BEDFORM produced by near-critical flows, forming in broad shallow channels, and comparable to a sand dune. However, antidunes are more temporary and are less common than dunes. Antidune formation requires a Froude Number (quantifying the relationship between the bedform and flow regime) greater than 0.8, with development often dependent upon channel depth and bed material. Antidunes migrate upstream as sediment is lost from their downstream side more rapidly than it is deposited, though they can also move downstream or remain stationary. Antidunes form directly in phase with standing waves on the water's surface, and are characterized by shallow foresets which dip upstream at an angle of about 10°. They show low resistance to flow and are rare in the rock record, probably due to reworking. Where antidunes are observed in ancient sediments they are characterized by fine, poorly developed laminae.

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STEVE WARD

APPLIED GEOMORPHOLOGY

The application of geomorphology to the solution of miscellaneous problems, especially to the development of resources and the diminution of hazards (Goudie 2001), to planning, conservation and specific engineering or environmental issues (Brunsden 2002). It incorporates what is sometimes called 'ENGINEERING GEOMORPHOLOGY' (Coates 1971).

In the last three decades applied geomorphology has become a much more central and accepted part of the discipline and a variety of texts have now appeared that review its nature (e.g. Hails 1977; Cooke and Doornkamp 1974; Thorne *et al.* 1997).

The reasons for what Jones (1980: 49) calls this 'significant transformation' are various. He cites three main reasons:

1 An increasing awareness on the part of environmental decision-makers as to the complexity of environmental conditions and the significance of geomorphological hazards (e.g. landslides and floods).
2 Demand from engineers for more information on ground conditions for construction purposes and for engineering geomorphological maps.
3 A decreasing level of insularity among geomorphologists and their feeling that they needed to justify their existence in a society that increasingly measured value in terms of practical achievement.

Other reasons have been the change of emphasis in geomorphology towards the study of contemporary processes and changes at the Earth's surface. The second has been the development of more precise techniques for mapping, monitoring and analysis. A third has been growing awareness of the finite limits to some resources and the importance of a seemingly growing number of environmental crises and catastrophes. Growing urban centres face many serious geomorphological hazards. A fourth stimulus has been the development in various countries of the need for environmental impact assessments.

A major new stimulus to applied geomorphological research has been a concern with global environmental change and the potential consequences of global warming. Matters such as the stability of the Antarctic ice sheets, the susceptibility of permafrost to thermokarst development, the sensitivity of coastal wetlands to sea-level rise, and the possible reactivation of sand seas are some of the major issues that have been investigated and for which land management solutions may be required. At a more local scale, human activities are modifying the rate and extent of particular geomorphological processes including soil erosion, salt weathering and river channel form.

Geomorphologists have a variety of applicable skills which, while they may not individually be unique, as a package are distinctive.

1 'An eye for country' and the ability to interpret landscapes and identify landforms.
2 The ability to interpret and produce maps, for these uniquely effective means of imparting spatial information are central to applied geomorphology. GEOMORPHOLOGICAL MAPPING, based on field surveys and the use of topographic base maps and remote sensing techniques, have for long been used by applied geomorphologists. Cartographic skills have been revolutionized in recent years through the use of new technologies, including differential GPS, GIS, DIGITAL ELEVATION MODELS and LIDAR. Maps are especially important for land use planning and zoning.
3 Competence in the use of techniques to measure the operation of geomorphological processes.
4 Appreciation of the relationships between environmental phenomena. This enables applied geomorphologists to see a site in its broader context and to appreciate that change in one place will have ramifications elsewhere. Thus an engineering scheme (e.g. the construction of an erosion control device on a coastline) can have a range of unintended impacts on slope stability or on downdrift beach nourishment.
5 Recognition of the importance of spatial scale. Geomorphologists appreciate, for example, that rates of sediment yield vary according to the area studied and that small erosion plots may give different orders of magnitude of rates than whole catchment studies in a large basin.
6 Recognition that all places are different and that a practice which may be appropriate in one place may not be appropriate in another.

Thus some areas may be peculiarly aggressive while others may be especially sensitive to change. In a permafrost area, for example, there may be profound local differences in permafrost stability because of local soil or microclimatic conditions.

7 Recognition that the landscape is subject to change at all temporal scales and that not only is the present always changing but also that the present is a poor guide to either past or future conditions. An example of such a skill is the recognition of the need to reconstruct long-term discharge records for rivers using a range of dating and sedimentological techniques.

8 Recognition of the importance of human activities and human attitudes. A natural science/social science mix is a unique attribute that will be of particular importance in the field of environmental management (Jones 1980: 70).

Plate 5 The demolition of a railway line in Swaziland, southern Africa, caused by floods associated with a large tropical cyclone. One role of the applied geomorphologist is to undertake post-event surveys in order to ascertain past discharges

The roles of the applied geomorphologist

The various roles of the applied geomorphologist are shown in Table 1.

1 A very basic, but highly important role is to map geomorphological phenomena as a basis for TERRAIN EVALUATION. Landforms, especially depositional ones, may be important sources of useful materials for construction, while maps of slope categories may help in the planning of land use and maps of hazardous ground may facilitate the optimal location of engineering structures.

Plate 4 The main railway line from Swakopmund to Walvis Bay in Namibia illustrates the hazard posed by sand and dune movement. One role of the applied geomorphologist is to identify optimum route corridors and to advise on their management

2 Use landforms as the basis for mapping other aspects of the environment, the distribution of which is related to their position on different landforms. This is important because landforms are relatively easily recognized on air photographs and other types of remote sensing imagery. An important example of the use of landforms as surrogates for other phenomena is the use of landform mapping to provide the basis of a soil map through the use of the CATENA concept and soil toposequences.

3 Recognize and measure the speed at which geomorphological change is taking place. Such changes (Table 2) may be hazardous to humans (e.g. coastal retreat, movement of river bluffs, surges of glaciers). By using sequential maps and remote sensing images, archival information or by monitoring processes with appropriate instrumentation, areas at potential risk can be identified, and predictions can be made as to the amount and direction of change.

4 Assess the causes of observed and measured changes and hazards, for without a knowledge of cause, attempts at amelioration and management may have limited success. There is an increasing need to assess the role that humans are playing in modifying rates

Table 1 The roles of the applied geomorphologist

1	Mapping of landforms, resources and hazards
2	Use of maps of landforms as surrogates for other phenomena (e.g. soils)
3	Establishment of rates of geomorphological change by direct monitoring, use of sequential maps, archives, etc.
4	Establishing causes of change
5	Assessment of management options
6	Post-construction assessment of engineering schemes
7	Post-event evaluations (e.g. palaeodischarges)
8	Prediction of future events and changes

Table 2 Examples of geomorphological hazards

Arid zones	*Coastal*
Dune encroachment	Sea-level change
Soil deflation	Dune blowouts
Arroyo formation	Cliff retreat
Dust storms	Saltmarsh siltation
Fan entrenchment	Coastal progradation
Flash floods	Spit growth and breaching
Salt weathering	
Ground subsidence	
Tundra areas	*General*
Thermokarst formation	Mass movement
	Karstic collapse
Frost leave	River floods
Thaw floods and ice jams	Shifting river courses
	Lake sedimentation
Glacier surges and glacier dams	Soil erosion
	Riverbank erosion
Avalanches	Neotectonic activity
Jökulhaups	

of geomorphological processes, particularly as a result of land-cover changes.

5 Having decided on the speed, location and causes of change, appropriate management solutions need to be adopted. Although the management solution to a particular geomorphological problem may involve the building of an engineering structure (e.g. a sand fence, a sea wall, a check dam, a shelterbelt), these structures may themselves create problems and their relative effectiveness needs to be assessed. The applied geomorphologist may make certain recommendations as to the likely consequences of building, for example, GROYNEs to reduce coastal erosion. Examples of engineering solutions having unforeseen environmental consequences, sometimes to the extent that the original problem is heightened and intensified rather than reduced, are all too common, especially in coastal situations (Viles and Spencer 1995). Management issues involve a consideration of ecological issues, such as when one decides on the most appropriate form for a river channelization scheme, and are likely to become increasingly important as decisions have to be made about how to manage the landscape in the face of global climate change. More and more alternatives are being sought to ecologically injurious 'hard engineering' solutions.

6 Related to environmental management and the use of engineering solutions, is the field of assessment of the success of particular schemes. An audit of performance is required as the basis for formulating best practice.

7 Undertake 'after-the-event' surveys. It is important to put on record the magnitude and consequences of extreme events as a basis for improving engineering designs and land-zoning policies. For instance, establishing the Holocene flood histories of rivers by surveying and dating slack water deposits give an important tool for predicting possible future flood peaks, especially in ungauged catchments.

8 This brings us to the final role of the applied geomorphologist, which is to look forward and to predict. When is a particular glacier likely to surge, how long will it take for an irrigation canal to be blocked by a wandering barchan, when is this slope likely to fail, how quickly will this reservoir be rendered useless through sedimentation, will the surface of a delta be built up by fluvial sediment inputs more quickly than sea-level rises? These are examples of where geomorphologists can help to answer questions about the future. Their answers can be based on studies of the past rate of operation of geomorphological processes or by developing their modelling capability.

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A.S. GOUDIE

ARCH, NATURAL

Natural arches are formed when weathering, together with mass collapse, and in arid areas with wind erosion, creates a tunnel through a slab of rock. They can thus be distinguished from NATURAL BRIDGES which are formed by fluvial or marine erosion. They most commonly occur in sandstone, which has sufficient permeability to provide the seepage that promotes weathering, yet which has the necessary cohesion for an arch to develop. Arches are most numerous where long and closely spaced joints have been eroded to form narrow fins of rock that are readily pierced by weathering. These characteristics, and thus a very high concentration of arches, occur in the Entrada and Cedar Mesa Sandstones of the Colorado Plateau.

In strongly bedded rock, widening of the initial tunnel may result in the development of a long slab or lintel. The load of the undercut rock creates tensional stress on the lower face of the slab. If the space continues to grow, the stress may exceed the tensile strength of the rock, causing the slab to collapse. An upward curving form, rather than a slab, will develop where there are curved joints in the rock, or, more commonly, where concave stress patterns in the undercut rock result in minor failures or surface spalling. The curved form of the true arch is much more stable than a slab, because the load is transmitted to the abutments, and virtually the entire structure is in compression. This is so even when the arch is split by joints, for compressive stress on each of the joint-bounded blocks keeps them in place.

Natural arches may take various forms, but will remain stable provided the load is transmitted into the abutments. This condition is met so long as the thrust line of the load remains within the arch. Arches are therefore very stable features. However, continued erosion may result in an unstable form, and the arch may then collapse by folding in on itself at several hinge points. Erosional weakening of the abutments into which the load is transmitted can also cause an arch to collapse. Conversely, rock pinnacles transmitting a vertical load down into the abutments, or natural buttresses supporting them laterally, increase the stability of the arch.

Especially where joints are widely spaced, the hollow developed in a cliff may not penetrate through the rock mass. Instead of a true arch, an alcove or apse develops. These are much more widespread than arches, but they form in essentially the same manner, being particularly well developed where seepage issues from massive sandstone.

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R.W. YOUNG

ARÊTE

A landform composed of a fretted, steep-sided ridge that separates valley or cirque GLACIERS. Arêtes are the result of glacial undercutting – basal sapping – of rock slopes. They are common whenever mountains and peaks rise above glaciers, as in the case with NUNATAKS.

A.S. GOUDIE

ARMOURED MUD BALL

Roughly spherical lumps of cohesive sediment which generally have a diameter of a few centimetres (Bell 1940). They are also called mud balls, mud pebbles, pudding balls, till balls and clay balls. Many examples are lumps of clay or cohesive mud that have been eroded by vigorous currents from stream beds or banks. They often occur in areas of badland topography and along ephemeral streams, but can also be found on beaches (Kale and Awasthi 1993), in tidal channels, and as ice-transported debris dumped on the seafloor (Goldschmidt 1994).

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A.S. GOUDIE

ARMOURING

'Armouring', the process whereby a clastic deposit develops a surface layer that is coarser than the substrate, is most commonly associated with warm deserts and gravel-bed rivers (see BOULDER PAVEMENT, STONE PAVEMENT, FLUVIAL ARMOUR). DEFLATION, the removal of fine-grained material by wind; the winnowing of fines by surface wash (see SHEET EROSION, SHEET FLOW, SHEET WASH); the upward migration of coarse particles, as a result of alternate wetting and drying and the associated swelling and shrinking of fine debris, or FREEZE–THAW CYCLE activity at high altitudes; the upward displacement of gravel clasts as fines accumulate; and the preferential weathering and breakdown of coarse debris at depth have all been proposed as mechanisms that produce concentrations of coarse particles at the ground surface in desert environments (Cooke 1970; Dan *et al.* 1982; McFadden *et al.* 1987). The process(es) operating in any given location depend on climate, geomorphic setting, and the nature of the clastic particles and local soils. The gravel clasts involved may be produced by mechanical weathering of the local bedrock, or be of fluvial origin. In rivers, armouring may involve the concentration of coarse clasts at the base of the active layer or the preferential winnowing of finer sediment from the surface during degradation; and vertical winnowing during active BEDLOAD transport which compensates for the disparity in mobility between coarse and fine particles (Andrews and Parker 1987; Parker and Sutherland 1990). Size segregation which produces concentrations of large particles at the surface, also occurs in gravity-driven, granular mass flows, including DEBRIS FLOWs and PYROCLASTIC FLOW DEPOSITs. Segregation mechanisms include size percolation, in which fine grains infiltrate beneath coarser particles; size exclusion, where coarse grains are excluded from narrow, convective downwellings; and cascading segregation, in which larger particles roll more rapidly downslope than small ones (Shinbrot and Muzzio 2000; Vallance and Savage 2000).

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BASIL GOMEZ

ARROYO

An incised valley bottom, particularly in the western USA (Figure 7), where many broad valleys and plains became deeply incised with valley-bottom gullies (arroyos) over a short period between 1865 and 1915, with the 1880s being especially important (Cooke and Reeves 1976). The arroyos can be cut as deeply as 20 m, be over 50 m wide and tens or even hundreds of kilometres long. There has been a long history of debate as to the causes of incision (Elliott *et al.* 1999) and an increasing

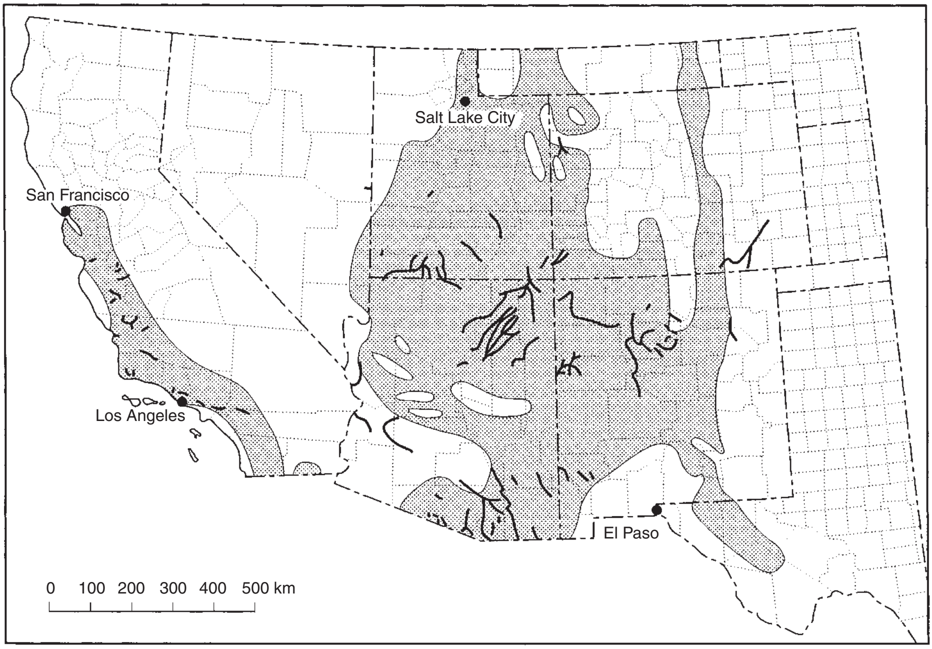

Figure 7 The distribution of arroyos in the southwestern USA (shaded), showing the course of some large examples (dark lines)

appreciation of the scale and frequency of climatic changes in the Holocene (McFadden and McAuliffe 1997), which could have led to changes in channel and slope behaviour. For example, Waters and Haynes (2001) have argued that arroyos first appeared in the American south-west after *c*.8,000 years ago, and that a dramatic increase in cutting and filling episodes occurred after *c*.4,000 years ago. They believe that this intensification could be related to a change in the frequency and strength of El Niño events.

Many students of this phenomenon have believed that human actions caused the entrenchment, and the apparent coincidence of white settlement and arroyo development in the late nineteenth century tended to give credence to this viewpoint. The range of actions that could have been culpable is large: timber-felling, overgrazing, cutting grass for hay in valley bottoms, compaction along well-travelled routes, channelling of runoff from trails and railways, disruption of valley-bottom sods by animals' feet, and the invasion of grasslands by scrub.

On the other hand, study of the long-term history of the valley fills shows that there have been repeated phases of aggradation and incision and that some of these took place before the influence of humans could have been a significant factor. Elliott *et al.* (1999) recognize various Holocene phases of channel incision at 700–1,200 BP, 1,700–2,300 BP and 6,500–7,400 BP.

A climatic interpretation was advanced by Leopold (1951), which involved a change in rainfall intensity. He indicated that a reduced frequency of low-intensity rains would weaken the vegetation cover, while an increased frequency of heavy rains at the same time would increase the incidence of erosion. Balling and Wells (1990), working in New Mexico, attributed early twentieth-century arroyo trenching to a run of years with intense and erosive rainfall that succeeded a phase of drought conditions in which the protective ability of the vegetation had declined. Large floods have also been important causative agents (Hereford 1986). Erosion and entrenchment result from a larger flood regime, with streams having a large sediment transport capacity. With lower flood regimes, a reduction in channel width and sediment storage occur, but if there are no floods, no alluviation of floodplains is possible. It is also possible, as Schumm *et al.* (1984) have pointed out, that arroyo incision could result from neither climatic change nor human influence. It could be the result of some intrinsic natural geomorphological threshold (see THRESHOLD, GEOMORPHIC) (such as stream gradient) being crossed. Under this argument, conditions of valley-floor stability decrease slowly over time until some triggering event initiates incision of the previously 'stable' reach.

It is possible that arroyo incision and alluviation result from a whole range of causes (Gonzalez 2001), that the timing of events will have varied from area to area and that individual arroyos will have had unique histories.

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A.S. GOUDIE

ASPECT AND GEOMORPHOLOGY

As the sun moves across the sky, through the course of each day and through the seasons, the intensity of short wave radiation at a point on the hillside changes. At night, there is little radiation. In the daytime, radiation is greatest when the sun is un-obscured, and not reduced by cloud cover or where the hillside is shaded by surrounding hills. Because, north of the equator, the sun is highest in the sky towards the south, sunny south-facing slopes receive more short wave radiation than north-facing slopes, while east- and west-facing slopes receive intermediate amounts, east-facing slopes receiving more in the mornings and west-facing in the evenings. In the southern hemisphere, relationships are exchanged, north-facing slopes receiving most radiation, although east-facing slopes still face the morning sun.

Some solar radiation is lost in passing through the atmosphere, partly through the scattering which gives blue sky light, and much more if there are clouds in front of the sun. The radiation from both a cloudy sky and a clear blue sky is diffuse, and comes from all directions, although some light is lost by shading in deep valleys. The direct beam of the un-obscured sun is strongly directional, and its intensity on the surface is directly proportional to the cosine of the angle between the sun's rays and a perpendicular to the slope surface. Thus solar radiation is highest where rays fall squarely on the surface, and is greatly reduced when the rays graze the surface.

The sun's path through the sky changes in a regular way through the year, so that the amount of radiation on a hillside can be computed trigonometrically, from the latitude, the slope gradient and the direction in which the slope faces.

The sun's azimuth Φ (bearing to the sun's position in the sky) and elevation θ (angle above the horizon) can be calculated with reasonable accuracy as:

$$\sin\theta = \sin\lambda\sin\beta - \cos\lambda\cos\beta\cos\gamma$$

$$\tan\Phi = \frac{\cos\beta\sin\gamma}{\sin\lambda\cos\beta\cos\gamma + \cos\lambda\sin\beta}$$

where λ is the latitude in degrees North, β is the sun's declination ~ $-23.5\cos J$ on Julian day J (0–360) and γ is $15\,h$ at hour h (0–24 hr local sun time). Even under a clear sun, some light is diffused (about 15 per cent under unpolluted skies) to provide blue sky light. Corrections must also be made for cloudiness and shading by any hills which form the local horizon. Making these calculations, Figure 8 shows that the difference in radiation received from clear skies on north- and south-facing slopes is greatest at about 60° latitude, but because cloudiness also increases with latitude on the continents from 30°, particularly in summer, the actual difference in radiation received is greatest at latitudes of 30°–40°.

Aspect affects geomorphology through the contrasts in radiation, most strongly between north- and south-facing slopes, which leads to differences in hydrology and sediment transport rates. Table 3 summarizes the main differences for the northern hemisphere, and north and south should be consistently exchanged for the southern hemisphere.

The effect of aspect differences is generally to create differences in the intensity of geomorphic processes between the two opposing hillsides. For example the greater radiation on south-facing semi-arid slopes increases evapotranspiration rates, so that water stress occurs in vegetation more quickly after rain. As a result, the vegetation cover is sparser and the species more drought adapted. Sparse vegetation encourages greater crusting of the soil surface, more overland flow runoff and more erosion by wash erosion. On north-facing slopes, soil moisture is maintained after rain for a longer period, so that humid vegetation can grow, usually providing greater ground cover, and better conditions for soil accumulation. Although these conditions improve infiltration rates and reduce overland flow and wash erosion, they can also provide better conditions for mass movements due to the greater depth of soil and higher moisture content.

In the short term, an increase in erosion may lead to steepening of the slope profile, but the longer term implications, as slope profiles approach some form of equilibrium, are less clear, although process differences due to aspect are commonly associated with ASYMMETRIC VALLEYs. Where there is pronounced slope asymmetry, short steep slopes on one side of the valley are matched by longer and gentler slopes on the opposite side. Two factors influence the form of the asymmetric valley cross-section. First, sediment transport depends on both slope length and slope gradient, so that the steeper slope does not necessarily deliver the more sediment. Second, at equilibrium, the valley form may not only be cutting vertically downwards, but may also be migrating laterally. For both these reasons, the hillside with the more intense process activity, due to aspect differences, may not become the gentler slope to compensate for its more intense geomorphological activity. Observations of semi-arid slopes generally suggest that radiation differences tend to maintain steep bedrock slopes on south-facing aspects, and gentler slopes mantled with soil and vegetation on north-facing aspects,

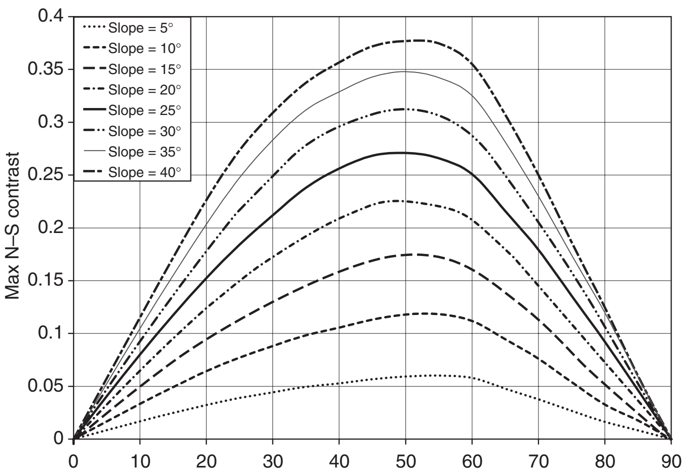

Figure 8 Difference in total annual radiation between north and south-facing slopes for clear sky conditions

Table 3 Summary of effects of aspect differences

Climatic regime	North-facing	South-facing	Geomorphic impact
Very cold (arctic or high altitude)	Permanently frozen	Some freeze-thaw	Greater solifluction and other activity on S-facing slopes
Moderately cold	Some freeze-thaw	Mainly unfrozen	Greater disturbance of vegetation and solifluction on N-facing slopes
Moist temperate	Cooler and moister	Warmer and dryer	Where water is not limiting, differences due to aspect are weak
Warm semi-arid	Cooler and moister	Warmer and dryer	S-facing slopes have sparser and more xeric vegetation, and greater runoff and erosion

although the strength of asymmetry is affected by a number of other factors, particularly geological structure and the meandering activity of rivers.

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MIKE KIRKBY

ASTROBLEME

The term *astrobleme* (literally 'star-wound') was introduced by Robert S. Dietz (1960) in reference to ancient erosional scars, usually circular in outline, that form on the Earth's surface through the impact of a cosmic body. This origin was recognized because of the presence of highly disturbed rocks that display evidence of intense shock (Dietz 1961). In the early debates over the origins of these features, it was not clear whether the intense pressures responsible for the disturbed rocks resulted from a bolide (an exploding meteor or comet) or from a volcanic explosion. Structures formed in the latter manner were termed *cryptovolcanic* by Branco and Frass (1901). However, the nongenetic term, *cryptoexplosion structure* (Dietz 1959), is preferred when the origin is uncertain. Nevertheless, modern research methods can nearly always confirm or reject an origin by meteor or comet impact.

The sites of relatively recent impacts on Earth, Quaternary to late Tertiary in age, will generally preserve the morphology of impact craters that is so commonly observed on the surfaces of other rocky planetary bodies in the solar system. The lack of long-term preservation of distinctive crater morphologies on the Earth's surface is the result of long-acting and relatively rapid erosional and depositional processes, when compared to circumstances on the other planetary objects. The ancient eroded impact structures of Earth (Plate 6) include circular features that are much larger than the better preserved, young impact craters. Debates over the cryptovolcanic versus impact origin of these large features raged until about the 1960s, when mineralogical studies confirmed that one of these structures, the Ries Kessel in Germany, was clearly the result of a large impact.

During the impact cratering process, immense pressures are imparted on target rocks by the high-velocity projectile. The highest pressures vaporize and melt rocks upon their release. Indeed, some large astroblemes, like Ries Kessel, are associated with huge amounts of impact melt, which early workers found difficult to distinguish from igneous rocks. Somewhat lower pressures are responsible for the metamorphic alteration of quartz to coesite and stishovite, minerals which do not form in the tectonic and volcanic processes of Earth's interior. Even lower pressures produce distinctive planar features in crystals, shocked quartz, and a distinctive cone-in-cone fracture pattern in target rocks, called *shatter cones*. The study of such features, along with their structural and geological settings, has led to the discovery of

Plate 6 Central uplift of the deeply eroded Gosses Bluff impact structure, an astrobleme in central Australia. The bluff comprises a ring of resistant sandstone, about 5 km in diameter, that was uplifted in the centre of a much larger transient crater created during the early Cretaceous (Milton *et al.* 1972). The larger structure has a diameter of about 22 km, but it is has been eroded to a nearly level plane. An ancient, higher planation surface bevels the crests of the sandstone ridges that mark the central uplift

well over a hundred terrestrial astroblemes over the last several decades.

Perhaps the most famous astrobleme is the Chicxulub structure, which is buried beneath cover rocks at the northern end of the Yucatan Peninsula, Mexico (Hildebrand *et al.* 1991). The recognition of this feature and its significance illustrates the highly interdisciplinary character of planetary science studies in application to the Earth. The story begins with the discovery in the late 1970s of an enrichment in the element iridium in a 3-cm thick layer of clay at the Cretaceous–Tertiary boundary in a thick section of marine sediments at Gubbio, Italy (Alvarez *et al.* 1980). This geochemical anomaly led the discoverers to propose that a 10-km diameter comet or asteroid collided with Earth 65 million years ago, ending the Cretaceous era and causing one of the most extensive mass extinctions of organisms in geological history, including the demise of the dinosaurs. This was indeed a provocative hypothesis, of immense potential importance to our understanding of Earth history. How could it be verified?

The iridium anomaly was subsequently identified at numerous other Cretaceous–Tertiary boundary sites around the world. Associated with the iridium were other, somewhat exotic elements in concentrations typical for chrondritic meteorites, as would be expected from the composition of the impactor. Also found were shocked quartz grains, stishovite, coesite and small glass spherules. The latter are interpreted as microtektites. Long considered a geological curiosity, relatively large, pebble-sized tektites have been found over extensive surfaces in local regions. They are clearly melted silicates, but their streamlined shapes showed that they had fallen through the atmosphere. Modern understanding of impact cratering mechanics shows that tektites are droplets of impact melt that achieve widespread ballistic dispersal from very large impact events.

The geochemical evidence all pointed to an object that would have produced a crater about 200 km in diameter, which was considerably larger than any astrobleme that had yet been identified on Earth. By following various indicators of proximity to the impact source, including tsunami deposits, tektite sizes and other features of world Cretaceous–Tertiary boundary deposits, the assembled evidence all pointed to the Caribbean and Gulf of Mexico as the likely target area. Interest then moved toward a previously obscure circular structural anomaly in northern Yucatan. The Chicxulub feature is about 180 km in diameter. Though buried, it has surface expression in a ring of cenotes (karstic sinkholes), and it is well displayed in geophysical surveys of the subsurface structure.

The discovery of astroblemes is accelerating. New features are being found on the ocean floor, aided by the extensive exploration for hydrocarbon resources. The techniques for identifying these anomalous forms make use of classical geomorphological reasoning. Moreover, it is now clear that impact cratering, the most prevalent geomorphological process on the rocky planetary bodies of the solar system, is not so rare on Earth as was once believed. It is just that the immense timescales involved for the larger impacts means that their landform consequences mostly appear as eroded, buried, and/or exhumed features that are intimately associated with the Earth's long-term geological record.

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SEE ALSO: crater; cryptovolcano; extraterrestrial geomorphology

VICTOR R. BAKER

ASYMMETRIC VALLEY

In very few valleys are the profiles of the opposite sides exact mirror images about the axis of the thalweg; the geomorphological definition of valley asymmetry, however, requires substantial differences in the shape and/or steepness of the two hillsides. This asymmetry may be localized, e.g. where a meander creates a river cliff opposite a slip-off slope, or valley-wide, e.g. in the case of the UNICLINAL SHIFTING characteristic of scarp and vale scenery. The ultimate in asymmetry is the case of 'one-sided' valleys, such as those of glaciated regions where the missing side was once provided by an ice sheet.

Asymmetry can, then, be the product of a whole range of circumstances relating to the orientation of valley axes and hillsides with respect to both the underlying geology and the past and present subaerial processes. Kennedy (1976) lists eight factors which have been considered to produce valley asymmetry: Coriolis force; differences in insolation and precipitation receipts; differences in slope dimensions; variable lithology; geologic structure; warping; evolution of the drainage net; and glaciation. Of these the role of geologic structure and of aspect-induced variations (see ASPECT AND GEOMORPHOLOGY) in microclimate are the two most commonly attributed causes of asymmetry.

To deal with geologic structure: faulting is evidently capable of producing dramatic asymmetry, either by opposing a fault scarp to a lower-angle hillside, or by creating hillsides with contrasting lithologies. More generally, it is accepted that the low-angle dips of domes such as the English Weald can lead to preferential down-dip migration of rivers, in the process of uniclinal shifting, resulting in broad and broadly asymmetric valleys. Whilst this is a widely observed geologic control, the question of any more general influence exerted by the dip of beds on the movement of stream channels has never been fully explored. M.J. Selby's ROCK MASS STRENGTH classification includes the dip of joints (and bedding planes), but his concept of strength-equilibrium slopes excludes those undercut by streams (1993: 104).

Far more attention has been directed towards the role of microclimatic variability and the asymmetry of slope processes which results. This was explicitly tested by A.N. Strahler (1950) in his quantitative investigation of the Davisian explanatory trio of 'structure, process and stage'. Working in the Verdugo Hills, California, Strahler found that marked vegetation contrasts between north- and south-facing hillsides were not reflected in significant angular differences. This study was extended and refined by M.A. Melton (1960) who revealed statistically significant asymmetry associated both with profile orientation and with the location of stream channels in the Laramie Mountains, Wyoming; the steepening of undercut profiles was shown to be additively linked to that associated with slope aspect (north-facing steeper).

Kennedy (1976) summarizes evidence for the presence or absence of localized and valley-wide asymmetry in seven areas of North America, ranging from 69°N to 31°N. There is no simple pattern, with the exception of the greater prevalence of valley-wide asymmetry in basins whose axes trend east-west, rather than north-south. This suggests strongly that it is the radiation balance, rather than differential precipitation inputs – at least in these cases – which is crucial to the development of process asymmetry. What is of particular interest, however, is the finding (Kennedy and Melton 1972) that an area of modern permafrost (the Caribou Hills, Northwest Territory) shows distinct, topographically determined cases where either north-facing or south-facing slopes are steeper. This must cast some doubt on the persistent attempts (cf. French 1996) to identify distinctive 'periglacial' asymmetry in terms of the orientation of steeper slopes. Kennedy found steeper north-facing slopes as far south as Kentucky (38°N), where it would seem improbable that they represent any legacy of periglaciation.

If there is any generalization to be made about the role of aspect in inducing valley-wide asymmetry, it is probably that it will develop in cases where the overall moisture balance is in some sense marginal: where this is the case, small topographic differences (or – cf. Schumm 1956 – lithologic ones) may create relatively dramatic variations in infiltration, runoff and mass movements and, ultimately, angular differences. That said, one must largely agree with Selby's assessment: 'few [studies] are based on critical examination of all slope units . . . and even in those that are, it has proved impossible to relate hillslope asymmetry to processes, because hillslopes develop over long periods' (Selby 1993: 289–290).

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SEE ALSO: aspect and geomorphology; cross-profile, valley; rock mass strength

BARBARA A. KENNEDY

ATOLL

Atolls are generally sub-circular rings of CORAL REEF surrounding a lagoon with no dry land other than occasional islands (called *motu*) made from sand and gravel-sized detritus thrown up on the reef during storms (Nunn 1994). The word 'atoll' should strictly be applied only to the reef and lagoon (as it is here) but is sometimes used more loosely to refer to *motu*.

On first encounter, it may come as a surprise how ancient many atolls are. In the Pacific, where some of the world's oldest atolls exist, many have reef foundations dating from at least the Oligocene. It may be even more of a surprise to hear how apparently strong such organic structures are, remaining intact despite the continuous buffering of storm waves, earthquakes and even nuclear weapons tests. Yet as we learn more about the structural history of such islands, so it is becoming clear that atolls do, even without such stresses, occasionally experience catastrophic collapses. Thus Johnston Atoll in the central Pacific, where the US chemical weapon stocks are being destroyed, lost its southern flank in a series of huge landslides predating its discovery by humans. On the other hand, part of Moruroa Atoll in French Polynesia, where 98 subterranean tests of nuclear bombs were carried out between 1981 and 1991, has subsided as a direct consequence and growing concern exists about the stability of the remainder – and the possibility of radioactive residues leaking out into the ocean from the test chambers (Keating 1998).

Atoll origins

The classic exposition of atoll origins was that given by Charles Darwin who, having reached Tahiti in 1835 on the *Beagle*, climbed the slopes above Papeete and, looking across at neighbouring Mo'orea Island surrounded by its barrier reef, realized that if the island disappeared, only the REEF would remain. Thus an atoll would be formed. So, even before he had seen an atoll, Darwin set out his Theory of Atoll Development which involved the upward growth of a coral reef in response to the subsidence of its foundations (Darwin 1842). Darwin figured, with considerable prescience, that it was the tendency of ancient volcanic islands in the world's oceans to subside but that their coral fringe could stay alive only if it was able to grow upwards at the same rate. Thus modern atoll reefs (and FRINGING REEFS and CORAL REEFS) had only veneers of living coral growing atop a coral framework composed largely of the skeletal remains of dead hermatypic (reef-building) corals.

For Darwin, atoll lagoons were places where the volcanic foundations of atolls were buried by reef detritus washed over the reef during storms. Organic and mechanical forces combined to make these lagoon sediments finer over time.

What Darwin was unaware of was that sea level had oscillated with amplitudes of 100 m or more during much of the past few million years and that, although this fact did not invalidate his basic model (which is still regarded as essentially correct), sea-level change needs to be incorporated into models of atoll formation. During every period of low sea level, atolls would be converted into high limestone islands analogous to modern Niue Island in the central Pacific and others. The surfaces of these limestone islands would be reduced by KARST erosion and, when sea level rose once again at the end of the low sea-level stand, the reef would begin growing on the reduced surface (Purdy 1974).

It is worth taking a closer look at what would happen when the reef began growing once again on these surfaces during postglacial periods, taking the last as an example. In places where oceanographic and other conditions were most favourable, the upward growth of coral reef was able to 'keep up' with sea-level rise. Yet in most places, it seems that coral could not grow upwards as fast as sea level rose and that only later did the reef surface 'catch up' with sea level. In some cases, reef upgrowth was altogether too slow to keep pace with sea-level rise and the reef 'gave up' resulting in the formation of a drowned atoll (Neumann and MacIntyre 1985). The presence of drowned atolls in many parts of the Pacific and Indian oceans may be a result of Holocene reefs 'giving up' an unequal race as well as their latitude, a proxy for calcium carbonate production (Grigg 1982) and other conditions (Flood 2001).

Atoll forms

It seems most likely that the aerial form of any atoll reflects the subsurface form of the island from which it grew most recently (Purdy 1974). But within that general principle, there is considerable variation of atoll form which cannot be so readily explained.

Like barrier reefs, atoll reefs tend to be broader and more biotically diverse along their windward sides. These are also the places where atoll *motu* are generally more abundant. Thus atolls in the central Pacific easterly wind and swell belt tend to have broader reefs and more (extensive and continuous) *motu* along their eastern sides than their western sides. In contrast, Diego Garcia Atoll in the central Indian Ocean, which experiences a reversal in swell direction every six months, has a symmetrical reef with a continuous *motu*.

On some atolls, particularly those with completely enclosed lagoons, *motu* are extending lagoonwards and beginning to fill in lagoons. Some islands in Tuvalu, such as Nanumaga, which have only a few small lakes and depressions in their central parts are thought to have formed in this manner.

Humans and atolls

It is clear that most atolls only became habitable in the late Holocene because a fall of sea level exposed the tops of atoll reefs which then became foci for the accumulation of sediment dredged up from reef-talus slopes by large waves to form *motu*. Thus the existence of atoll *motu* is clear evidence for the occurrence of a higher-than-present sea level during the middle Holocene, about 4,000 cal. yr BP (Nunn 1994).

Humans have occupied many atolls continuously since that time but stress comes today from many sources. Not only has atoll life become more difficult as populations have increased and demands on naturally resource-poor environments have become more complicated, but today the fabric of atoll islands is threatened by sea-level rise. The rise of sea level during the twentieth century has caused erosion along many atoll shorelines, although often a direct link has been difficult to demonstrate because of a lack of understanding of lagoon sediment dynamics and because of the construction of artificial structures like causeways to link *motu*. In this regard, the effects of creating or enlarging reef passes to enable larger vessels to enter atoll lagoons has created problems for some atoll communities (Nunn 1994).

For the future, there is a widespread perception that the projected rise in sea level in the twenty-first century will result both in the comprehensive destruction of many atoll islands and in many atoll dwellers becoming environmental refugees in consequence. There is certainly cause for concern; one of the best geomorphological studies of recent years (Dickinson 1999) showed that the sea was currently attacking the lithified foundation of Funafuti Atoll in Tuvalu but that soon it was likely that this level would be overtopped and the sea would find itself eroding only the unconsolidated cover of the *motu* resulting in their rapid removal.

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PATRICK D. NUNN

AVALANCHE BOULDER TONGUE

Avalanche boulder tongues are distinctive accumulations of coarse debris resulting from the long continued, downslope transport of debris by snow avalanches (see AVALANCHE, SNOW). Two basic forms were identified (Rapp 1959). Fan tongues are thin veneers of angular debris in the avalanche runout zone. Many larger boulders and vegetation have a scattered surface cover of smaller 'perched' boulders that have been let down in precarious positions from an ablating snow cover. These fans may extend for several hundred metres across slopes of as little as 8°. Similar low-angled fans may also result from the activity of SLUSHFLOW in subarctic environments.

Where avalanches run across accumulations of loose debris (e.g. SCREE slopes below major couloirs) they erode loose debris from the upper surface of these slopes and, redistributing it downslope, produce a raised tongue of debris extending from the base of the original deposit. These 'roadbank' tongues have a pronounced basal concavity and are often asymmetric in cross section with a smoothed bevelled top, flanked by steep side slopes and a lobate front.

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SEE ALSO: hillslope, form; hillslope, process; mass movement; slushflow

BRIAN LUCKMAN

AVALANCHE, SNOW

Controls and characteristics of snow avalanches

In mountainous areas above the snowline, topographical controls result in snow avalanches recurring in locations referred to as avalanche paths. The paths are conventionally divided into the starting zone, track and runout zone. Starting zones occur at elevations above the winter snowline, have slopes in the range of 30° to 45° and are generally lee to the main storm wind directions. Below local treeline elevation, presence of forest cover also influences avalanche location while some secondary topographic factors such as slope form and roughness also play a role. Downslope convexities and transitions from anchoring points to smooth slopes are zones of tension often reflected in avalanche starting points. Across slopes, concavities (bowls and gullies) provide local snow accumulation areas. Tracks and runout zones are at lower angles. The extent of avalanche influence can be determined by their effect on vegetation and contribution to fan-shaped landforms (see Plate 7).

In addition to highest frequencies in winter and spring, avalanche timing is often related to storm events though strong melt may also be significant. Avalanches may be classified in relation to storms as either direct action or climax. The

Plate 7 An avalanche path in the Canadian Rocky Mountains showing a broad bowl-type starting zone, a track devoid of vegetation and a fan-type runout zone

former are initiated by storms (particularly heavy snowfall) and involve only the new snow while the latter owe more to instabilities that develop within the snowpack over periods of at least several days. Climax avalanches may be triggered by new snowfalls but they involve old as well as new snow and are common in spring when the snowpack may contain significant quantities of liquid water.

Snow avalanches are conventionally described in terms of a number of criteria including type of snow, form of motion, snow wetness and depth of failure. The type of snow effects the form of release, such that loose snow avalanches form in cohesionless dry or wet snow, beginning from a point and broadening downslope, while slab avalanches resulting from the existence of a strong layer of snow over a weakly resistant layer, propagate first in a line across the slope. The form of movement after initiation depends on slope steepness, roughness and the nature of the snow. For smooth, relatively gentle slopes and/or wet snow, movement is by flowing in contact with the surface. If the snow is dry and slopes steep and rough, then airborne flow (a powder avalanche) is likely, though most powder avalanches also contain a surface flowing component. Although dry avalanches will generally travel more quickly than wet snow avalanches, wet snow avalanches are capable of transporting considerable quantities of debris and are more likely to be full-depth avalanches. The extreme case of wet avalanches are slush avalanches or SLUSHFLOWS.

Snow avalanches and landforms

Studies of the accumulation of debris transported by avalanches in Scandinavia (Rapp 1960) and Canada (Luckman 1977, 1988; Gardner 1983) show that they may contribute up to several tens of mm a^{-1} of accretion on debris slopes with higher rates near the apex. Erosion also occurs

Table 4 Geomorphic effects of snow avalanches

Direct effects				
	Erosional		Depositional	
Bedrock abrasion, transport	Avalanche impact at breaks of slope		Surface sediment redistribution	
Chutes	*Pits and pools*	*Mounds and ramparts*	*Boulder tongues*	
		Debris tails	*Road-bank*	*Fan*
Indirect effects				
Sediments	Soils	Hydrology, glacier nourishment, snow melt floods	Nivation	

Note: Landforms in italics

though the effects are much more variable and difficult to investigate. While they often occur in association with other processes typical of mountainous areas such as rockfall and DEBRIS FLOWs, in some areas, for example the Himalayas, snow avalanches may be the dominant process in specific elevation zones (Hewitt 1989).

The geomorphic effects of snow avalanches may be either direct or indirect (Table 4). The former may involve both erosional and depositional processes and forms and in the case of some landforms, elements of both. The latter relate to aspects of the environment that influence other geomorphic processes.

The presence of downslope aligned alternating ribs and furrows in rock slopes affected by snow avalanches has long been known (Allix 1924). Abrasion of bedrock by rocks in avalanches is thought to play a role in chute formation but it is probably secondary to the transport of material loosened by other processes (Luckman 1977).

Where snow avalanches occur in locations with significant concave breaks of slope (often where formerly glaciated valleys have been subsequently filled with alluvium), they may generate great impact pressures on the landscape resulting in features collectively referred to as avalanche impact landforms (Luckman *et al.* 1994). The erosional parts of these may form circular or elongated pits but more commonly the pits intersect the local water table or the landforms may occur in water bodies such as lakes or rivers to form pools. They have been described for many areas of the world but seem particularly characteristic in areas of North America, Norway and New Zealand where resistant bedrock has resulted in the preservation of steep-sided formerly glaciated valleys.

Mounds and ramparts are made up of material scooped out by avalanche impacts and often form arcuate ridges at the distal edge of pits or pools. However, there may also be some contribution to their formation by accumulation of debris from frequent small 'dirty' avalanches once the pit or pool is full of previously avalanched snow.

In mountainous areas above the treeline and the winter snowline, snow avalanches redistribute material on debris slopes in association with other processes such as DEBRIS FLOWs and rockfall. Where avalanche transport of debris onto slopes of about 20° to 30° dominates, landforms referred to as avalanche boulder tongues occur. In the pioneering study of these features, Rapp (1959) identified two types – road-bank and fan. Road-banks are smooth flat-topped accumulations of debris often with an asymmetrical profile while fans are fan-shaped tongues of debris extending on relatively low angles towards valley floors. Rapp suggested that fan tongues result from larger avalanches and that the asymmetry of road-bank tongues resulted from preferential deposition of wind-drifted snow on the down valley side of the deposit. However, subsequent studies have suggested that other factors may lead to differences in form with road-bank tongues tending to be favoured where there is a plentiful debris supply and a confined track (Luckman 1977). Boulder tongues are characteristically 100 to 1,000 m long, up to 200 m wide and 10 to 30 m thick. They generally have a strongly concave long profile by which they can be distinguished from debris slopes formed by other processes.

Debris tails are small-scale forms which often occur on boulder tongues where there is a large range in debris sizes. They take the form of streamlined deposits of small to medium-sized particles usually downslope and more rarely upslope of large boulders. As indicated in Table 4, both erosion and deposition may play a role in their formation.

The most significant of the indirect effects of snow avalanches in relation to geomorphology is their influence on the characteristics of sediments and soils, as these may often be used to infer which processes have been responsible for building debris slopes under present or past conditions. Blikra and Nemec (1998) showed that while snow avalanches may transfer surface debris in a similar manner to debris flows, there are significant differences in the resulting sedimentary deposits, including the existence of precariously perched melt-out debris. In addition, snow avalanche deposits are often patchy, ranging from lobes of unsegregated debris to areas with better sorted sands or granules arising from water deposition following snow flows. Snow avalanche deposits may be distinguished from those of rockfalls which are characterized by openwork structure, weaker fabric strength (see FABRIC ANALYSIS) and existence of pronounced downslope increase in sediment size, as shown by Blikra and Nemec (1998) in Norway and Jomelli and Francou (2000) in the French Alps.

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IAN OWENS

AVULSION

Shift of a part or the whole of a river channel to another location on the FLOODPLAIN. It seems to be caused by local superelevation of part of the channel (see CHANNEL, ALLUVIAL) or channel system above the floodplain, as a result of river AGGRADATION, creating a local gradient advantage. Most avulsions occur when a triggering event forces a river across the stability threshold (see THRESHOLD, GEOMORPHIC). The closer the river is to the threshold, the smaller the trigger needed to initiate an avulsion (Jones and Schumm 1999). Local short-term processes triggering avulsions include: tectonic movements, variations in discharge and SEDIMENT LOAD AND YIELD, mass failure, aeolian processes (e.g. the formation of dunes) and log or ice jams. Regional long-term factors controlling avulsion include: BASE LEVEL change, climatic change, tectonic movements and discharge variation (Stouthamer and Berendsen 2000).

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SEE ALSO: anabranching and anastomosing river; palaeochannel; river delta

ESTHER STOUTHAMER

B

BADLAND

Badlands are deeply dissected erosional landscapes (Plate 8), formed in softrock terrain, commonly but not exclusively in semi-arid regions. Badland processes are dominated by overland-flow erosion. Badlands usually have a high DRAINAGE DENSITY of rill and gully systems, and at most support sparse vegetation. Badlands may comprise zones of coalesced hillslope gullies (see GULLY) within which little of the pre-gullying terrain remains. Badlands are common in areas with at least seasonal drought, in semi-arid and arid areas, Mediterranean and dry-season tropical areas. However, they also occur in humid regions, for example on eroding coastal and river cliffs. Badlands may result from natural processes, but their extent may be accentuated by human activity. Some badlands may be the result of human-induced soil erosion. Two prerequisites for badland development are erodible rock, typically marl, clay, or shale, and available relief. Badlands are common in uplifted and dissected softrock terrain (Plate 8).

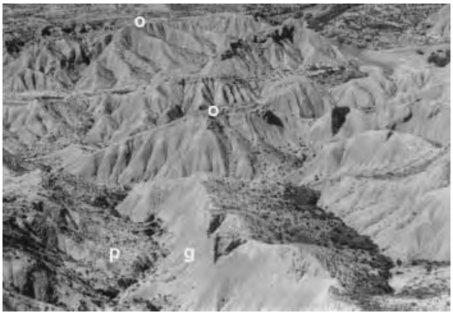

Plate 8 Extensive badland development, Tabernas, south-east Spain. These badlands are cut into older pediment surfaces (o), and owe their origin to tectonically induced dissection of an uplifted Neogene sedimentary basin, under semi-arid climatic conditions. The badland slopes are dominated by surface processes, but note the strong aspect-related contrasts. Note also the differences between micropediment-based badland slopes (p) and gully-based badland slopes (g)

Badland processes and morphology

Processes on badland slopes are dominated by surface erosion by Hortonian (see HORTON'S LAWS) OVERLAND FLOW (Horton 1945), created by rainfall intensity exceeding infiltration capacity. Away from the drainage divide, runoff increases in depth, at first as non-erosive laminar flow, then as turbulent sheet flow. Then, when shear stresses exceed the resistance of the surface, erosion is possible. Initially erosion is by surface winnowing as sheet erosion (see SHEET EROSION, SHEET FLOW, SHEET WASH) or linearly as RILL erosion. As runoff increases further downslope, shear stresses exceed the strength of the underlying material, and channel incision is possible. At that point, defined by a minimum drainage area (Schumm's (1956a) constant of channel maintenance), sheet flow and rill flow give way to open channel flow, and sheet and rill erosion to gully erosion and stream channel processes. The requirements for Hortonian processes are intense storm rainfall, little vegetation cover, low infiltration capacity, easily erodible material and relatively steep slopes.

On slopes dominated by Hortonian processes, smooth rounded divides (Horton's (1945) belt of no erosion) give way downslope to straight rilled slopes, which in turn feed v-shaped gullies at the slope base. Rill and gully networks generally accord with the 'laws' of drainage composition (Schumm 1956a; Strahler 1957) (see HORTON'S LAWS). Local variations in the drainage density of the rill networks or other microtopographic features, such as erosion pinnacles (see HOODOO), reflect local lithological variations in infiltration capacity or erosional resistance.

On many badlands, other processes also operate. Repeated WETTING AND DRYING WEATHERING, and in some areas freezing and thawing, weather the surface materials. These processes have two main effects. Desiccation cracking or frost heave may greatly increase infiltration capacity, so that rainfall-excess overland flow becomes unlikely. In that case surface water infiltrates, and reaches rill systems by lateral flow through the weathered surface layers, reducing interill, but not rill erosion. These processes may be exacerbated by the geochemistry of the material, particularly the presence of exchangeable sodium salts. On wetting, such materials may be prone to slaking, greatly enhancing the potential for mudslides (Benito *et al.* 1993).

Weathering processes, especially on materials rich in swelling clays (see EXPANSIVE SOIL), create their own microtopography of pinnacles (Finlayson *et al.* 1987), crack patterns and so-called 'popcorn textures' (Hodges and Bryan 1982). At a larger scale, the relation between weathering and removal rates (i.e. WEATHERING-LIMITED AND TRANSPORT-LIMITED conditions) may be important. Under transport-limited conditions a weathered mantle may accumulate, ultimately to fail as a shallow mass movement, whereas on an equivalent weathering-limited slope Hortonian processes may dominate. Aspect (see ASPECT AND GEOMORPHOLOGY) often controls these processes (Plate 8), either directly or through its influence on vegetation.

A major process in some badland areas is subsurface erosion by piping (Bryan and Yair 1982; Harvey 1982) (see PIPE AND PIPING; TUNNEL EROSION). Pipes may be induced mechanically by the channelling of overland flow below the surface along animal burrows, vegetation rootways and, particularly in dissected terrain, down tension cracks. Piping is enhanced by the geochemical properties mentioned above (Gutierrez *et al.* 1988; Faulkner *et al.* 2000).

On piped badlands surfaces may have lower rill network densities, and pipe inlets and outlets may add to the morphological diversity. There may be modifications to channel alignments, when the major gullies result from pipe collapse rather than from Hortonian processes.

On some badlands, processes are dominated by single (Hortonian) processes, but on many badlands interactions between several processes take place (Schumm 1956b; Faulkner 1987; Harvey and Calvo-Cases 1991). Spatially, process interactions include on-slope interactions between weathering, Hortonian runoff, mass movements and piping, and also include HILLSLOPE-CHANNEL COUPLING relationships involving interactions between the slope processes and basal stream or gully activity (Harvey 2002). This may involve the build-up of material derived from slope erosion, and its periodic flushing by stream processes, maintaining erosional activity at the slope base (Harvey 1992).

Temporal characteristics of process interactions result from discrepancies between effectiveness, rates and frequency of the various processes. They may relate to individual storm events and recovery periods. However, a common timescale is one of seasonal cyclicity, often related to a seasonal process regime, generating for example seasonal rill development cycles (Schumm 1956a; Harvey 1992). Another type of seasonality or longer-term cyclicity may relate to flushing, when there are different frequencies of sediment generation and removal rates (Harvey 1992; Faulkner 1994). Cyclicity may also be due to discrepancies between weathering and removal rates (Harvey and Calvo-Cases 1991). Over an even longer term there may be progressive changes related to longer-term morphological development (Harvey 1992).

In a wider context, badlands show relationships to GULLY systems. Gully systems may develop as valley-floor (ARROYOs) or hillslope gullies (Campbell 1997). Badlands result from the coalescence of both basally- and midslope-induced hillslope gullies. Of fundamental importance for badland morphology and development is the local base level. An incising or laterally migrating gully channel maintains an active base level, influencing all badland processes, surface processes, slope stability, sediment removal, and even subsurface processes through its influence on hydraulic gradients. Basally-induced gullying and badland development are more likely to have effective base-level control, but even there slope retreat may transform gully/channel-based badlands by micropediment-based badlands (Plate 8).

Badland dynamics

Badlands have two major roles within the context of the wider geomorphic system: (1) as major sediment sources to the fluvial system, and (2) as a major influence on slope evolution. Badlands, especially when coupled with the stream network, represent a zone of drainage network expansion, an increase in drainage density and an increase in stream order (Strahler 1957). This has hydrological implications, but above all increases the sediment supply to the fluvial system to the extent that a zone of badlands may dominate sediment dynamics of a drainage basin (Campbell 1997). Badlands may act as a major influence on slope evolution, especially in semi-arid areas, producing extensive pediment areas at the base of retreating escarpments.

Badlands are erosional not equilibrium forms. In addition to the results of process interactions, badland morphology progressively changes as the badlands develop. This, in turn, modifies the processes. Ultimately badland development depends on the relative rates of extension and stabilization. Harvey (1992) has demonstrated that in a humid environment, once a gully system is decoupled from a basal stream, stabilization by vegetation colonization operates faster than gully extension. Those gullies do not develop into badlands. However, in many semi-arid areas, although auto-stabilization mechanisms have been recognized (Alexander *et al.* 1994), stabilization processes are slower than gully extension so that badlands develop and persist. Under conditions of incising base levels, basal v-shaped gully systems would maintain characteristic badland processes and morphology on the slopes. Under stable base-level conditions the badland slopes progressively retreat, forming pediment-based badlands, which if they do not self-stabilize, would ultimately produce a landscape of extensive pediments and small badland residuals.

One factor of fundamental importance is the interaction between vegetation and geomorphic processes which affects both the generation of overland flow and the stabilization of eroded slopes (Alexander *et al.* 1994; Gallert *et al.* 2002). However, of the main factors influencing badland geomorphology, it is the interaction of base level with the surface processes that has the greatest influence on badland evolution.

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ADRIAN HARVEY

BAJADA

The broad zone of coalesced or compound ALLUVIAL FANs that form a more or less continuous piedmont alluvial apron lying between the mountain front and the basin floor in areas like the semi-arid south western United States. They are in contrast to rock-cut PEDIMENTs. The term was introduced by Tolman (1909).

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A.S. GOUDIE

BANK EROSION

Bank erosion is the detachment and entrainment of bank material as grains, aggregates or blocks by fluvial, subaerial or geotechnical processes. Riverbank erosion processes are important for the evolution of MEANDERING and BRAIDED RIVER systems and FLOODPLAINs, catchment sediment output and biodiversity on floodplains. Bank erosion can also lead to loss of agricultural land and riparian structures, exacerbated sedimentation problems and riverine boundary disputes, sometimes necessitating bank stabilization works.

Bank erosion measurement

The many methods of bank retreat measurement can be classified into long-timescale, medium-timescale, and short-timescale techniques (Lawler 1993a). Long-timescale methods employ sedimentological, botanical or cartographic evidence to reveal channel change over decades or thousands of years. For example, sequences of channel movements can be preserved in datable fluvial deposits, or quantified through dendrochronological analysis (see DENDROCHRONOLOGY) of trees colonizing bar (see BAR, RIVER) surfaces (e.g. Hickin and Nanson 1984). River course changes can be quantified by superimposing early maps, aerial photographs and satellite imagery (e.g. Hooke and Redmond 1989; Lewin 1987), often using analytical photogrammetry and GIS (e.g. Lane *et al.* 1993). Medium-timescale techniques include the periodic field resurvey of bank lines with theodolites, EDMs (Electronic Distance Measurers), Total Stations or GPS (Global Positioning Systems) (Lawler 1993a). Cross-section resurvey, however, using levelling or Total Station techniques, can be more sensitive to subtle changes. Airborne laser altimetry and side-scan sonar methods have also been applied.

The following short-timescale techniques are more useful for process studies, because the geomorphological change can be related to forcing hydrological and meteorological events. Erosion pins can be inserted into banks: erosion then exposes more pin, the length of which is recorded periodically (Lawler 1993a). Terrestrial photogrammetric monitoring involves the repeated capture of ground photograms using stereometric cameras, from which the three-dimensional bank form (DIGITAL ELEVATION MODEL (DEM)) is derived. 'Subtracting' successive DEMs reveals the intervening bank erosion rate (Lawler 1993a; Lane *et al.* 1993). However, all the methods above reveal little about the *timing* of bank erosion *events*, knowledge which is crucial to process inference. Lawler (1992), therefore, developed the *automatic* Photo-Electronic Erosion Pin (PEEP) system. When erosion occurs, the PEEP signal increases; if deposition occurs, voltages are decreased. The system thus allows the magnitude, timing and frequency of erosional and depositional activity to be monitored more precisely, including for TIDAL CREEK (Lawler *et al.* 2001), and is now used by twenty research groups worldwide. The example in Figure 9 shows how the PEEP approach, for the first time, fixes the time of an erosion event to forty-three hours *after* the hydrograph peak; this suggests the operation of mass failure processes rather than fluid entrainment.

Bank erosion rates

Bank erosion rates range from 0–1,000 m a^{-1} and tend to increase with boundary shear stress, STREAM POWER, FREEZE–THAW CYCLE activity and for silty or

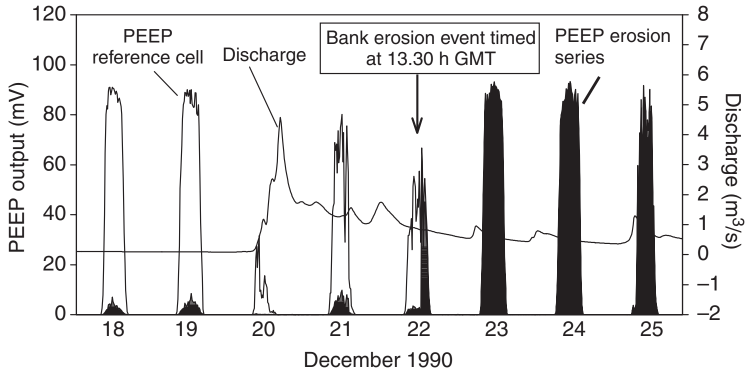

Figure 9 Timing of a bank erosion event at 13.30 h GMT on 22 December 1990 detected by the automatic Photo-Electronic Erosion Pin (PEEP) monitoring system on the Upper River Severn, UK. The bank erosion event lags the flow peak by forty-three hours, suggesting mass failure processes are responsible (from Lawler *et al.* 1997)

saturated bank materials of low cohesion (Lawler *et al.* 1997). Within some basins bank erosion rates peak in the *middle* reaches (e.g. Lewin 1987; Lawler *et al.* 1999), possibly related to stream power increases (Lawler 1992; Abernethy and Rutherfurd, 1998; Knighton, 1999).

Riverbank erosion processes

The many bank erosion mechanisms identified can be grouped into fluid entrainment, preparation or mass failure processes (Lawler 1992; Lawler *et al.* 1997; Prosser *et al.* 2000).

FLUID ENTRAINMENT PROCESSES

Entrainment occurs when the motivating forces due to the flow (lift and drag, often indexed by boundary shear stress) and particle mass exceed the friction and cohesion forces holding the particle in place (Lawler *et al.* 1997). On non-cohesive (e.g. sandy) banks, material is usually entrained grain by grain, while on cohesive, fine-grained banks, material is eroded as aggregates or crumbs bound by cohesive forces. Cohesion results from a combination of physico-chemical, inter-granular, bonding forces, driven by the mineralogy, dispersivity, moisture content and particle size distribution of the bank material, and the temperature, pH and electrical conductivities of the pore fluid and river water (Osman and Thorne 1988). Cohesive banks are normally much more resistant to fluid entrainment than non-cohesive ones.

PREPARATION PROCESSES

The ERODIBILITY of cohesive bank materials, however, can change because of preparation or weakening processes. Crucially, then, the critical shear stress value for entrainment varies with antecedent material preparation (Lawler 1992; Prosser *et al.* 2000). For example, desiccated banks may crack as moisture is thermally driven off and clay minerals shrink (Plate 9). For instance, in summer, east-facing banks of the river Arrow, Warwickshire, UK reach early-morning warming rates of 7 °C h^{-1}, peak daily temperatures above 30 °C and diurnal temperature ranges of 20 °C (Lawler 1992). Flowing waters then exploit cracks to enhance erosion (e.g. Lawler 1992). Freeze–thaw activity takes many forms (e.g. Lawler 1993b; Prosser *et al.* 2000). For example, NEEDLE-ICE can lift or incorporate material, and transport it downslope on ablation. Much disturbed sediment remains, though, to be readily removed by subsequent flow rises (Lawler 1993b).

MASS FAILURE PROCESSES

Mass failure occurs when blocks of material collapse or slide towards the bank toe (Plate 9). Banks are vulnerable to mass failure if steep, high, fine-grained, of high bulk unit weight and subject to high or fluctuating PORE-WATER PRESSURES – indeed any variable which increases the mass of material above a potential failure

Plate 9 Bank erosion on the lowland river Arrow, Warwickshire, UK. A failed block of bank material (length ~2 m) lies under water at the bank toe, with the scar of the failure surface visible above. Desiccation cracking is above the scar. Flow is from right to left

surface. Hence, bed scour can induce bank failure by increasing bank height and angle. Also, failure can follow increases in block mass due to moisture uptake, often after submergence. Bank failures should thus occur on hydrograph recession limbs, following saturation. This is confirmed by the PEEP automatic erosion monitoring system (Figure 9). Figure 10 shows bank failure characteristics. For the most common type, the failure surface is almost planar (Plate 9 and Figure 10). Cantilever failure occurs on composite riverbanks if, because of faster erosion of the lower more erodible layers, an overhang develops then collapses (Lawler *et al.* 1997) (Figure 10g and h).

Mass failures can be analysed using geotechnical slope stability theory (e.g. Osman and Thorne 1988; Darby and Thorne 1996; the CONCEPTS model of the United States Department of Agriculture (USDA)). One example is the Culmann formula for planar failure (Figure 10c), which predicts the critical height for a bank, H_{crit}:

$$H_{crit} = \frac{4c}{\gamma} \cdot \frac{\sin\alpha \cdot \cos\phi}{\{1 - \cos(\alpha - \phi)\}}$$

where c = material cohesion (kPa), γ = *in situ* unit weight of material ($kN\,m^{-3}$), α = slope angle (°), ϕ = friction angle (°). Many of the data required can be collected using the Stream Reconnaissance Record Sheets of Thorne *et al.* (1996).

Bank processes may change in a longitudinal direction (Lawler 1992). This idea, developed into the DOCPROBE model (DOwnstream Change in the PROcesses of Bank Erosion), suggests that, in upstream reaches, preparation processes are most effective, because stream power and bank heights are too low for significant fluid entrainment and mass failure respectively. In middle reaches, where stream power is high, fluid entrainment dominates. Further downstream, bank heights and material properties exceed critical values and mass failure processes prevail. Evidence has now emerged to support this model (e.g. Lawler 1992; Knighton 1999; Abernethy and Rutherfurd 1998). Vegetation can considerably reduce fluid erosion, partly through root reinforcement or foliage protection (Thorne 1990; Abernethy and Rutherfurd 1998; Simon and Collinson 2002). However, forest canopy shade may suppress shorter riparian vegetation, increasing erosion.

A much richer mix of bank erosion processes is now recognized. Though flow processes are important, research has shifted to temporal change in bank erodibility, vegetation, riparian hydrology and the dynamics of bank erosion events, often using novel automated monitoring techniques.

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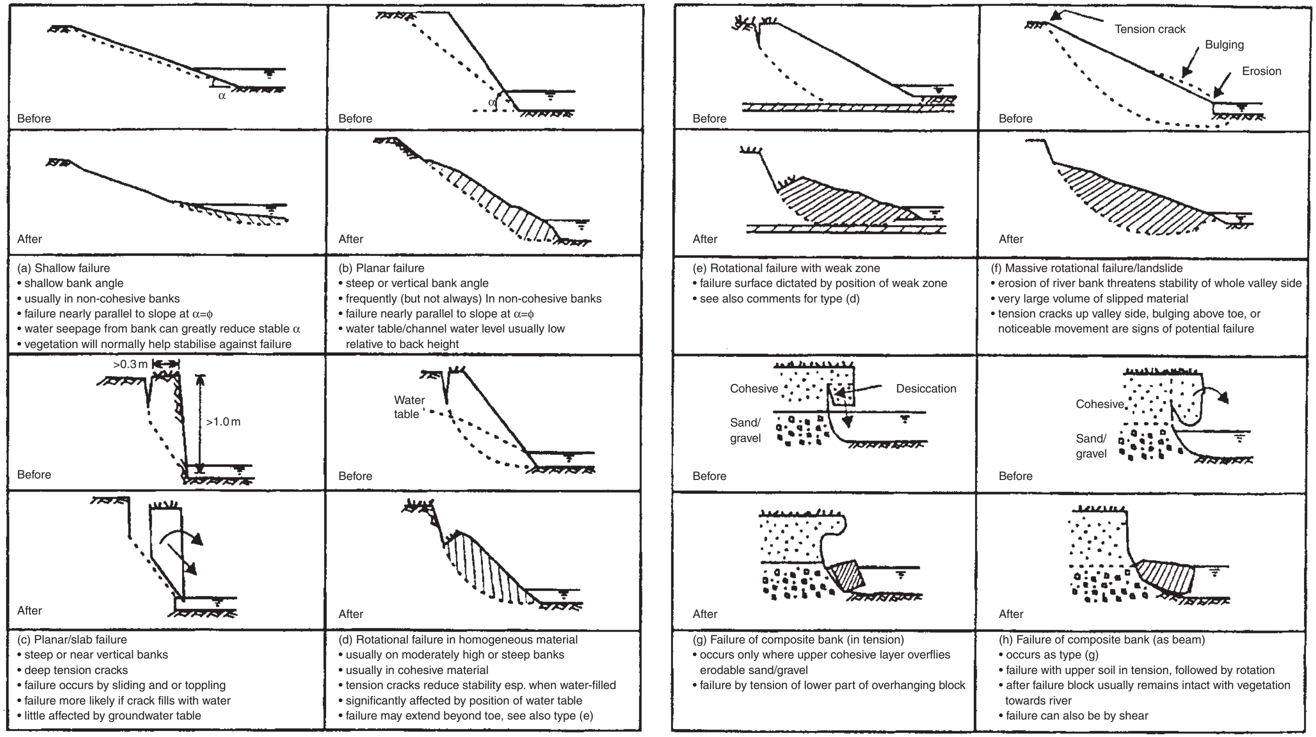

Figure 10 Characteristics of bank failure (from Hey *et al.* 1991; cited in Lawler *et al.* 1997)

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SEE ALSO: fluvial geomorphology; hydraulic geometry

DAMIAN LAWLER

BANKFULL DISCHARGE

Bankfull discharge, a hydrologic term, is the flow rate when the stage (height) of a stream is coincident with the uppermost level of the banks – the water level at channel capacity, or bankfull stage. Bankfull stage is a fluvial-geomorphic term (see FLUVIAL GEOMORPHOLOGY) requiring an interpretation of site-specific landforms. In this context, bank typically refers to a sloping margin of a natural, stream-formed, alluvial channel (see CHANNEL, ALLUVIAL) that confines discharge during non-flood flow. Although the term bankfull stage can refer to various channel-bank levels, it generally applies to alluvial-stream channels (1) having sizes and shapes adjusted to recent fluxes of water and sediment, (2) that are principal conduits for discharges moving through a length of alluvial bottomland, and (3) that are bounded by FLOODPLAINS upon which water and sediment spill when the flow rate exceeds that of bankfull discharge. Thus, the concept of bankfull discharge, which often approximates the mean annual flood for perennial streams, includes the floodplain as a unique, identifiable geomorphic surface, all higher surfaces of alluvial bottomlands being terraces (see TERRACE, RIVER) (generally former floodplain surfaces), and acknowledgement that bankfull discharge occurs only when stream stage is at floodplain level.

Previous studies

Numerous discussions, only a few of which are cited here, have addressed the bankfull concept. A review by Williams (1978) documented a variety of published definitions for bankfull discharge and listed a wide range of flood frequencies related to the bankfull stage; almost without doubt the range resulted from observer misidentification of surfaces higher and lower than the floodplain as that of bankfull. Radecki-Pawlik (2002), among many others, showed that field determinations of bankfull stage and therefore floodplain level are interpretive, and thus the related bankfull discharge may differ greatly from that of the mean annual flood.

Papers by Woodyer (1968) and Osterkamp and Hupp (1984) recognized a variety of bottomland surfaces, including the floodplain, and related an approximate flow characteristic to each. Petit and Pauquet (1997) and Castro and Jackson (2001), respectively, determined return intervals for bankfull discharge of generally 1.2 to 3.3 years at sites of gravel-bed streams in France, and return intervals of 1.0 to 3.11 years for bankfull discharge at seventy-five stream sites of the north-western United States.

Significance

Bankfull discharge is significant owing to the hydraulic and related physical changes that occur for most alluvial stream channels when flow increases from in-channel to overbank conditions. Resistance to flow 'decreases with increasing water depth to reach a minimum at bankfull stage, so that the channel operates most efficiently with regard to water conveyance when the flow is at bankfull level' (Petts and Foster 1985: 150). Thus, the change in hydraulics as flow depth increases exerts a basic control on the geomorphic processes that are related to floodplain formation, regardless of the return period that may be associated empirically with the discharge at bankfull stage. The approximate height above the channel bed at which overbank flow begins is the level to which riverine bars develop (see BAR, RIVER), a process that can be described in terms of flow field, channel bathymetry and characteristics of sediment size and transport (Nelson and Smith 1989). Significant also is the observation that numerous data collected from perennial streams suggest that floodplain level is roughly equivalent to the stage of the mean annual flood, about 2.3 years (Wolman and Leopold 1957).

Floodplain formation

Floodplains typically form through POINT BAR deposition and, generally to a lesser extent, by deposition of sediment during overbank flows. Overbank deposits underlying floodplain and alluvial-terrace surfaces are typically poorly sorted and generally exhibit thinly bedded and alternating layers of silt, sand and possibly gravel. When a succession of floods causes overbank deposition, each flood elevating the surface higher above the channel, the deposits tend to grade from relatively coarse particles at the bottom upward to finer sizes. Because the thickness of overbank sediment deposited by large floods is generally small, averaging about 20 mm (Wolman and Leopold 1957), numerous episodes of overbank deposition are ordinarily needed to result in the accumulation of sediment on a gravel bar to a flood-plain level. AGGRADATION above flood-plain level is minimal owing to the infrequency of overtopping discharges, scour of flood-plain sediment by large floods and EROSION of accumulated deposits by lateral channel migration (Wolman and Leopold 1957). Nanson (1986), among others, suggested that processes resulting in the development of floodplains are influenced by prevailing conditions of energy (largely channel gradient) and the availability of sediment for entrainment. It follows, therefore, that many high-gradient streams have little potential for lateral ACCRETION (point-bar formation) and that flood-plain development occurs principally by vertical accretion.

Considerations and problems

The legitimate application of bankfull stage, bankfull discharge and floodplain to hydrologic and geomorphic studies requires accurate field determination of bankfull (flood-plain) level, which may be difficult if streamflow data are unavailable. Bankfull level is recognized easily along channels with point-bar deposits, especially if recent overbank deposits overlie point-bar sediment. For channels lacking point-bar features, interpretation of bankfull stage must rely on observations of channel morphology and gradient, bed and bank sediment, vegetation, root exposure and indications of flood processes.

The bankfull concept has been valuable as a means of describing clearly and effectively the processes and landforms of perennial-stream channels in humid areas. In recent decades, however, the bankfull concept has been so prevalent that it often has been overextended by Earth scientists who have confused the floodplain with other prominent alluvial landforms and corresponding flow frequencies. Thus, 'bankfull' should be used cautiously and consistently within the constraints by which it was described.

Common difficulties of the bankfull concept include its misapplication to non-alluvial conditions, such as streams incising till or debris-flow deposits, where fluvial adjustment is incomplete and bank height may be poorly related to flood frequency. Alluvial surfaces adjacent to high-energy (especially alpine and subalpine) streams, which typically correspond to the approximate stage of mean discharge (Osterkamp and Hupp 1984; Hupp 1986), commonly are misidentified as floodplain. As noted previously, the approximate correlation of bankfull stage (flood-plain level) and discharge with the mean annual flood pertains principally to perennial streams of moist areas; in drier regions with intermittent to highly ephemeral streams, the floodplain may correspond to floods with return periods of 100 years or more.

Flows smaller than bankfull discharge, those that occur more frequently than that of bankfull stage, typically cause in-channel features unrelated to bankfull discharge (such as BEDFORMS and bars). Processes and channel features resulting from these common events should not be confused with processes that correspond to bankfull discharge, and it should be recognized that all flows transport and sort sediment, thereby modifying the stream-channel morphology. Inappropriate emphasis on the bankfull concept has given rise to terms such as 'dominant discharge' and 'channel-forming discharge'. Such terminology, which focuses on a single flow rate, fails to recognize that all flows contribute to channel shape. As demonstrated by Wolman and Miller (1960), bankfull discharge, if related to geomorphic work accomplished, may indeed be dominant when applied to well-adjusted channels of perennial streams, but the dominance of a bankfull discharge becomes increasingly questionable as rates of precipitation and runoff decrease. Because all flows alter the shape of alluvial-stream channels, the term 'channel-forming discharge' may be inappropriate.

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W.R. OSTERKAMP

BAR, COASTAL

Coastal bars can broadly be defined as aggradational ridges of sediments whose formation, morphology and behaviour are determined by interactions between WAVES, CURRENTS, tides, local slope and grain size. Some confusion exists regarding usage of the terms *bar* and *ridge*, however, since features such as BEACH RIDGES and CHENIER RIDGES are not normally considered bars. Furthermore, bars are found in BEACH, RIVER DELTA, ESTUARY and CONTINENTAL SHELF environments with a wide range of sizes, types and orientation. However, comparatively more COASTAL GEOMORPHOLOGY research studies have focused on bars which exist in the nearshore zone of sandy wave-dominated beaches.

Early studies (e.g. Shepard 1950) identified a seasonal cycle of beach morphology with winter storms promoting offshore sediment transport and bar formation and calmer conditions in summer favouring landward migration of the bar and eventual welding to the beach face. However, the existence of such 'winter' and 'summer' profiles is not universal as both barred and non-barred profiles occur at all times in some areas, while in others only one type may persist throughout the year. Furthermore, cyclic beach response at timescales much shorter than seasons can result in barred/non-barred profiles (Short 1979).

Cross-shore barred profiles are most commonly asymmetrical, having a distinct crest and a steeper landward slope than seaward slope (Figure 11a). Types of bars are often distinguished based on their alongshore planform shape and orientation relative to the shoreline. They may be linear (also referred to as longshore or shore-parallel; see Figure 11b), sinuous or crescentic (often termed

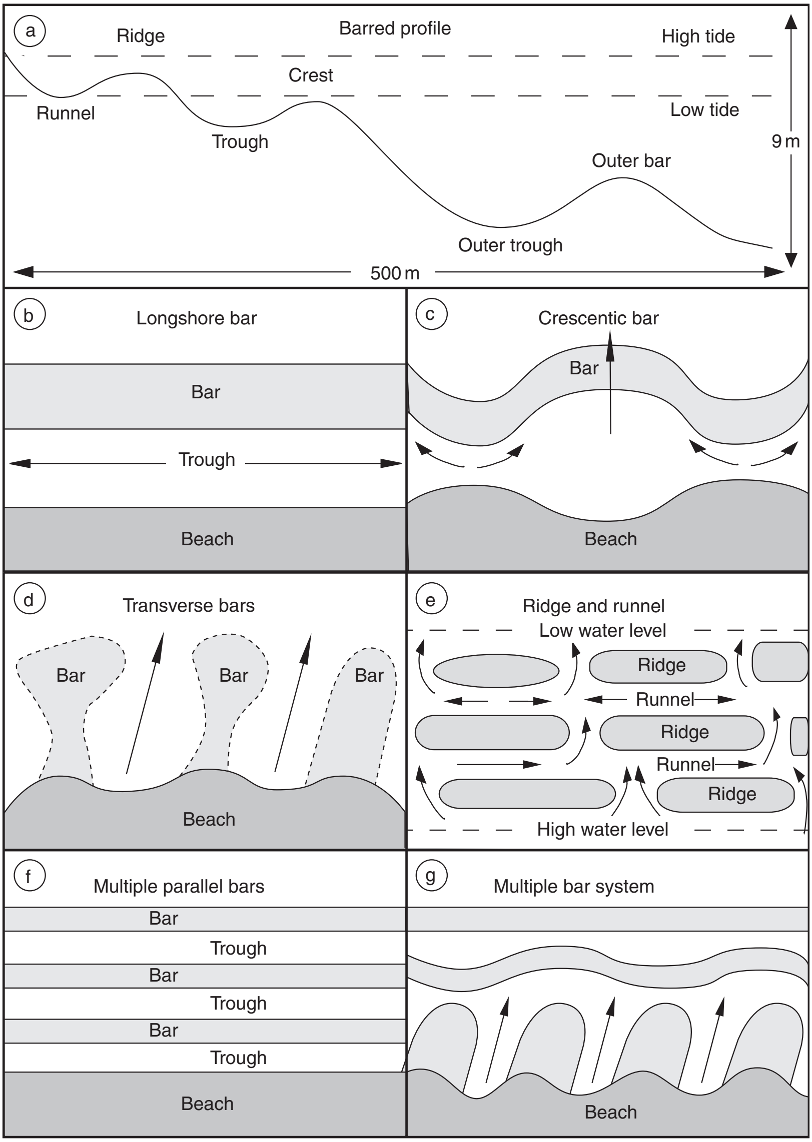

Figure 11 Idealized cross-shore barred profile (a) and some examples of bar types (b–g)

rhythmic) with a trough separating them from the shoreline (Figure 11c), or they may consist of alternating transverse bars, which are welded to the shoreline and are separated by channels occupied by RIP CURRENTs (Figure 11d). Bar type appears to be strongly related to wave energy level with linear bars developing under high-energy conditions, crescentic bars during intermediate energy, and transverse bars during lower wave energy levels. Under very low-energy conditions, a bar may become fully welded to the beach and appear as a flat terrace at low tide. These types of bar configurations are common on microtidal beaches and may grade into each other as energy levels vary. A number of classifications exist describing both bar types and the continuum of bar evolution (e.g. Greenwood and Davidson-Arnott 1979; Short and Aagaard 1993; Wijnberg and Kroon 2002).

Many beaches are characterized by the existence of multiple offshore bars (Figure 11f, g), ranging in number from two to over a dozen in some cases. Although high-energy swell wave environments can exhibit two or three bars, multi-barred profiles are most commonly developed in storm wave dominated sea and lacustrine environments where the outer bars are only active during short intense storms, remaining stationary during longer periods of low-wave energy. The number of bars seems to be related to the overall nearshore slope with lower gradients characterized by more bars. Both the spacing and size of bars has been observed to increase offshore. Bars are commonly absent on steep beaches.

Sandy beaches characterized by significant tidal ranges and low-energy conditions typically have gentle gradients and although some of the previously described bar types may be present, the presence of RIDGE AND RUNNEL TOPOGRAPHY (Figure 11e) in the intertidal zone is more common (Masselink and Anthony 2001). These exist as a series of low amplitude bars (ridges), which are usually stable in form and position, separated by subdued channels (runnels) associated with tidal drainage, and should not be confused with the multi-barred profiles described above.

As reviewed by Komar (1998) and Aagaard and Masselink (1999), some uncertainty remains regarding the formation of nearshore bars despite considerable theoretical, laboratory and field research. Bars develop as a result of sediment convergence and most mechanisms for their formation attempt to explain this. An early theory, that vortices under plunging breakers move sediment seaward forming a bar just seaward of the breakpoint, has largely been discounted since the breakpoint location on natural beaches with irregular waves varies considerably. It is more likely that sediment convergence results from onshore sediment transport outside the surf zone due to wave asymmetry and offshore transport in the surf zone due to bed return flow, with the bar forming somewhere near the breakpoint. Single and multiple bar formation has also been attributed to net sediment transport patterns associated with standing infragravity waves. According to this theory, if the sediment transport is predominantly bedload the bars will form at nodal positions, whereas if suspension dominates, they will form at antinodal positions (Bowen 1980). This theory has also proven useful in explaining rhythmic bar morphology (Holman and Bowen 1982) although nearshore cell circulation and rip currents are also important factors.

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SEE ALSO: beach; beach sediment transport; current; ridge and runnel topography; wave

ROBERT W. BRANDER

BAR, RIVER

The nature and distribution of alluvial instream geomorphic units is fashioned by the interaction between unit stream power along a river reach and sediment calibre and availability. If a reach has excess energy relative to available sediment of sufficient size, flushing is likely to occur.

Alternatively, with excess sediment availability or insufficient flow energy, continuous instream sedimentation is likely to occur, commonly in the form of near-homogenous sheets. In most cases, rivers fall somewhere between these extremes, with transient sediment stores of differing calibre and bed material organization in differing landforms along the channel bed.

The most common instream geomorphic units are accumulations of deposits referred to as bars. These areas of net sedimentation of comparable size to the channels in which they occur are key indicators of within-channel processes. Interpretation of bar type is often critical in elucidation of river character and behaviour. There are two main components in bar form. The basal feature, or platform, is made up of coarse material and is overlain by supraplatform deposits of varying forms which is subject to removal and replacement during floods. Bars are readily reworked as channels shift position over the valley floor. Bank-attached features are much less likely to be reworked than mid-channel forms. The long-term preservation of bars is conditioned by factors such as the aggradational regime and the manner of channel movement.

Bars adopt many varied morphologies, ranging from simple unit bars (Smith 1970) to complex compound features (Brierley 1991, 1996). Bar character is controlled primarily by local-scale flow and grain size characteristics. Unit bars are simple features composed of one depositional style. The sediments of a unit bar (whether they be sand or gravel in composition) tend to fine in a downstream direction. As unit bars are found at characteristic locations along long profiles under particular sets of flow energy (stream power) and bed material texture relationships, a 'typical' down-valley transition in forms can be discerned (Church and Jones 1982). Bed material character, and the competence of flow to transport it, determine formation of longitudinal bars as flow divides around a tear-drop shaped feature. When flow is oriented obliquely to the long axis of the bar, a diagonal feature is produced. This is commonly associated with a dissected riffle. In highly sediment-charged sandy conditions, flow divergence results in transverse or linguoid bars, which extend across rather than down the channel (Collinson 1970; Cant and Walker 1978). Alternatively, the entire channel bed may comprise a homogenous sandsheet.

Instances in which patterns of sedimentation are dominated by within-channel bars reflect situations in which the material on the channel bed is either too coarse to be transported or the volume of material is too great to be transported. These scenarios are generally associated with gravel and sand bed systems respectively, such that competence and capacity limits are exceeded and flow divides around sediment stored in the channel zone.

In contrast to various mid-channel sedimentation features, rivers that are more readily able to accommodate their sediment load or have lower available energy are commonly characterized by bank-attached bars. Dependent on channel/flow alignment, lateral and POINT BARS are found at channel margins under both sand and gravel conditions. These features record sediment accretion on the convex slopes of river bends. Lateral bars tend to occur along straighter river reaches, while point bars are formed on bends. Scroll bars on the inside of bends may form a distinct element in themselves, while former positions of the channel may be recorded by a series of accretionary ridges and intervening swales (Nanson 1980). A range of bar forms have also been characterized for laterally constrained sinuous channels, such as point dunes (Hickin 1969), gravel counterpoint bars (Smith 1987) and convex bar deposits (Goodwin and Steidtmann 1981).

Most river bars are not simple unit features, but are complex, compound features made up of a mosaic of depositional forms such as bar platforms, ridges, chute channels, etc. Compound bars can be differentiated into mid-channel and bank-attached forms. On mid-channel compound bars, chute channels may dissect the bar surface into a chaotic pattern of remnant units. Variants of within-channel compound bar features in sand-bed channels include linguoid bars (Collinson 1970), macroforms (Crowley 1983), sand flats (Cant and Walker 1978), sand waves (Coleman 1969) and sandsheets (Smith 1970). Vegetated mid-channel compound bars are referred to as islands. The array of smaller-scale geomorphic units that make up an island provides key insights into its formation and reworking. On bank-attached compound bars a range of erosional and depositional features such as chute channels, ridges and ramps can be formed under varying flow conditions. Chute channels short-circuit the main body of flow in a river. Enlargement of the chute channel and plugging of the old channel proceed gradually, resulting in a chute cut-off. Because of the small angular difference between the old channel and the chute channel, the stream continues to flow through the old channel for some time, depositing bedload sediment at the upstream and downstream ends and on the floor and sides until terminal closure of the cut-off is complete. Ramp units are depositional forms that result from deposition of coarse gravels within a partially-filled chute channel. These features have a steep upstream facing surface that effectively plugs the chute channel, disconnecting it from the downstream outlet. These chute channel fills are notably straighter in outline than either meander cut-offs or swales.

In quite different environmental settings, bedrock accretionary forms occur on low slopes. These bedrock core bars are characterized by bedrock ridges atop which alluvial materials have been deposited during the waning stages of floods. A positive feedback mechanism is induced when vegetation colonizes these surfaces inducing further deposition and the vertical building of a bar feature. These features are common along bedrock-anastomosing rivers (e.g. Van Niekerk *et al.* 1999).

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SEE ALSO: channel, alluvial; fluvial geomorphology, point bar

KIRSTIE FRYIRS AND GARY BRIERLEY

BARCHAN

Barchan is an active crescentic-shaped dune (see DUNE, AEOLIAN), developing in areas of unidirectional winds and limited sand supply. Most barchans are arranged in belts, in which they follow each other. Such belts contain dunes of different sizes: small dunes, which do not exceed several tens of metres in length or width and a few metres in height, develop in a short time; and meso-dunes, which rise up to 40 m and attain several hundred metres in length or width. In various deserts such as in the Western Desert of Egypt (Embabi 1982), Qatar, Peru and California, dimensions of simple barchans show strong linear allometric relationships, such as between length of windward side and dune height, and dune length and width of horns (Plate 10). Mega-barchans that attain heights up to 120 m and lengths of 2–4 km are less common than small and meso-barchans, and are recorded in areas such as the northern parts of Rub' al-Khali in Arabia, and Taklamakan. Sand supply, wind environment, atmospheric motion and age are the main controlling factors of barchan size.

Slope form is concave–convex on the windward side of simple barchans, with angles varying between 1° and 10°. As the barchan grows in size, concave segments occupy a higher percentage of the total length of the windward side. The form of the leeward side changes from

Plate 10 Barchans are crescentic dunes, the horns of which point downwind. These examples are in the Western Desert of Egypt in the Kharga Depression

convex–concave to convex–straight to mostly straight when it acquires the angle of repose.

The internal structure of barchans reflects the dynamics of sand removal and deposition on both dune sides. The dominant structure is composed of thin steeply dipping cross-strata resulting from grainflow and grainfall deposited on the slip face, and is preserved as the dune migrates downwind. A secondary horizontal to low dipping structure develops due to deposition on the top of the dune. The sets of cross-strata are separated by horizontal to steeply dipping bounding surfaces.

Barchan moves in the downwind direction due to the dynamics of sand removal and deposition on dune sides. Sand is removed from the lower part of the windward side, and is deposited on the dune crest or on the leeward side. Accumulation of sand on the dune crest leads to periodic sand avalanches on the surface of the slip face. By time, sand removed from the windward side is deposited on the leeward side/slip face, resulting in dune advancement in the downwind direction. Average annual net migration of barchans varies between a few metres to 100 m. Wind energy, dune size and surface relief are the most significant controlling factors in dune advancement. As barchans move forward, they encroach on highways, railways, fields and buildings, representing a permanent hazard to all sorts of human activities, unless checked.

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SEE ALSO: dune, aeolian

NABIL S. EMBABI

BARRIER AND BARRIER ISLAND

Coastal barriers and SPITS are often regarded as similar coastal forms in terms of beach deposition projecting across coastal re-entrants/bays. While barriers tend to bridge the re-entrant by joining the mainland at each end, spits are only attached at one end. This distinction is not rigid, as many barriers show cross-barrier breaks or breaches through which the sea may enter on a permanent or intermittent basis, thus forming barrier segments or barrier islands. The degree of segmentation required to allow the 'island' nomenclature is not defined. Studies on the US eastern seaboard (Johnson 1925; Hoyt 1967; Kraft 1971; Leatherman 1979) have concentrated on barrier formation and reworking, though studies on barrier stability and defending barriers in the face of rising sea levels have come to the fore, pressured by the extensive and expensive real estate located on them (Titus 1990).

Coastal barriers are complex constructional morphology involving deposition by waves, wave-generated currents, tidal currents and wind activity (Hayes 1979). By morphological definition a barrier must exhibit two morpho-dynamic environments/units: a seaward beach face and a landward facing back-barrier slope (Plate 11). These two units are exposed most clearly when the barrier is gravel-dominated (Orford *et al.* 1996). As sand becomes the dominant sediment, a third environment comprising aeolian dunes can appear at the top of the beach face (barrier crest) and spread onto the backslope. Current flows may have been responsible for the initial submarine platform under the barrier, but over time and with sea-level rise, wave action forcing barrier onshore-migration generates much of the later barrier's basal-stratigraphy in combination with fine sedimentation characteristic of the low-energy back-barrier bay. Tidal currents become influential once barrier islands appear.

Plate 11 Sand-dominated coastal barrier, Long Island, New York State. Photo by permission of Whittles Publishing, Caithness

Some barriers along the US coast are thought to have evolved out of spits (Fisher 1967). However, other barrier origin mechanisms have been proposed that involve onshore rather than longshore sediment supply. Offshore shoals and longshore bars have been proposed as accretional cores to barriers, but as bars are now recognized as being beach face dependent, it is unlikely that bars can appear before the barrier defines the beach face. Roy *et al.* (1994) suggested that some Australian barriers emerged through shoreface accretion when sea level achieved stationarity after major early-Holocene fluctuations and the shelf area was reworked with excess sediment moving onshore. This latter position sees barriers accreting throughout their length, their plan-view position being set by wave refraction of constructional swell waves. Regardless of origin per se, barriers should be seen as multi-phase morphology relating to changes in controlling variables. Barriers should also be seen in terms of a palimpsest imposed by the structure and roughness of the terrestrial platform that they are superimposed on.

Sediment type and supply are major controls on barrier development with a behavioural distinction to be drawn between sand-dominated barriers (SDB) and gravel-dominated barriers (GDB). This distinction has a spatial basis with GDB being more prevalent in mid-upper latitudes compared to SDB, a dominance reflecting the greater potential of coarse material in high latitudes as a residue of late Quaternary glacigenic processes. SDB were often associated with low angle coastal plains, but barriers are potentially viable wherever sand sinks can be found, e.g. Holocene restructuring of deltas.

Barriers are generally seen as having a Holocene timescale, though this may be truncated to periods since the last major eustatic sea-level variation occurred. In particular many barriers around the North Atlantic have a history commensurate with the mid-Holocene decline in the rate of relative sea-level (RSL) rise. This emphasizes RSL as a datum control on wave activity and barrier development, and the rate of RSL change as a control on the tempo of barrier migration. Early work on the US east coast barriers identified their development with a Holocene TRANSGRESSION that swept up available sediment and concentrated it into the barriers. This perspective has been challenged in that, although it sounds intuitively correct, the actual initial mechanism for onshore concentration of sediment in the surf zone with rising RSL has not been verified, hence the switch to spit elongation as a more coherent model of barrier building in the face of rising RSL. An alternative perspective is to consider aggraded barriers as a consequence of falling RSL. This suggestion has been made for some Florida and Texas barriers, identifying a higher than present sea level during the mid to late Holocene, as a consequence of which barriers developed and aggraded during the subsequent regression. The lack of an obvious mechanism for barrier build-up during a transgression should not be confused with the more understandable behaviour of an existing barrier during subsequent transgressive phases. Jennings *et al.* (1998) suggest that the longshore coherency of GDB relates to the rate of RSL rise: slower rates ($<2\,\mathrm{mm\,a^{-1}}$) mean reduced longshore sediment supply and the cannibalization of existing barrier segments to the point of barrier breaching. Barrier migration rates generally relate to RSL rise rate, though severely reduced GDB may be overwhelmed by the ambient RSL rise and flattened, to be rebuilt further onshore (i.e. overstepped; Forbes *et al.* 1991).

Wave climate is a major constraint on barriers. Many barriers are prominent in areas that are dominated by oceanic swell. These swells are likely to be transformed as constructional breakers, refracting parallel to existing shorelines and minimizing offshore sediment losses. This does not mean that barriers cannot emerge in storm-wave dominated areas, indeed such areas often show GDB, which tend to move onshore during storms, e.g. Atlantic Canada (Orford *et al.* 1996) and Patagonia (Isla and Bujalesky 2000). Storms are of crucial importance to barrier development. Most storms are expected to work at moving sediment offshore (gravel barriers aside), however as the severity of the storm goes upscale, then the emphasis of sediment transport switches from offshore to onshore. This threshold is reached when run-up physically reaches beyond the barrier crest and transports beach face material over the crest on to the back slope. This process is overwashing and its product is known as washover sedimentation. The latter is most obvious where overwash generates distinct flow channels (throats) through the crest and down the back slope ending in fan splay deposition beyond the previous back-barrier shoreline; such splays are the basis for barrier retreat. The position of throats and fans are partially dictated by the longshore gradients of the breaking wave and morphology of the barrier's seaward face. Beach faces showing cuspate morphology can preferentially set a rhythmic spacing to overwash, which in turn forces a consistent barrier retreat. As storm severity increases then the volume and depth of surface flows over the crest cause lateral extension of throats to the point of coalescence and the overt channels are lost in the face of generalized mobilization (sluicing overwash) of the barrier top into the back-barrier bay area. As sediment is washed into the back-barrier bay, it helps to build up a sub-aqueous sedimentation base for later marsh sedimentation (pads). These pads clearly help to fill in an accommodation space over which the barrier will migrate, such that the shallower the back-barrier area becomes the faster the potential for the barrier to migrate. Some controversy has been generated by the perceived influence of dunes on this migration (and hence survival) process, as dunes will block overwash, or spatially defuse the longshore consistency of overwash and hence reduce migration rate. This led to a short-lived management policy on the US barriers of advocating bulldozing the dune so as to promote overwash – scarcely a recipe for short-term barrier stability. The reverse is now in favour, that of promoting dune sedimentation as a 'sustainable' coastal defence.

Severe storms are also important for the development of cross-barrier breaches. US east coast barriers are vulnerable to hurricanes generating storm surge flow whose elevations are higher than the barrier crest. Hurricane overwash can back-up in the back-barrier bays, impounded by onshore surges, and flow laterally to escape; (1) through old breaches in the barrier; (2) along old overwash channels and (3) at any topographic low point on the barrier that erodes to form a new cross-barrier breach. These breaches can be quickly sealed up by post-hurricane beach face longshore sediment transport, or breaches may persist over decades. Sediment can be transported either way through breaches and surge deposits on the bay side can act as shallow water platforms for future barrier retreat. It is these overwash events that are responsible for most coarser sediment (i.e. high-energy) found in low-energy back-barrier environments.

Tidal range is considered as an indirect control of barrier development. As the tidal range expands, then so does the effectiveness of the tidal current flow regime. Increased tidal prisms maintain the storm-generated breaches. This tidal regime prevents post-storm breach healing and diverts longshore sediment into stores formed as flood or ebb deltas offset from the barrier. The larger the tidal range, the more breaches can be maintained in terms of hydraulic efficiency. Texas has one of the longest barriers in the USA – Padre Island is over 100 km in length and though subjected to hurricane attacks and suffering some breaches (mostly sealed) does not have a sufficient tidal prism (micro-tidal) to maintain the hurricane openings. New Jersey also has storm breaches with a low tidal range, but low longshore sediment availability reinforced by human interference is holding open breaches that would be closed elsewhere. South Carolina experiences a meso-tidal regime that maintains a greater longshore density of breaches or tidal passes, sufficient to define barrier islands. As a concomitant of inlet development and maintenance, there are substantial flood and ebb deltas linking the barrier islands into a hydraulically efficient network conditioned by tidal prism. If the prism alters due to back-barrier reclamation then inlet dimensions will alter, e.g. the Friesian Islands, German Wadden Sea (FitzGerald *et al.* 1984). The ebb/flow characteristics may be conditioned by saltmarsh growth within the bay, acting as a retardant to balanced flood/ebb tidal flow asymmetry. It is rare to find barriers in macro-tidal ranges, but when they do occur (north Norfolk, England) the forcing of the barrier is more to do with longshore sediment supply than tidal inlet forcing, however subsequent evolution in the face of diminishing sediment may mean that the potential for segmentation is great and the apparent morphology of barrier islands may be superimposed.

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JULIAN ORFORD

BASE LEVEL

The concept that there is an effective lower limit to erosional processes. Powell (1875) first named the concept: ‘we may consider the level of the sea to be a grand base level, below which the dry lands cannot be eroded; but we may also have, for local and temporary purposes, other base levels of erosion, which are the levels of the beds of the principal streams which carry away the products of erosion.’ Chorley and Beckinsale (1968) recognized four main interpretations of the term.

1. Grand base level or ‘ultimate base level’ which is the plane surface forming the extension of the sea under the lands.
2. Temporary or structural base level, whereby there is a limit to downward erosion of an ephemeral character imposed headwards of a resistant outcrop.
3. Base-levelled surface, which is an ultimate or penultimate topographic surface.
4. Local base level, as for example in areas of interior drainage under an arid cycle.

The first of these usages occupied a fundamental place in the cycle of erosion concept of W.M. Davis (1902). Base-level changes are also crucial in the study of fluvial terraces, deltas and other depositional systems (Koss *et al.* 1994). They can result from tectonic activity, sea-level change and river capture (Mather 2000).

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A.S. GOUDIE

BEACH

A beach is a wave deposited accumulation of sediment located between modal wave base and the upper swash limit. Beaches may be composed of fine sand through boulders and may range from low energy, narrow strips of sand lapped by low wind waves, to high energy systems exposed to persistent 2–3 m high swell which breaks across 500 m wide surf zones. Beach systems are also located in all tide ranges, in all latitudes, in all climates and on all manner of coast, from the low coastal plains fronted by long beaches to small pockets of sand wedged in at the base of massive cliffs. Therefore beaches exist in a wide spectrum of wave, tide and sediment combinations and geological settings.

In two dimensions, however, all beaches will contain three dynamic zones – of wave shoaling, wave breaking and swash–backwash. The wave shoaling zone extends from the modal wave base where average waves can entrain and move sediment shoreward, to the outer breakpoint. The wave shoaling zone is dominated by asymmetrical wave orbital motions which produce a concave upward profile sloping less than 1°, dominated by wave ripples which become increasingly three dimensional close to the breaker zone. The depth and width of this zone is dependent on wave height and sediment size. On high energy coasts it extends out to depths of 30 m or more which may lie 2–3 km offshore, while on low energy systems it may only extend to low tide a few metres from the shore. Sediment is often graded across the system and fines seaward.

The surf zone is located between the breakpoint and shoreline. The surf zone has the greatest potential for complex dynamic processes and resulting topography and bedforms. The dominant processes start with wave breaking. Onshore currents are generated by wave bores and orbital currents associated with reformed waves. Longshore currents result from oblique waves and bi-directional rip feeder currents. Offshore flows are driven by wave reflection, discrete rip currents and bed return flow. The width of the surf zone is dependent on surf zone gradient, a function of sand size and wave height. It will be as narrow as a few metres on a steep reflective beach, typically 50–100 m wide on a single bar intermediate beach and up to several hundred metres on a high energy dissipative beach. The presence and number of bars will increase with decreasing wave period and decreasing gradient. Longshore variations in form and processes are driven by longshore changes in wave conditions, and by three-dimensional bar and rip topography.

Surf zone topographic features include shore parallel bars and troughs with waves breaking over the bars and reforming in the troughs. Swell coasts rarely have more than two bars (see BAR, COASTAL), while energetic sea coasts may have several bars. On intermediate beaches with cellular rip circulation, alternating shore transverse bars and rips, also known as crescentic bars, can occur. When present they dominate the inner bar and can lead to a rearranging of the shoreline to produce rhythmic topography. Surf zone BEDFORMS reflect the changing velocity and direction of currents and depth of water, and range from flat bed over the shallow bars, to wave orbital and shore perpendicular current ripples in the troughs, to shore parallel seaward migrating ripples in the rip channels.

The swash zone extends up the beach from the shoreline, from where the wave breaks or bore collapses, to the limit of swash. The swash zone is always an upward sloping zone of wave uprush (swash) and backwash. Its slope is directly related to grain size and inversely related to wave height and may range from 1° to 10°. It is usually featureless, with ripples only produced by strong backwash in the lower swash zone. The swash may, however, be superimposed on high tide beach cusps or a berm, and mesoscale megacusps. Beyond the limit of normal swash is the backshore, a zone of either wind-blown aeolian deposits and/or storm wave overwash.

In three dimensions beaches respond to a greater range of variables and become increasingly complex. First is the alignment of the shoreline to the dominant wave crest, which produces swash aligned beaches. As waves refract around headlands and nearshore topography, the wave crests bend to parallel the contours. At the shoreline the beach also moves to parallel the wave crests so as to minimize longshore transport, and thereby produce a more stable shoreline in equilibrium with the wave crest. Where waves arrive at a persistent oblique angle to the shore, particularly on long beaches, then sediment is moved downdrift by the longshore surf zone currents generated by the waves, producing a drift aligned shore.

Beach type

Beach type refers to the morphodynamic character of a beach system, which is a product of the interaction of waves, tide and sediment. Beaches may be of three types: wave dominated, tide modified and tide dominated. Wave-dominated beaches occur where waves are high relative to the tide range. This can be defined quantitatively by the relative tide range

$$RTR = TR/H_b \qquad (1)$$

where TR is the spring tide range and H_b the average breaker wave height. When $RTR < 3$ beaches are tide dominated, when $3 < RTR < 15$ they are tide modified and when the $RTR > 15$ they become tide dominated.

Within each of these beach types a range of wave and sediment combinations can occur which will influence the actual state of the beach. The dimensionless fall velocity

$$\Omega = H_b/T\ W_s \qquad (2)$$

where T is wave period (s) and Ws sediment fall velocity ($m\,s^{-1}$) can be used to quantify beach state. They range from the lower energy reflective ($\Omega < 1$) favoured by low waves, longer periods and coarser sediments, to dissipative ($\Omega > 6$) with high waves, shorter periods and fine sand. In between are the more rip dominated intermediate beaches ($\Omega = 2$–5) produced by moderate to high wave conditions.

WAVE-DOMINATED BEACHES

Wave-dominated beaches consist of three types: reflective, intermediate and dissipative.

Reflective beaches

Reflective beaches are produced by combinations of lower waves, longer periods and particularly coarser sands. They occur on sandy open swell coasts when waves average less than 0.5 m, and on all coasts when beach sediments are composed of coarse sand or coarser, including all gravel and boulder beaches, even under higher waves. However, they are all characterized by a concave upward nearshore zone of wave shoaling that extends to the shoreline. Waves then break by plunging and/or surging across the base of the beach face. The ensuing strong swash rushes up the beach, combining with the coarse sediment to build a steep beach face (4°–10°), commonly capped by well-developed beach cusps and/or a *berm* (Plate 12), possibly backed by a runnel where the high tide swash may temporarily accumulate. When the sediment consists of a range of grain sizes, the coarser grains accumulate as a coarser steep *step* below the zone of wave breaking, at the base of the beach face.

The cusps are a product of cellular circulation on the high tide beach resulting from sub-harmonic edge waves produced from the interaction of the incoming and reflected backwash. The high degree of incident wave reflection off the beach face is responsible for the naming of this beach type, i.e. reflective. Apart from the cosmetic beach cusps and swash circulation these are essentially two-dimensional beaches with no longshore variation in either processes or morphology. On sand beaches they also represent the lower energy end of the beach spectrum and as such are relatively stable systems only responding to an increase in wave height. Such an increase induces a growth in the swash energy and erosion of the swash zone.

Plate 12 A lower energy reflective beach with wave surging up the moderately steep beach, Horseshoe Bay, South Australia (Andrew D. Short)

Intermediate beaches

Intermediate beaches are called such as they represent a suite of beach types between the lower energy reflective and higher energy dissipative. They are the beaches that form under moderate to high waves, on swell and sea coasts, in fine to medium sand. The two most distinguishing characteristics of intermediate beaches are (1) a surf zone, and (2) cellular rip circulation (see RIP CURRENT) commonly associated with rhythmic bar and beach topography (Plate 13). Since intermediate beaches can occur across a wide range of wave conditions, they consist of four beach states ranging from the lower energy low tide terrace to the rip dominated transverse bar and rip and rhythmic bar and beach, and the high energy straighter longshore bar and trough.

Intermediate beaches are controlled by processes related to wave dissipation across the surf zone which transfers energy from incident

Plate 13 Well-developed transverse bar and rips, Lighthouse Beach, New South Wales, Australia (Andrew D. Short)

waves with periods of 2 to 20 s, to longer infragravity waves with periods >30 s. As a consequence, incoming long waves associated with wave groupiness, increase in energy and amplitude across the surf zone and are manifest at the shoreline as wave set up (crest) and set down (trough). The long wave then reflects off the beach leading an interaction between the incoming and outgoing waves to produce a standing wave across the surf zone. It is believed that standing edge waves trapped in the surf zone are responsible for the cellular circulation that develops into rip current circulation and associated transverse bars and rips. These are in turn responsible for the high degree of spatial and temporal variability in intermediate beach morphodynamics.

The low tide terrace (or ridge and runnel) beaches are characterized by a continuous attached bar or terrace located at low tide. They form under lower waves (0.5–1 m) and usually undergo temporal variation between low tide when the waves break and dissipate across the bar, while at high tide they may remain unbroken and surge up the reflective high tide beach face. Weak rips may occur at mid to low tide.

Transverse bar and rip beaches form under moderate waves (1–1.5 m) on swell coasts and consist of well-developed rip channels, which are separated by shallow bars, the bars attached and perpendicular or transverse to the beach (Plate 13). Variable wave breaking and refraction across the shallow bars and deeper rip channels lead to a longshore variation in swash height and approach, which reworks the beach to form prominent megacusp horns in lee of bars, and embayments in lee of channels. Water tends to flow shoreward over the bars, then into the rip feeder channels. The flow moves close to the shoreline and converges laterally in the rip embayment. It then moves seaward in the rip channel as a relatively narrow (few metres), strong flow (0.5–1 $m\,s^{-1}$), called a *rip current*. This beach state has extreme spatial-longshore variation in wave breaking, surf zone and swash circulation and beach and surf zone topography, leading to a highly unstable and variable beach system.

The rhythmic bar and beach state forms during periods of moderate to high waves (1.5–2 m) on swell coasts. The high waves lead to greater surf zone discharges that require deeper and wider rip feeder and rip channels to accommodate the flows. Rips flow in well-developed rip channels, separated by transverse bars, however the bars are detached from the beach by the wider feeder channels.

The longshore bar and trough systems are a product of periods of higher waves which excavate a continuous longshore trough between the bar and the beach. Waves break heavily on the outer bar, reform in the trough and then break again at the shoreline, often producing a steep reflective beach face (coarser sand) or low tide terrace (finer sand). Surf zone circulation consists of both cellular rip flows as well as increasingly shore normal bed return flows (see below).

Dissipative beaches

Dissipative beaches represent the high-energy end of the beach spectrum. They occur in areas of high waves, prefer short wave periods, and must have fine sand. They are relatively common in exposed sea environments where occasional periods of high, short storm waves produce multi-barred dissipative beach systems, as in the North and Baltic seas. They also occur on high-energy mid-latitude swell and storm wave coasts as in northwest USA, southern Africa, southern Australia and New Zealand. On swell coasts waves must exceed 2–3 m for weeks to generate fully dissipative beaches. They are characterized by a wide long gradient beach face and surf zone, with two and more shore parallel bars forming across the surf zone (Plate 14). The

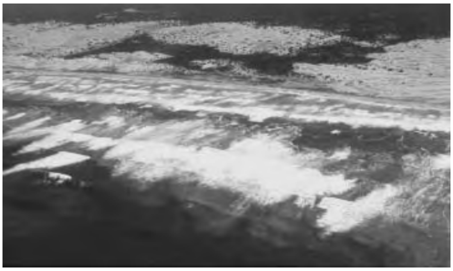

Plate 14 High energy dissipative beach containing an inner bar, trough and wide outer bar, Dog Fence Beach, South Australia (Andrew D. Short)

low gradient is a product of both the fine sand, as well as the dominance of lower frequency infragravity swash and surf zone circulation, which act to plane down the beach. The name comes from the fact that waves dissipate their energy across the many bars and wide surf zone.

The dissipation of the incident wave energy leads to a growth in the longer period infragravity energy, which becomes manifest as a strong set up and set down at the shoreline. The standing wave generated by the interaction of the incoming and outgoing waves may have two and more nodes across the surf zone. It is believed that the bar crests form under the standing wave nodes and troughs under the antinodes. Surf zone circulation is vertically segregated. Wave bores move water shoreward toward the surface of the water column. This water builds up against the shoreline as wave set up. As it sets down the return flows tends to concentrate toward the bed, which propels a current across the bed of the surf zone (below the wave bores) called *bed return flow*.

Like the reflective end of the beach spectrum dissipative beaches are remarkably stable systems. They are designed to accommodate high waves, and can accommodate still higher waves by simply widening their surf zone and increasing the amplitude of the standing waves, while periods of lower waves are often too short to permit substantial onshore sediment migration.

TIDE-MODIFIED BEACHES

Most of the world's beaches are affected by tide. On most open coasts where tides are low (<2 m) waves dominate and tidal impacts are minimized. However, as tide range increases and/or wave height decreases tidal influences become increasingly important. To accommodate these influences beaches, still by definition wave-formed, can be divided into tide-modified and tide-dominated types, as defined by equation 1 (see p. 64).

The major impact of increasing tide range is to shift the location of the shoreline between high and low tide, which – depending on the shoreline gradient – will be tens to hundreds of metres. This shift not only moves the shoreline and accompanying swash zone, but also the surf zone, if present, and the nearshore zone. While wave-dominated beaches have a relatively 'fixed' swash–surf–nearshore zone, on tide-modified beaches they are more mobile and transient. The net result is a smearing of the three dominant wave processes of shoaling (nearshore zone), breaking (surf zone) and swash (swash zone). A section of intertidal beach can be exposed to all three processes at different states of the tide. Second, because all three zones are mobile, except for a brief period at high and low tide, there is a reduction in the time any one process can fully imprint its dynamics on a particular part of the beach. As a consequence there is a tendency for swash processes only to dominate the spring high tide beach, for surf zone processes only to dominate the beach morphology around low tide, during the turn of the low tide, while shoaling wave processes become increasingly dominant overall, producing a lower gradient, featureless, concave beach cross section.

Tide-modified beaches can contain three states. When waves are lower ($\Omega < 1$–2) they consist of a steep reflective high tide beach face fronted by a wide low gradient low tide terrace, often with a sharp break in slope between the two. At low tide waves dissipate across the terrace, while at high tide they pass unbroken across the now submerged terrace to surge up the steep reflective high tide beach. In areas of moderate waves ($\Omega = 2$–5) and tide range ($RTR = 3$–7) the tide-modified beaches consist of a high tide reflective beach, a usually wider intertidal zone, and a low tide zone dominated by surf zone morphology, which may include transverse and rhythmic bars and rips. In moderate energy sea environments a series of shore parallel ridges and runnels may develop (Plate 15). Higher energy tide-modified beaches ($\Omega > 5$) composed of fine sand are characterized by a wide, low gradient concave upward, flat and featureless, beach and intertidal system, called an ultradissipative beach (Plate 16).

Plate 15 Reflective high tide beach (foreground) fronted by three ridges and runnels, Omaha Beach, France (Andrew D. Short)

Plate 16 Rhossili Beach, Wales, a high energy ultradissipative tide-modified beach, shown here at low tide (Andrew D. Short)

TIDE-DOMINATED BEACHES

Tide-dominated beaches occur when the RTR > 15, that is the tide range is more than 15 times the wave height. As the maximum global tide range is about 12 m, and usually much less, this means that most tide-dominated beaches also receive low (< 1 m) to very low waves, and are commonly dominated by locally generated wind waves. Tide-dominated beaches are characterized by a usually steep, narrow high tide beach, and a wide, low gradient (< 1°) sandy intertidal zone, which in temperate to tropical locations is usually bordered by subtidal seagrass meadows. They consist with decreasing wave energy of three states: (1) a beach and sand ridges, containing multiple, low amplitude, shore parallel sand ridges across the intertidal; (2) a reflective beach fronted by a usually wide, flat, featureless intertidal sand flat; (3) a tidal sand flat, which is a beach in so far as it has a high tide beach. However, the fronting often wide tidal flats are dominated by the tides and not waves, and may contain intertidal biota and tidal drainage features. It is part of the transition between the beaches and the often muddy, tidal flats.

Further reading

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SEE ALSO: bar, coastal; bedform; rip current

ANDREW D. SHORT

BEACH CUSP

Beach cusps are crescentic concave-seaward regularly spaced features occurring along the shoreline. The term has been used for features with spacing ranging from 10 cm to many hundreds of metres, although larger examples tend to be called rhythmic beach features with the term swash cusp being used for features with a spacing less than tens of metres (Hughes and Turner 1999). Beach cusps (swash cusps) are most commonly associated with medium to coarse sands, shingle or mixed sand-shingle sediments, on steep beaches demonstrating significant wave reflection. Their amplitude ranges from almost zero to over 1 m. On beaches with high tidal range, multiple sets of cusps may be present at different levels. Beach cusps consist of embayments or swales separated by triangular horns which are normally comprised of coarser sediments.

A number of different swash flow patterns in and around cusps have been reported in the literature. Under low energy conditions, oscillatory flows (with swash largely unaffected by cusp morphology), horn divergent flows (uprush flow separation at the horn with water returning from the embayment), and horn convergent flows (uprush entering the cusp in the embayment and returning along the sides of the horn, converging at its apex) have all been reported. Under high energy conditions sweeping flow (alongshore directed water movement) and swash-jet flows (where incoming swash is held back by the backwash until it develops sufficient head to break through as a jet in the centre of the embayment) can also occur (Masselink and Pattiaratchi 1998).

Numerous conflicting ideas have been proposed for the mode of formation of beach cusps, including processes of accretion, processes of erosion or a combination of both, and theories based on instabilities on wave breaking, alongshore sediment transport, and intersecting wave trains. Debate continues on the formation of beach cusps, with two theories based on fundamentally different mechanisms dominating.

Cusp development caused by standing edge WAVES at either twice the period (subharmonic) of, or synchronous with the incident waves has been proposed. This hypothesis can explain regular spacing (cusp spacing equal to edge wave length for synchronous edge waves, or half edge wave length for sub-harmonic edge waves), but can also explain complex quasi-regular spacing if more than one mode of edge wave is present. Werner and Fink (1993), however, proposed a self-organization model for beach cusp formation, with topographic depressions resulting in feedback mechanisms between swash and morphology resulting in self-emergence of beach cusps. Cusp spacing is proposed as being proportional to swash excursion length. Predications of similar beach cusp spacing based on the quite distinct self-organization and edge wave theories of cusp formation have meant that field studies have been unable to discriminate between the two models. It is also possible that edge waves may initiate cusp development with self-organization then allowing for the growth of the features. Many field studies have reinforced the importance of feedback processes between swash and morphology (Masselink *et al.* 1997).

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SEE ALSO: beach; wave

KEVIN PARNELL

BEACH–DUNE INTERACTION

The original generation of the wave–beach–dune model of beach and dune interactions was formulated by Hesp (1982). It was followed by the publication of a robust micro-tidal beach model with reasonably high predictability (Wright and Short 1984). The beach model enabled scientists to classify micro-tidal beaches into six states with characteristic morphologies, mobility, and modes of erosion and accretion. Subsequent research has extended the original model to meso- and macro-tidal beaches (see BEACH). An understanding of beach and backshore morphology for different surfzone-beach types allowed Hesp (1982) to develop actual and theoretical links between backshore morphology, potential aeolian transport, foredune state and morphology, and dunefield type and development.

Surfzone-beach state

The micro-tidal beach models classified beaches into six states, with the dissipative state at the high wave energy (>2.5 m) extreme and reflective state at the low wave energy (<1 m) extreme. Four intermediate beach states occur between these states. Dissipative beaches are characteristically high wave energy beaches and have the highest potential onshore sediment supply (Hesp 1988). Note, however, that beaches may also be dissipative because of the presence of very fine sand (hence low gradient) or abundant sand, so some dissipative beaches may, in fact, be low wave energy beaches. They are typically wide, display flat to concave morphologies (no berms), low gradients and minimal backshore mobility. The latter refers to the coefficient of variation of mean shoreline position, and in reality refers to the amount of volumetric and profile change the beach and backshore experiences over time and through erosion to accretion phases. Reflective beaches are characteristically low wave energy beaches with low potential onshore wave driven sediment transport. Note that they may also be moderate to high wave energy where sediments are coarse sand to boulders. They are relatively steep, narrow, linear to terraced (i.e. display a berm form) morphologies, with low backshore mobility. Intermediate beaches range from wide, relatively flat beaches with low mobility at the higher energy end of the spectrum, through moderate width beaches with pronounced berms and high mobility to narrow beaches and moderate to low mobility berms at the reflective end of the range. Rips dominate surfzone processes in the intermediate range.

Beach-backshore width and morphology, fetch and potential aeolian transport

Beach width is important in determining fetch which is critical for determining the volume of sand delivered across the backshore and to dunes (Davidson-Arnott 1988). Beach morphology is important because the greater the morphological variability, the more likely that wind velocity decelerations and variations take place across the backshore. Hesp (1982, 1999) showed that the wind flow across a wide, low gradient, dissipative beach displayed minimal flow variation and gradually accelerated across the backshore, thus maximizing potential aeolian transport. The wind flow over the berm crest of an intermediate beach was accelerated but decelerated leeward of the berm crest. High narrow berms typical of some reflective beaches display significant flow disturbance and deceleration leeward of the berm crest (Short and Hesp 1982). Sherman and Lyons (1994) modelled wind flow and potential sediment transport on a flat beach, low berm and high berm profiles, and found that sand transport off the dissipative beach was 20 per cent higher than off the reflective beach if just slope and grain size were taken into account. When moisture content was added, transport rates were nearly two orders of magnitude higher off the dissipative beach compared to the reflective beach. Note, however, that each beach had the same width (100 m wide), whereas actual reflective beaches and many intermediate beaches are considerably narrower.

Beach mobility is important because the greater the beach mobility, the greater the morphological variability. The latter affects the fetch such that the beach width is at times quite narrow, at times quite wide, particularly for intermediate beaches. It is also important because alternating episodes of cut-and-fill result in varying beach morphologies which then affect airflow and sediment transport as indicated above.

Thus, the link between surfzone beach state, aeolian sediment transport and landward dunes is that modal dissipative beaches have maximum potential aeolian sediment transport, reflective beaches minimal potential aeolian sediment transport, and intermediate beaches range from relatively high potential at the dissipative end to low potential at the reflective end. Note that a minimal sediment supply ('minimal' is currently undetermined) is required.

Aeolian sediment transport and foredune morphology

An examination of foredune heights and volumes on dissipative to reflective beaches allows one to examine the validity of the links above. Since established foredunes occupy a foremost backshore position, they are a medium-term indicator of beach and backshore processes. Hesp (1988) measured incipient and established foredune volumetric changes over several years at Myall Lakes National Park in New South Wales, Australia to find that a modal reflective beach with the same wind exposure as a modal dissipative beach received 60 per cent less sand than the dissipative beach over the same survey period. Intermediate beach volumes ranged from relatively high to relatively low between the dissipative and reflective beaches.

Surveys of established foredunes, which have been present for potentially several hundred years, provide further evidence that there is a strong link between surfzone-beach type and foredune height and volume. Hesp (1982, 1988) demonstrated that in the Myall Lakes National Park the smallest established foredunes, with lowest sediment volumes were found on reflective beaches, while the highest and largest foredunes occurred on dissipative beaches. Similar results are reported by Davidson-Arnott and Law (1990). Intermediate beaches followed a trend from low to high volumes on lowest to highest energy intermediate beaches respectively (see reviews in Sherman and Bauer 1993 and Bauer and Sherman 1999).

Foredune ecology

The vegetation cover, species richness and zonation of foredunes is determined by several factors, but sediment supply and sand deposition rate, and salt spray aerosol levels are two important factors (Hesp 1991). Simultaneous studies carried out on adjoining reflective, intermediate and dissipative beaches show that salt spray aerosol levels are related to surfzone-beach type. Dissipative beaches have the widest surfzones, the greatest number of breaking waves, and highest wave heights and the highest salt aerosol levels. Reflective beaches often have only one breaking wave, narrow to very narrow surfzones, and low wave heights and the lowest salt aerosol levels. All other factors being equal, foredune species richness and zonation tends to be greatest and narrowest respectively on reflective beaches (low sediment supply and salt aerosols), and lowest and widest on dissipative beaches (highest sediment supply, high salt aerosol levels) (Hesp 1988).

Foredune stability and type, erosion processes and dunefield development

Foredunes bear a morphological imprint dictated, in part, by modal surfzone-beach erosion and accretion modes, and the wind often accentuates this morphological imprint. Dissipative beaches are typically eroded by swash bores and undertow commonly associated with elevated water levels and storm surge. Beach erosion and dune scarping is laterally continuous alongshore, and at times catastrophic. Hesp (1988) and Short and Hesp (1982) theorized that such laterally continuous alongshore, large-scale foredune scarping would on occasions lead to large-scale foredune destabilization. Transgressive dunefields would most likely result from the breakdown of the large established foredune. In fact, transgressive dunefields are most commonly found on high energy dissipative surfzone-beach systems (e.g. Australian and South African coasts below the tropics; west coast USA; west coast North Island, New Zealand).

Intermediate beaches are characterized by localized, arcuate rip embayment erosion during storms. Such arcuate erosion extends well into the foredune during extreme events resulting in large-scale but localized foredune scarping. Topographic funnelling of the wind may result in the evolution of blowouts and eventually parabolic dunes at these locations. On average, many higher energy intermediate beaches display parabolic dune complexes (Hesp 1982, 1988; Short and Hesp 1982).

On south-east Australian beach systems where overwash events are minor to absent, where sediment supply is generally not limited, and where an aggressive pioneer grass (*Spinifex* sp.) exists, relict foredune plains are common, particularly on the moderate energy intermediate beaches. Here established foredune stability is maintained to various degrees, and progradation over the last 6,000–7,000 years has led to the development of foredune plains.

Reflective beaches are characterized by accentuated swash during storms and laterally continuous alongshore beach erosion. Recovery is fairly rapid. Foredunes remain relatively stable over time, and because they are typically small, with limited sediment supply, little dune transgression results. Thus reflective beaches are characterized by a single foredune, or a few relict foredunes.

Role of sediment supply and other factors

There is no doubt that sediment supply, wind energy, sea-level state (transgressive, stable, regressive), return interval and magnitude of extreme storm events, and Pleistocene inheritance factors will all, at times, and in some places, be a controlling variable in beach–dune interactions. If sediment supply is limited, sea level is rising, and coastal erosion is the general rule, the model as outlined above may not work in part or perhaps at all (Psuty 1988).

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PATRICK HESP

BEACH NOURISHMENT

Beach nourishment is the act of placing sediment (termed fill) on a beach by artificial means using sources outside the nourished area. Nourishment is primarily conducted to overcome a sediment deficit and create a beach of sufficient width to protect existing buildings and infrastructure from wave attack, but it also can enhance the value of urban locations for tourism or create new natural environments.

Debates have occurred over the cost effectiveness of nourishment and whether nourished beaches erode more rapidly than predicted in design studies (Houston 1991; Pilkey 1992), but projects are increasingly implemented, and nourishment is now the principal option for shore protection in some countries. Nourishment is conducted at all levels of management, from the national level to private homeowner groups. Borrow areas for fill materials include offshore, inlets, backbays, rivers and glaciated uplands. Opportunistic sources from dredging of harbours, marinas, lagoons and inland construction sites are also used (Nordstrom 2000).

Nourishment occurs on the upper beach, the nearshore, offshore on stable berms (designed to alter wave conditions) and active berms (that change shape or migrate onshore), and on existing dunes or on the backbeach to create new dunes. Large-scale nourishment operations on the upper beach commonly use a pipeline to transport a sand/water slurry. Small-scale operations transport sediment in dump trucks. The nourished beach is then often reshaped by bulldozers. The result of upper beach nourishment is a high, wide beach with an unstable shape, but the fill is easy to place, provides good initial protection against wave overwash, and creates a wide recreation platform. Conspicuous losses may subsequently occur on the upper beach as the fill adjusts to a more natural equilibrium shape. Fill sediments on the foreshore are reworked by waves and often become similar to pre-nourished sediments in size and sorting, but fill sediments on the backbeach above the zone of wave reworking retain characteristics that differ from native sediments.

Nearshore nourishment occurs by spraying sediment as a sand/water slurry or dumping it from shallow-draught barges. By placing sediments directly in the dynamic surf zones, losses through time are not visible and aesthetics are not spoiled by different sediments on the backbeach, but a beach nourished this way evolves slowly. Offshore berms are often implemented as disposal areas for sediment dredged from navigation channels, and more study is required to evaluate their use as protection structures.

Sediment bypassing (artificially transporting sediment to the downdrift side of obstacles to littoral drift) and backpassing (artificially transporting sediment from downdrift deposits to updrift eroding zones) may also be considered nourishment projects. Bypassing is gradually becoming more common at inlets where jetties or dredging of navigation channels interrupt longshore sediment transport. Backpassing is now most frequently conducted in small-scale trucking operations, but it may become more significant in the future as ready supplies of external sediment for nourishment projects become exhausted.

Nourishment of the upper beach can alter aeolian transport by (1) increasing the source width for entrainment of sediments; (2) adding fine sediment that is more readily transported; (3) changing moisture-retention characteristics; (4) changing the shape of the beach or dune profile; and (5) changing the likelihood of marine erosion of the incipient foredune (van der Wal 1998). Rapid dune growth can occur on nourished beaches, especially when sand fences and vegetation plantings are used to trap sand.

Dunes may be created directly by mechanically depositing sediment. Most dune nourishment operations place the new fill in front of the existing dune to create a sacrificial structure or on top of it to increase the level of protection against wave overwash; more rarely, a new dune may be built behind an existing foredune (Nordstrom 2000). Dunes built and used as protection structures can evolve into a condition that functions naturally or appears natural in terms of surface vegetation (Nordstrom *et al.* 2002).

Nourished beaches benefit threatened species by providing habitat that would otherwise be unavailable, but detrimental ecological impacts can occur due to (1) mechanical removal of habitat in borrow areas; (2) burial of habitat in nourished areas; (3) increased turbidity and sedimentation; (4) disruptions to foraging, nesting, nursing and breeding; (5) change in sediment characteristics, wave action and beach state; and (6) change in community structure and evolutionary trajectories, including enhancement of undesirable species (Nelson 1993; National Research Council 1995). Detrimental effects are often considered temporary, but little is known about long-term, cumulative impacts and critical thresholds.

Human activities, such as driving on the beach or raking the beach to remove litter, can eliminate incipient topography and vegetation and prevent formation of natural landforms, so true restoration of landforms and habitats may not occur in the absence of controls on subsequent human activities (Nordstrom 2000). The great importance of nourishment as a form of shore protection and as a sediment resource that can evolve into naturally functioning landforms makes this option an important area of future geomorphological research. To be effective, the nourished beach must be considered as a landform in its own right and as a source of sediment for evolution of other landforms landward and downdrift of it, rather than merely as an engineering structure or recreation platform.

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KARL F. NORDSTROM

BEACH RIDGE

Beach ridges are azonal accumulation forms created on the shores of seas and lakes. They are usually subparallel ridges of sand, gravel or pebble, as well as detritus of shell, situated in the foreshore zone, which is the boundary of low and high water's range. Older, subfossil complexes of beach ridges may actually appear in the backshore zone, which lies above the high water's range. Beach ridges forming at the present day are roughly parallel to the coast. The height of their crests is usually a little bit higher than mean high tide or storm, and the bottom of the adjacent troughs or swales have elevations not far from mean low water (Stapor 1975).

In Carter's (1986) opinion two types of beach ridges may develop on a progradational sea coast. The first type is a result of gradual accretion and coalescing of swash bars during a deposit's transport by wave action. This type of beach ridge is constructed by seaward dipping laminae of sand or gravel (swash lamination). The second type is connected with longshore bar emergence during low wave energy conditions and simultaneous fall of sea level. The morphology of these ridges is more complicated. They are constructed mainly by landward dipping laminae. However, tabular planar cross-lamination connected with landward migration of the lee slope of the emerge feature are also situated here. On the seaward slope of this type of beach ridge a thin layer of swash lamination can be present.

Beach ridges are also partially created by the processes of aeolian deflation and accumulation. There is often an accumulated cover of aeolian deposits on earlier formed ridges, stabilized by vegetation. As a result, on the beach ridges irregular hummock dunes or parallel foredune ridges can be situated. In this case, the sediments of beach ridges are usually separated from aeolian covers by fossil erosional surfaces with a shell or gravel pavement (Carter and Wilson 1990).

The formation of beach ridges is very dependent on conditions of beach supply by littoral deposits. Beach ridges are created only when wave action, and connected with it sea currents, provide more deposits to the beach than the waves can remove (Johnson 1919). Important factors during the creation of beach ridges are the bathymetry of the inner shelf, abundant sediment supply in the nearshore zone and also the wave energy regime and fluctuations of sea level. The average size of a beach's material is also a very important factor. On sandy beaches, the beach ridges accumulate during the low wave energy events, but on gravelly beaches the formation of ridges is usually connected with high wave energy events.

Beach ridges may appear as a single form, as well as a complex of forms, creating often expansive plains of beach ridges. These plains are especially characteristic of progradational coasts. The relief of the individual beach ridges is different. Their height may reach values from a few dozens of centimetres to about a few metres. The distance between two different beach ridges also varies. It is usually thought that the smaller, closely spaced ridges are formed during rapid beach progradation, and that the ridges of larger dimensions and greater spacing are connected with a relatively slower rate of growth (Taylor and Stone 1996).

Beach ridges are good palaeogeography indicators of past wave regimes, sediment supply, sediment source, climatic conditions, sea-level change and also isostatic emergence or submergence of land. If we can measure the absolute age of beach ridges, e.g. using radiocarbon or archaeological methods, we will be able to reconstruct the ancient shorelines' position as well as speed of coast progradation. Beach ridges can also be used to understand past relative sea-level changes and the history of deposit availability and abundance within the inner shelf.

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SEE ALSO: beach; beach–dune interaction; beach sediment transport; chenier ridge

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BEACH ROCK

'A consolidated deposit that results from lithification by calcium carbonate of sediment in the intertidal and spray zones of mainly tropical coasts' (Scoffin and Stoddart 1983: 401). Some authors have also used the term 'cay sandstone' to distinguish between rocks that are formed in the supratidal zone and beach rock or beach sandstone formed in the intertidal zone of coral reefs (see, for example, Gischler and Lomando 1997). Although the latitudinal limits of most contemporary beach rocks are approximately 35°N to 35°S, from time to time they do occur in higher latitudes, for example in north-west Scotland (Kneale and Viles 2000).

Beach rock is also widespread on beaches around the Mediterranean Sea (Plate 17) but is perhaps best known for its association with the calcium carbonate beaches of coral reef islands. Beach rock is geomorphologically important in that it preserves coastal landforms, provides a record of former sea levels, creates distinctive pavements and forms very rapidly. It also displays suites of characteristic erosional landforms that include micro-scarps, ridges and runnels and various weathering forms produced by biological processes, chemical erosion and salt attack.

The origin of beach rock has perplexed investigators ever since it was described in the early nineteenth century by travellers like Admiral Beaufort and Charles Darwin. Proposed mechanisms of formation include both physico-chemical and biological models. The former involve cement precipitation resulting from evaporation, CO_2 degassing owing to wave agitation and increasing temperature, and mixing of alkaline fresh water with sea water. Such models tend to have dominated the literature. However, the role of micro-organisms is now being seriously considered as a result of both

Plate 17 Beach rock developed on the south-east coast of Turkey at Arsuz near Iskenderun

field (Webb *et al.* 1999) and laboratory evidence (Neumeier 1999). High Mg calcite (often micritic) and aragonite are the commonest cement types, although low Mg calcite cements are common from temperate zone beach rocks. The cements occur most commonly as clean isopachous fringes of acicular crystals around grains, but meniscus and gravitational cements are also known (Scoffin and Stoddart 1983).

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A.S. GOUDIE

BEACH SEDIMENT TRANSPORT

Beach sediment transport occurs over the whole area where wave-induced currents are capable of moving sediment: in the shoaling zone, surf zone, breakers and swash zone, and also as aeolian sediment transport on beaches. Spatial gradients in the sediment transport rate determine positions of erosion and deposition and thus three-dimensional changes in the beach shape. Research on beach sediment transport concentrates on predicting the mechanics of transport, direction of transport, transport rate, transport volume, and changes in beach morphology.

The essential difference between sediment transport under a steady unidirectional flow and under waves is that oscillatory BOUNDARY LAYERS show a temporal variation which is not present under steady flow, growing and decaying twice every wave cycle as flows accelerate, decelerate and change direction. Boundary layers are not able to develop fully under oscillatory flow, and consequently are always thinner than under an equivalent unidirectional flow. This means that for a given free-stream velocity and bed roughness, the bed shear stress in oscillatory flow is always larger than beneath steady flow.

Beach sediment transport can be in the form of SUSPENDED LOAD, BEDLOAD and sheet flow (see SHEET EROSION, SHEET FLOW, SHEET WASH). Bedload transport is often modelled as a function of the shear stress acting on the sediment grains, while suspended transport is generally calculated as the product of velocity and concentration profiles. Suspended sediment transport generally receives more attention than bedload transport; however, this is mainly because instruments such as the optical backscatter sensor and acoustic backscatter sensor have been developed which are capable of making high-frequency measurements of suspended sediment concentrations simultaneously with velocity measurements. Because only suspended sediment transport can be measured in this way, many researchers conclude that the most reasonable approach at present is to assume that suspended sediment transport dominates when strong wave motion is present. However, this assumption will remain essentially untested until instruments are developed which are capable of high-frequency measurement of bedload transport.

A number of mechanisms contribute to beach sediment transport, including turbulence, mean currents, currents generated by oscillatory waves at incident and infragravity frequencies, and wave–current interaction. Wave-induced currents include both unidirectional currents (such as longshore currents, RIP CURRENTs and undertow) and rapidly reversing asymmetrical cross-shore currents which flow onshore under the crest of the wave and offshore under the trough. In combined flows, oscillatory wave motion is generally assumed to entrain sediment which is then moved by a steady current. Beach sediment transport is also affected by local bed slope and the presence of RIPPLES and other bedforms, and can be modified by human activity such as BEACH NOURISHMENT and coastal engineering structures.

The easiest approach to beach sediment transport is to consider cross-shore and longshore transport separately. Longshore transport is responsible for changes in the beach plan shape, and is usually considered to be unidirectional in the direction of the longshore current. In the simplest formulation, the longshore sediment transport rate is proportional to the longshore wave energy flux at the breakpoint. (See LONGSHORE (LITTORAL) DRIFT.)

Cross-shore sediment transport is responsible for changes in the beach profile. This includes features on the subaerial profile such as beach face slope and berms, and submerged features such as nearshore bars. Net cross-shore sediment transport is difficult to calculate because it occurs as an accumulation of small differences between the large values of onshore and offshore transport, which must be evaluated separately and correctly. Most field data and model predictions indicate that offshore sediment transport dominates under breaking waves due to the seaward-directed undertow. During non-breaking wave conditions, transport is generally onshore-directed due to the effects of incident wave asymmetry.

Governing equations based on fundamental physics have not yet been established, and no unified theory of sediment transport presently exists that is valid for all water depths and fluid motions in the nearshore. Instead, there are many sediment transport models, ranging from quasi-steady formulas such as the energetics approach described below to complex numerical models involving higher-order turbulence closure schemes that attempt to resolve the flow field at small scales. Models can be classified by direction (cross-shore or longshore), driving force (e.g. bottom fluid velocity, bed shear stress, wave energy dissipation, stream power, etc.), or mode of transport (bedload, suspended load, total load). Reviews of sediment transport models are given by Schoones and Theron (1995), Bayram *et al.* (2001) and Davies *et al.* (2002), and measurement of coastal sediment transport is reviewed by White (1998).

One of the preferred approaches to modelling both longshore and cross-shore wave and current-induced sediment transport is based on the energetics approach of Bagnold (1963), formulated for a time-varying flow field (e.g. Bailard 1981). The energetics approach assumes that sediment is mobilized by the oscillatory flow under waves and is related to some power of the instantaneous velocity. Once mobilized, sediment can be transported by a number of different mechanisms: time-averaged flows (undertow or longshore currents), asymmetric orbital velocities and gravity in the downslope direction. The fluid forces which drive the energetics model are based on the calculation of various moments of the fluid velocity, which give the direction and magnitude of both oscillatory and mean flows. Use of the energetics model requires knowledge of the moments of the instantaneous flow field, often decomposed into mean, gravity and infragravity band components and then time-averaged. Net transport is calculated from the integral of the instantaneous rate through a particular time interval.

The energetics model is regarded as one of the best theoretical models available at the moment for time-dependent nearshore sediment transport because of its capacity to represent a wide variety of transport conditions in a computationally efficient manner. However, it does not include a number of factors which are believed to be important in beach sediment transport, such as turbulence, fluid accelerations, threshold of motion, and transport over bedforms. In particular, the energetics model may not be appropriate for use in the swash zone, where beach accretion is most likely to occur. Additional processes are likely to be important in swash sediment transport, such as infiltration/exfiltration, bore-generated turbulence, water depth, inertial forces on coarse grains, and sediment advection and convection.

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SEE ALSO: bar, coastal; beach; beach–dune interaction; beach nourishment; bedload; current; longshore (littoral) drift; sediment budget; sediment transport; sheet erosion, sheet flow, sheet wash; suspended load; wave

DIANE HORN

BEDFORM

The transport of sand or gravel as BEDLOAD (and see SEDIMENT LOAD AND YIELD) creates on the bed a variety of features the size, form and relative orientation of which depend through complex interactions on the density, shape and coarseness of the sediment particles, and on the strength, uniformity and steadiness of the current. These features are bedforms. They participate in the sediment transport and are normally very small or small in height compared to flow thickness. Although the main kinds are, generally speaking, agreed upon, there is little uniformity as to their nomenclature. Bedforms are known from rivers, tidal systems (especially estuaries) and the deep sea, and are most widely familiar from deserts (see DUNE, AEOLIAN). They contribute significantly to contemporary landscapes and seascapes and, where preserved in Quaternary sediments, are valuable indicators of environment and sediment transport strength and direction. Bedforms and their internal structures can be used to establish sediment-transport directions not only on a regional scale but also locally, where changes in strength and direction have occurred on small geographical and stratigraphical scales. For the river and irrigation engineer, bedforms are among the most important determinants of channel hydraulic ROUGHNESS and resistance to flow.

Extensive field and laboratory studies show that river bedforms and their distinctive patterns of internal stratification can be placed in a number of fields defined more or less closely by grain size and flow strength (Figure 12). At flow strengths below the entrainment threshold (see INITIATION OF MOTION) there is neither sediment transport nor bedforms. As flow strength rises, the bedforms first to appear are current ripples (medium- and finer-grained quartz-density sands) and lower-stage plane beds (coarser sands, granules, gravel). At equilibrium, current ripples are ridges of linguoid plan about 0.02 m high and 0.1–0.2 m in wavelength, which increases with sediment size (see RIPPLE). They move very slowly beneath the current as grains are eroded from the long upstream slope and then deposited by settling and avalanching on the steep leeward face (see REPOSE, ANGLE OF) overlain by a turbulent, recirculating vortex. An internal pattern of cross-lamination records the successive positions of the migrating downstream face. Lower-stage plane beds are underlain by subhorizontal parallel laminae on a millimetre to coarser scale. In sediments coarser than medium-grained sand, but at increasingly large flow strengths for progressively finer grades, lower-stage plane beds and current ripples are replaced by dunes. Like current ripples, these forms are transverse ridges which migrate by the erosion of particles from the upstream side followed by their deposition on the downstream face after settling through and avalanching beneath a recirculating, leeward vortex (see REPOSE, ANGLE OF). Patterns of cross-bedding occur internally, the foresets dipping in the sediment-transport direction. Unlike current ripples, dunes scale on flow depth, varying in height from about one-twentieth to about one-fifth of the depth. In large rivers, such as the Mississippi and Brahmaputra, they are several metres high and one to two hundred metres in wavelength. Extensive fields of even larger dunes, composed of cobble and boulder gravel, have been created by some major catastrophic floods (see DAM; ICE DAM, GLACIER DAM; OUTBURST FLOOD). At large enough flow strengths, current ripples and dunes become increasingly round-crested and flat, and are replaced by upper-stage plane beds over which sediment transport, in the form of very low bed waves, is intense. Internally, forms of subhorizontal laminae and bedding occur within such beds. In the case of sands, the surfaces of the laminae carry faint flow-parallel ridges, called parting or primary current

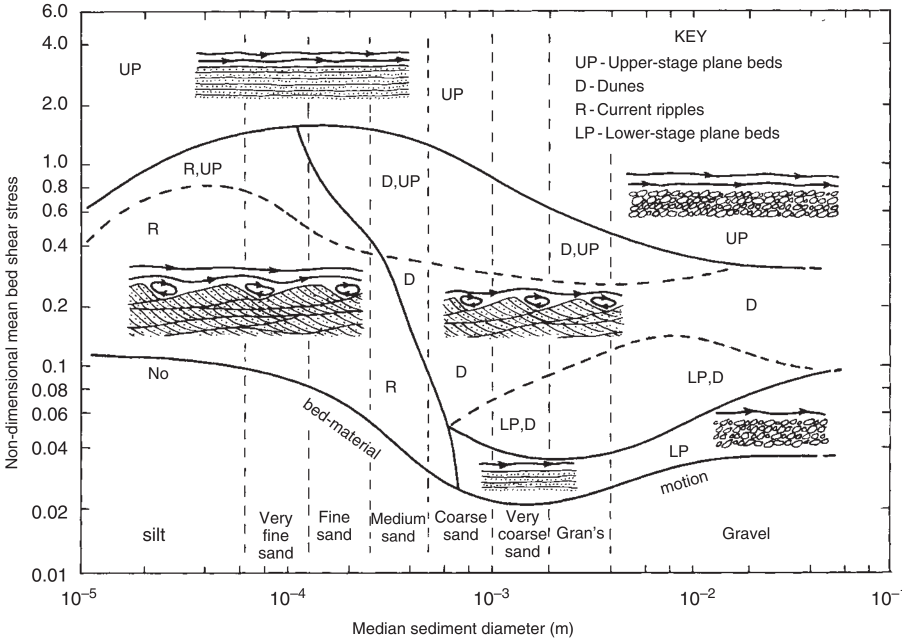

Figure 12 Existence fields, defined by flow strength and median grain diameter, for bedforms and their internal sedimentary structure as shaped by unidirectional water currents in quartz-density sediments. The diagram is based on many hundreds of individual observations. Flow strength is given in a non-dimensional form, defined as the quotient of the mean bed shear stress and the product of the relative particle density, acceleration due to gravity and median particle diameter

lineations, which may be related to flow patterns within the grain-dense lower part of the turbulent BOUNDARY LAYER.

These bedforms are restricted to subcritical flows (see ANTIDUNE), marked by comparatively smooth water surfaces and low values of flow velocity compared to flow depth. Supercritical flows over loose beds consist of unstable, transverse, surface waves more or less in phase with similar waves on the bed, but of a lower height. These are antidune bedforms. At high levels of supercriticality, various very flat bed waves arise, which include rhomboid ripples and dunes related to surface shocks. As supercriticality is promoted by shallow depths, supercritical bedforms can arise at almost any flow strength capable of sediment transport.

Tidal currents, reversing and rotating on a semi-diurnal or diurnal scale, are as non-uniform as river flows, but hugely more unsteady, and this factor complicates the shape, internal structure and relationship to flow conditions of the bedforms encountered in estuaries (see ESTUARY) and shallow seas. Additional kinds of bedform are recorded from these environments, as well as those familiar from rivers. Current ripples, sand and gravel dunes, upper-stage plane beds and antidunes are chiefly restricted to the shallower channels and intertidal shoals of estuaries. As an expression of the unsteady conditions, drapes of mud deposited when the water is slack may accumulate in the troughs of the ripples and dunes, and later become preserved within the cross-stratified interiors of the forms. Large bedforms – sand ribbons and sand waves – are found below tide level in the deeper channels and on tide-swept floors of confined seas such as the English Channel, the Southern Bight of the North Sea and Cook Inlet, Alaska. Subtidal sand waves were discovered in the 1920s and 1930s as the result of detailed hydrographic surveys and the appearance of practical echo-sounders. A few decades elapsed before the development of side-scan sonar allowed sand ribbons to be recognized.

Sand ribbons are long, flow-parallel belts of ripple- or dune-covered sand or fine gravel of low relief with a spacing across the current of a few to several times the flow depth. They express bedload transport under conditions of restricted sediment supply. Sand waves are series of long-crested ridges of sediment arranged transversely to the stronger of the tidal currents. The largest, found in the deepest waters, measure 5 m or more in height and a few hundred metres in spacing. They assume a roughly symmetrical, trochoidal profile where flood and ebb tidal streams are comparably strong, the small- to medium-sized dunes on their backs reversing in direction of travel with each change in tidal phase. Sand waves become increasingly asymmetrical in profile as the ebb and flood tidal streams become more unequal in their ability to transport sediment, and the dunes they carry may migrate only in the direction of the stronger flow, although being slightly rounded by the weaker stream. The internal structure of sand waves is not well known but is certainly complex, reflecting the presence of superimposed dunes which may change and reverse their direction of movement as the tidal streams reverse and rotate. Internally, the more symmetrical waves seem to consist of comparatively thin cross-bedded units recording sediment transport in many different but largely opposed directions. The more asymmetrical ones reveal internally a 'master-bedding' that dips gently in the direction of the stronger tidal stream and between which appears to lie cross-bedding facing chiefly in that same direction.

Currents strong enough to transport sand-sized particles in places affect large areas of the ocean floor and the deeper parts of open continental shelves. These variable but essentially unidirectional flows are not of tidal origin but depend on various thermohaline effects. Sand ribbons and transverse structures which have been called sand waves (very large dunes) have been described from many of the places where these currents operate, such as the long and intricate narrows between the Baltic Sea and the North Sea, the ocean floor immediately west of Gibraltar, swept by the Mediterranean Undercurrent, the continental shelf of south-east Africa, affected by the Agulhas Current, and the level tops of several oceanic guyots. The sediments involved are of diverse origins. They range from terrigenous sands, in some cases reworked after being introduced from shallower depths by TURBIDITY CURRENTs, to bioclastic debris (chiefly shells or foraminifera) eroded from adjacent parts of the ocean floor. Other than their location, general form and link with strong currents, little is known or understood about these deep-sea bedforms.

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J.R.L. ALLEN

BEDLOAD

By mode of transport, the sediment load is divided into SUSPENDED LOAD and bedload. The bedload typically consists of coarse particles derived from the bed material. The immersed weight of these particles is supported by a combination of fluid forces and solid reactive forces exerted at intermittent or continuous contacts with the bed (Abbott and Francis 1977). Bedload is dispersed in a zone immediately above the bed surface and is transported in the rolling/sliding or saltating modes. Particles comprising the bedload continually move in and out of storage on the bed. The pattern of particle motion can be characterized as a series of relatively short steps of random length, each of which is followed by a rest period of random duration, and each particle spends a negligible time in motion compared to the time spent at rest. Consequently, the virtual velocity of bedload is much lower than the flow velocity; in water, for example, it is only of the order of metres per hour, compared to the flow velocity which may be of the order of metres per second (Haschenburger and Church 1998).

In wind, most sand particles move by SALTATION. The saltating particles interact with the bed surface and disturb stationary grains (Anderson and Haff 1991). This not only reduces the threshold of particle motion, it also promotes reptation (the movement of particles impacted by saltating grains over short distances) and surface creep. Within the saltation layer most grains move within 1 to 2 cm of the sand surface. The size of the mobile grains decreases and there is an exponential decline in sediment and mass flux concentration with height (Anderson and Hallet 1986). Accurate data on sediment fluxes are difficult to obtain (Greeley *et al.* 1996; Iversen and Rasmussen 1999), and sand transport equations are frequently used to predict aeolian sand transport rates (Sarre 1989). Above the threshold for sand movement, sand transport rates are commonly assumed to be proportional to the wind (shear) velocity. Aeolian sand transport in bulk is associated with the formation of BEDFORMS of varying size, that develop as regularly repeating patterns (Anderson 1987; Lancaster 1988), and field observation suggests there is close agreement between measured and simulated patterns of erosion and deposition on dunes and wind velocity and direction (Howard *et al.* 1978).

In water, the maximum size of sediment that can be moved by a given flow condition defines flow competence, but the size and amount of sediment moved as bedload is constrained by a river's transport capacity. Transport rates may also be limited by sediment availability. Continuity of bedload transport typically is not maintained along a river, because transport capacity usually does not match the sediment supply. This may promote scour or fill of the river bed and other adjustments to channel geometry. For this reason, although bedload typically constitutes only a few per cent of the total sediment load of most rivers, bedload transport is a very significant process as it governs virtually all aspects of morphological change in river channels. Downstream through a drainage basin, the bed material generally becomes finer through the action of sorting and abrasion; in consequence, the suspended load increasingly dominates over the bedload.

At the lower limit of active transport, where rolling is the dominant transport mode, bed pocket geometry determines which particle sizes are mobile (Andrews 1994). When conditions are below the threshold for general bed motion or the sediment supply is limited the bedload transport rate is moderated by the interaction of coarse and fine size fractions in the bed material, as well as by the available shear stress (Gomez 1995). ARMOURING compensates for the intrinsically lower mobility of coarse particles relative to that of fine grains and renders all particle sizes on the bed surface equally mobile (Parker and Klingeman 1982). Equal mobility arises as a consequence of the shielding of small grains from the flow and the preferential exposure of large particles, coupled with their relative abundance on the bed surface. The adjustments combine to counteract the absolute size effect of particle weight by making coarse particles more available to the flow, and enhancing their probability for entrainment. There are two facets to equal mobility. Equal entrainment mobility is defined as the case when all particle sizes comprising the bed material begin to move at the same flow strength. Equal transport mobility refers to the situation where all particle sizes are transported according to their relative proportions in the bed material, so that the bedload and bed material grain-size distributions are identical. Departures from these conditions give rise to differences in the transport rate of individual size fractions (Wilcock and McArdell 1993), and to hydraulic sorting. Hydraulic sorting is known to occur during the entrainment, transport and deposition of heterogeneous bedload; it is important because of its links to channel armouring and downstream fining (Paola *et al.* 1992).

In most rivers, bedload transport is highly variable in time and space. Temporal variability in bedload transport rates, which is independent of variations in flow conditions, arises from three main sources (Gomez *et al.* 1989). First, variations may result from long- to intermediate-term changes in the rate at which sediment is supplied to or distributed within a channel or reach. Second, short-term, often quasi-cyclic, variations in bedload-transport rates may occur in response to the temporary exhaustion of the supply of transportable material, to the migration of BEDFORMS or groups of particles, or to processes such as ARMOURING. Third, instantaneous fluctuations in bedload transport rates result from the inherently stochastic nature of the physical processes that govern the entrainment and transport of bedload. Spatial variability in bedload transport rates results from downstream and cross-channel changes in the transport field, that occur primarily in response to differences in the shear stress and to changes in the local relation between boundary shear stress and sediment transport (Dietrich and Whiting 1989).

Commonly utilized approaches for gaining knowledge of the bedload transfer through a river reach involve field sampling or measurement, and the application of a formula. Sampling involves the collection of discrete quantities of bedload at various points across a channel, over limited time intervals. Measurement involves the continuous or time-integrated monitoring of bedload over the entire cross-section or reach. The presence of a sampling device on the river bed necessarily alters the pattern of the flow and sediment transport in its vicinity. Thus, bedload samplers must be calibrated to determine their efficiency under different hydraulic and sediment transport conditions. Determining the hydraulic efficiency of a bedload sampler has proved to be a relatively simple task, but determining the sampling efficiency is considerably more complex. Consequently, the sampler calibration process remains incomplete because none of the tests performed to date on any sampling device has provided definitive results (Thomas and Lewis 1993). Since bedload transport rates vary across channel and with time, appropriate temporal and spatial sampling strategies also are required to minimize sampling errors, which decrease as the number of samples collected increases and the number of traverses of the channel over which the samples are collected increases (Gomez and Troutman 1997).

Measurements are usually regarded as exact, and most commonly are obtained using a pit trap. Traps also have a distinct advantage over samplers in as much as, if the trap spans the entire width of the river, it is not only possible to catch all the bedload that passes through the measuring section in a given period of time but also to continuously measure the rate at which sediment accumulates. The simplest traps consist of lined pits or slots in the streambed in which the bedload collects over one or more events (Church *et al.* 1991). More sophisticated traps continuously weigh the mass of sediment (Reid *et al.* 1980).

Bedload formulae equate the rate at which bedload is transported with a specific set of hydraulic and sedimentological variables, and predict bedload transport capacity under given flow conditions. The underlying physics appear quite straightforward (Bagnold 1966), but the conditions governing fluvial bedload transport are complex and there is little consensus about the fundamental hydraulic and sedimentological quantities involved. Consequently, there are numerous bedload transport formulae (Gomez and Church 1989), and none has been universally accepted or recognized as being especially appropriate for practical application.

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BASIL GOMEZ

BEDROCK CHANNEL

Bedrock channels are those with frequent exposures of bedrock in their bed and banks. More precisely, these channels lack a coherent cover of alluvial sediments, even at low flow, although a thin and patchy alluvial cover may be present. However, short-term pulses of rapid sediment delivery may produce temporary sediment fills (see SEDIMENT ROUTING; SEDIMENT WAVE). Bedrock channels exist only where transport capacity exceeds sediment supply over the long term. Contrary to classical definitions, bedrock channels are self-formed. Bed and banks are not composed of transportable sediment, but are erodible. Flow, sediment flux and base-level conditions (see BASE LEVEL) dictate self-adjusted combinations of channel gradient, width and bed morphology.

Bedrock channels are important because (1) they set much of the RELIEF structure of unglaciated mountain ranges, and (2) the controls on their incision rates largely dictate the relationships among climate, lithology, tectonics and topography. Moreover, because river incision rate sets the boundary condition for hillslope EROSION, regional DENUDATION rates and patterns are dictated by the bedrock river network (see HILLSLOPE-CHANNEL COUPLING). Finally, bedrock rivers transmit signals of base level (tectonic/eustatic (see EUSTASY)) and climate change through landscape, and therefore set the timescale of response to perturbation.

Similar to alluvial channels, the longitudinal profiles of bedrock channels are typically smoothly concave-up (see LONG PROFILE, RIVER). These profiles are well described by Flint's law relating local channel gradient (S) to upstream drainage area (A):

$$S = k_s A^{-\theta} \tag{1}$$

where k_s is the steepness index and θ the concavity index. Steepness index is known to be a function of rock uplift rate, lithology and climate (see GRADE, CONCEPT OF). The concavity index is typically in the range 0.4–0.6, and is apparently insensitive to differences in uplift rate, lithology and climate where these are uniform within a DRAINAGE BASIN. However, θ does vary beyond this typical range, usually where downstream fining is particularly strong, or where lithology or uplift rate vary systematically downstream.

Bedrock channel width also varies with drainage area in a manner similar to that observed in alluvial channels (see HYDRAULIC GEOMETRY):

$$W \propto A^{0.3-0.4} \quad (2)$$

Bed morphology also appears to be dynamically adjusted to hydraulic and sediment-flux conditions, and in bedrock-dominated reaches includes STEP-POOL SYSTEMs, plane bed and incised inner channel forms. Discrete KNICKPOINTs and erosional forms such as flutes, POT-HOLEs, longitudinal grooves and undulating canyon walls are common.

Processes of erosion in bedrock channels include plucking, macro-abrasion, wear, chemical and physical WEATHERING, and possibly CAVITATION (see CORROSION; FROST AND FROST WEATHERING). These processes all include critical thresholds (see THRESHOLD, GEOMORPHIC) and most work is probably done by large storms (see INITIATION OF MOTION; MAGNITUDE–FREQUENCY CONCEPT). The relative roles of extraction of joint blocks (plucking plus macro-abrasion) and incremental wear are debated, but appear to depend on properties of the substrate lithology and flow conditions (see ROCK MASS STRENGTH). The relative contributions of BEDLOAD and SUSPENDED LOAD to ABRASION (macro- and wear) are also debated, but most agree sediment flux plays a dual role: providing tools for abrasion, but protecting the bed when overly abundant. The exact nature of the dependence of incision rate on sediment flux and grain size, and the different mechanics of plucking, macro-abrasion and wear all have far-reaching consequences for the relations among climate, tectonics and topography. Both DEBRIS FLOWs and FLOODs likely contribute to bedrock channel erosion in mountainous areas. Their relative importance is not well known, but apparently depends on position in the landscape and setting (tectonics, lithology and climate).

Rates of incision of bedrock rivers are highly variable (from $\mathrm{m\,Ma^{-1}}$ to $\mathrm{cm\,yr^{-1}}$), and depend primarily on tectonic setting and other controls on base-level fall. Where they have both been measured, long-term bedrock river incision rates match the highest rock uplift rates. Burbank *et al.* (1996) estimated incision rates up to $12\,\mathrm{mm\,yr^{-1}}$ on the basis of cosmogenic exposure ages of strath terraces along the Indus River in north-west Pakistan (see COSMOGENIC DATING; TERRACE, RIVER). Short-term incision rates up to $10\,\mathrm{cm\,yr^{-1}}$ have been measured under extreme circumstances.

Incision rates are positively correlated with channel gradient and drainage area, and are often modelled as a function of bed shear stress. The best known, semi-successful, bedrock river incision model is the shear stress or unit STREAM POWER model:

$$E = KA^mS^n \quad (3)$$

where E denotes vertical incision rate (L/T), A upstream drainage area (L^2), S channel gradient, K is a dimensional coefficient of erosion (L^{1-2m}/T) (see EROSIVITY), and m and n are positive constants that depend on erosion process, channel hydraulics and basin hydrology. Although this simple model has been useful for exploration of interactions among erosion, topography, climate and tectonics, much uncertainty remains regarding the physical controls on the model parameters K, m and n. In addition, equation (3) neglects an incision threshold and therefore misses an important aspect of the physics. Further field and laboratory studies are needed to resolve important outstanding issues.

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SEE ALSO: channel, alluvial; dynamic equilibrium; palaeoflood; tectonic geomorphology; valley

KELIN X. WHIPPLE

BEHEADED VALLEY

Fluvial valleys running across an active strike-slip fault react to lateral movement along the fault in the way that their lower reaches, located downstream from the fault, become horizontally displaced in relation to the upper reaches, situated upstream from the fault. In this way the continuity of the valley is lost and the lower reach becomes beheaded. Streams may deflect at the fault and follow the fault zone until they turn into the displaced lower reach, or abandon the original valley and continue without deflection. In the latter case the beheaded section of a valley becomes dry. It is usually only small, narrow valleys occupied by minor streams which become beheaded. For larger rivers, floodplains are normally sufficiently wide to retain spatial continuity.

If a beheaded valley can be clearly defined in the field and contains an alluvial suite which can be dated, then slip rate along the fault can be determined.

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SEE ALSO: seismotectonic geomorphology; tectonic geomorphology

PIOTR MIGOŃ

BERGSCHRUND

Deep, transverse or extensional crevasses that occur at the heads of valley or cirque glaciers (see CIRQUE, GLACIAL) are called bergschrunds. They differ from randklufts in that they occur in glacier ice rather than between the ice and the bedrock headwall. The randkluft of a glacier exists due to a combination of preferential ablation adjacent to warm rock surfaces and ice movement away from the rock wall. Both types of crevasse form formidable barriers for climbers in glacierized mountainous terrain and are particularly dangerous when covered in snow. Although numerous studies have suggested that the bergschrund of a glacier separates immobile, cold based ice at the head of a glacier from the active, sliding ice lower down, Mair and Kuhn (1994) have demonstrated that ice was sliding both above and below the position of a bergschrund in a glacier in the Austrian Alps. Early work on bergschrunds suggested that they were the location of intense FREEZE–THAW CYCLE activity and were, therefore, crucial to the excavation of cirques. Several problems with this hypothesis became apparent once bergschrunds were visited more frequently, most notably by W.R.B. Battle. Essentially, the base of bergschrunds, where bedrock is only occasionally encountered, do not experience appreciable freeze–thaw cycles. It is now accepted that the most effective conditions for freeze–thaw activity lie in the randkluft rather than in the bergschrund of a glacier (Gardner 1987).

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DAVID J.A. EVANS

BIOGEOMORPHOLOGY

Biogeomorphology is sometimes used as an umbrella term to describe studies which focus on the linkages between ecology and geomorphology. Because biogeomorphology deals with the interface between two disciplines it is necessarily diverse, interdisciplinary and hard to define in detail. Biogeomorphological studies have a long history, with several nineteenth-century workers focusing on the interrelationships between communities and landscapes at a range of scales, although the term itself was only coined in the late twentieth century (Viles 1990). Amongst nineteenth-century pioneers, Charles Lyell in 1835 noted the importance of the agency of organic beings in causing superficial modifications on the Earth's surface, and Charles Darwin undertook a classic piece of work on the role of earthworms in influencing soils (Darwin 1881). In recent years several volumes of collected papers on biogeomorphology have been published which provide varying pictures of the scope and nature of biogeomorphological research (see, for example, Viles 1988; Thornes 1990; Hupp *et al.* 1995; Viles and Naylor 2002). Papers within these volumes cover a whole range of organism: geomorphology interactions in riparian, hillslope and coastal settings in environments ranging from arid to humid tropical.

Similar terms have also been used in the literature, including zoogeomorphology or the interrelationship between animals and geomorphology (Butler 1995) and phytogeomorphology (Howard and Mitchell 1985) which investigates the influence of topography on plant communities. Furthermore, geoecology is a commonly used term, especially in the European literature, which also encompasses work addressing interactions between ecology and geomorphology (often at a large scale). Dendrogeomorphology, or the use of tree ring and allied evidence to study geomorphic processes, makes use of the influence of geomorphic processes on plant growth to throw light on the nature and timing of those processes – a neat way of linking ecology and geomorphology from a rather different perspective. Biogeomorphology and these other, similar, umbrella terms reflect a recent trend within the Earth and environmental sciences to investigate links between biotic and abiotic processes (as shown by the flowering of biogeochemistry as a study area, and the growth of interest in Gaia).

Three common themes within present-day biogeomorphological research are the effects of organisms on geomorphic processes, the contribution made by organic processes to the development of landforms, and the impact of geomorphological processes on ecological community development. Many studies have been made in recent years within these themes. For example, in terms of the impact of organisms on geomorphic processes studies have been made of the role of isopods and other fauna in sediment movement in the Negev desert by Yair (1995); the role of *Sabellaria alveolata* reefs in storing coastal sand on the Welsh coast by Naylor and Viles (2000), and the role of plants in influencing splash erosion in Mediterranean matorral environments by Bochet *et al.* (2002). Examples of studies of the role of organic processes in landform development include the study of Fiol *et al.* (1996) which investigates the role of biological weathering in the creation of solutional rillenkarren, and the work of Whitford and Kay (1999) on the role of mammal bioturbation in the production of long-lived mound structures (often called mima mounds). Investigations of the influence of geomorphic processes on ecosystems have been carried out by many ecologists and geomorphologists, such as Scatena and Lugo (1995) in subtropical forests and Hayden *et al.* (1995) on coastal barrier islands. Overall, research into these three major biogeomorphological themes is characterized to date by being largely empirical, field based and focused on a limited range of interactions. There are clear links between the three themes, as for example mammal burrowing produces mima mounds which then influence subsequent vegetation patterning.

Geomorphology and ecology are linked in detail in a range of different ways and understanding and measuring these links has provided much work for biogeomorphologists. Looking at the impact of ecology on geomorphology, organisms can have passive and/or active impacts on geomorphological processes. For example, a micro-organic biofilm can produce chemical weathering of the underlying rock (an active link) whilst also retarding the action of other weathering processes (passively). Biological impacts of geomorphological processes are often referred to by specific terms such as bioerosion, bioweathering, bioturbation, bioconstruction and bioprotection (see Naylor *et al.* 2002 for further details). Considerable research effort has gone into defining these terms and developing ways of studying and quantifying these processes. For example, bioerosion of coastal rocks by a suite of sessile and motile organisms has necessitated measurement of burrow dimensions and calculation of ages of the organisms creating them, as well as quantification of grazing trails through measurement of faecal contents. On the other side of the equation, geomorphology can exert an active and/or passive control on ecosystems. For example, topography influences microclimate which in turn affects plant communities (a passive geomorphological impact) whilst geomorphic processes such as mudflows and rockfalls provide an active control on vegetation development. A whole host of different techniques have been developed to study such influences, often involving mapping and correlation.

All exogenetic geomorphological processes have the potential to be influenced by biological activity; even in some quite hostile environments, as work on subglacial bacterial involvement in chemical weathering has demonstrated (Sharp *et al.* 1999). Indeed, there have been some suggestions that the harsher the environment the more closely interlinked biotic and geomorphic processes are, as organisms extract nutrients, shelter and water from sediments and rocks (Viles 1995). The whole spectrum of biological life forms is involved in biogeomorphological interactions, with animals, plants and a host of micro-organisms all recorded as influencing geomorphic processes. Bacteria have been found to contribute to the precipitation of sinter and travertine in hot spring environments, for example, and tree roots commonly enhance riverbank stability, whilst beaver dams have been recorded as having major impacts on some river networks. As a general rule, micro-organic and plant impacts are more widespread and important to geomorphology than those of animals, which tend to be spatially and temporally patchy in occurrence. Biogeomorphic interactions range greatly in scale and complexity: from the impact of one single organism on rock weathering at the microscale to the involvement of dynamic forest communities in whole catchments. One of the biggest challenges awaiting biogeomorphology in the future is to develop further studies of large-scale ecosystem: geomorphological system interactions over hundreds to thousands of year timescales.

Biogeomorphology is not simply an esoteric scientific backwater dealing with a few bizarre processes (although there are some notably weird examples of biogeomorphic studies such as the work of Splettstoesser in 1985 which discusses the role of rockhopper penguin (*Eudyptes crestatus*) feet in sandstone weathering); it has many applications. Identification of current biological:geomorphological linkages can help geologists interpret unusual sedimentary structures. Recognition of distinctive signatures of biogenic contributions to geomorphic processes on Earth can help scientists search for evidence of former life on other planets such as Mars. More practically, environmental engineering can harness the protective role of organisms in many environments to retard the action of some geomorphological processes. For example, stabilization of coastal dunes through revegetation is an essentially biogeomorphological project. At the smaller scale, understanding links between biofilms and rock surface weathering can aid the conservation of cultural heritage through reducing the threat of biodeterioration.

The future development of biogeomorphology depends both upon its capacity to answer fundamental questions and its ability to provide practical solutions to environmental problems. In some areas, such as the riparian environment, biogeomorphological studies are blossoming and providing much practical information on the mechanical role of roots, the influence of fluvial processes on seed banks and the biochemical role of riparian vegetation. In other areas, biogeomorphological studies remain more narrowly focused on unusual links between single organisms and one geomorphic process. In order to prosper further biogeomorphological studies need to establish novel research methodologies and techniques to investigate the varied links between the biotic and geomorphic worlds, many of which have proved quite hard to quantify and monitor. Furthermore, biogeomorphic studies need to move away from simple empirical, short-term studies to looking at longer term and larger scale situations. For this, numerical modelling may provide a way forward. Also, biogeomorphic studies must try and encompass the evolving two-way interplay between geomorphic and ecological processes, rather than simply focus on the impact of organisms on geomorphic processes or the influence of geomorphology on ecosystem development. Finally, biogeomorphological studies need to continue and expand their essential bridging role – by considering the links between a whole host of organic and inorganic processes in a wide range of environments within a broadly defined Earth surface systems science. The term biogeomorphology is far less important than the scientific terrain it describes – one part of the fertile, dynamic, boundary between the inorganic and organic worlds.

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HEATHER A. VILES

BIOKARST

Biokarst refers to karst landforms created, or influenced to a significant degree, by biological processes. In turn, the processes involved in the formation of such landforms are often called biokarstic. Biokarst features can be erosional or depositional, or involve a combination of the two processes, and are commonly found on exposed limestone surfaces in a range of environmental settings. An early paper by Jones (1965) described many of the erosional features found on limestone pavements as being at least partly biokarstic in origin. Some TUFAS AND TRAVERTINEs are largely influenced by biological processes and thus can be seen to be biokarstic, as can some organically influenced cave deposits. Most landforms recognized as biokarst are quite small (maximum of tens of metres in dimensions), but there is an indirect biokarstic element to most karst landscapes as organisms play a key influence on soil acidity and CO_2 levels which in turn are a vital control of karst development.

Similar terms in the literature include phytokarst, which is more narrowly defined as karst landforms produced by the action of plants, and zookarst, which refers to features produced by animal action. Both phytokarst and zookarst are subsumed within biokarst which can be produced by animal, plant or micro-organism action (and commonly involves a combination of organisms). The classic phytokarst landscape is that described by Folk *et al.* (1973) at Hell, Grand Cayman Island. Here, a series of limestone pinnacles in a low-lying swampy environment have been blackened and dissected in a random spongework pattern which Folk *et al.* ascribe to the action of cyanobacteria (blue-green algae). Another commonly identified type of phytokarst are the light-oriented erosional pinnacles found in the lit zone of many cave entrances (as reported by Bull and Laverty in 1982 in Mulu, Borneo, for example, and sometimes given the alternative name of photokarren). Other phytokarst features are the root holes produced in many limestone surfaces. Zookarstic features are rather rare and localized, but include small-scale erosional relief produced by rock wallaby urine in parts of Australia, and grooves produced by the giant tortoise (*Geochelone gigantea*) on Aldabra Atoll, Indian Ocean. By far the most important group of organisms contributing to biokarstic processes are micro-organisms and lower plants, which in mixed biofilm communities coat most subaerial limestone surfaces in a wide range of environments. Such biofilms play a range of active and passive roles in geochemical transformations, aiding both solution and re-precipitation of calcite.

Biokarst has been recorded from most karst areas, with many studies emanating from the great Chinese karst landscapes. Spectacular biologically influenced erosional relief is also found on many coastal limestone platforms, where bioerosion by a range of organisms produces a complex coastal biokarst. Although karst scientists have often been quick to note biokarst features, it has proved difficult to provide convincing process-form links in order to identify the exact nature and importance of biological influences. The major reason for this difficulty is the multi-factorial nature of karst development, which makes it impossible to untangle the interaction of interlinked processes and emerging forms. Some progress has been made with experimental studies, for example the work of Fiol *et al.* (1996) on the influence of rock surface micro-organism communities in rillenkarren development and the work of Moses and Smith (1993) on the role of physical weathering by lichens on kamenitza evolution.

The small-scale nature of many biokarstic features and the difficulties of positively ascribing their genesis to specific biological processes has made several karst scientists doubt their importance either to karst landscape development or as diagnostic landforms. The most important goal for future work is to provide a more general explanation of the role of a whole range of organisms and biological processes in karst landscape development as a whole, rather than worry whether any individual landform can be defined as biokarstic.

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HEATHER A. VILES

BLIND VALLEY

This is a valley formed by fluvial processes that terminates downstream against a steep and sometimes precipitous slope at the foot of which the stream that carved the valley disappears underground into a cave system. The headwaters are usually on relatively impervious rocks such as sandstones or granites and the surface stream disappears underground when it crosses a lithological contact onto a KARST rock such as limestone. The larger the stream, the further it penetrates into the karst before sinking underground, and hence the longer the associated blind valley. In the early stages of development of blind valleys, the downstream wall is not very steep or high, so if the capacity of the stream-sink (swallow hole, ponore) is exceeded during flood the excess water will overflow downstream along its former course, which is usually dry and abandoned. Such cases are referred to as semi-blind valleys. The incision of the blind valley is controlled by the rate of lowering of the cave system into which it drains. This can proceed in stages as the cave stream breaks through to lower levels. Incision is propagated upstream into the blind valley and results in stream terraces. Thus terraces are often found in blind valleys that grade to the position of a former stream-sink in the terminal face high above the modern swallow hole. Over 10^4 to 10^5 years blind valley incision can attain tens to hundreds of metres.

PAUL W. WILLIAMS

BLOCKFIELD AND BLOCKSTREAM

The term blockfield (or block field) is used to describe an extensive cover of coarse rubble on flat or gently sloping terrain, with an absence of fine material at the ground surface. The German term *felsenmeer* ('stone sea') is sometimes used to describe the same phenomenon. Three types of blockfield are recognized: autochthonous blockfields, formed *in situ* by WEATHERING of the underlying bedrock; para-autochthonous blockfields, in which boulders produced by weathering of bedrock have undergone downslope mass movement over low gradients; and allochthonous blockfields derived from GLACIAL DEPOSITION by upfreezing of boulders and washing out of fine sediments. Blockstreams (or block streams) are covers of coarse debris that have accumulated by mass movement on valley floors.

Most blockfields and blockstreams occur in areas of present or former periglacial conditions (see PERIGLACIAL GEOMORPHOLOGY), particularly in arctic environments and on mid-latitude mountains that lay in the periglacial zone outside the limits of the last Pleistocene ice sheets. Blockfields are particularly widespread on mid- and high-latitude plateaux such as those of Scandinavia and Scotland.

Blockfields and blockstreams occur on a wide range of rock types, but are particularly common on well-jointed igneous and metamorphic rocks that have weathered to produce abundant boulders but only limited amounts of fine sediment. Most blockfields comprise boulders less than 1–2 m in length. In autochthonous blockfields the largest boulders usually occur at the surface and boulder size diminishes with depth. Below the openwork surface layer, blockfields and blockstreams usually contain an infill or matrix of fine sediment (sand, silt and clay), and interstitial organic material has also been recorded. Plateau blockfields tend to be 0.5–4.0 m deep, but blockstreams consisting of accumulated valley-floor boulder deposits reach depths of 10 m or more.

Surface boulders in blockfields may be angular or, more commonly, edge-rounded by GRANULAR DISINTEGRATION. Where downslope mass movement has occurred, elongate boulders often exhibit preferred downslope orientation and upslope imbrication. PATTERNED GROUND may be present in the form of large sorted circles on level ground and sorted stripes on slopes, and blockstreams sometimes support lobate structures indicative of movement by SOLIFLUCTION of underlying fine sediments.

In a perceptive early (1906) account of blockfields and blockstreams on the Falkland Islands, J.G. Anderrson attributed their formation to frost weathering (see FROST AND FROST WEATHERING) of the underlying bedrock, slow downslope movement of the weathered debris by solifluction, and immobilization by eluviation (see ELUVIUM AND ELUVIATION) of fine sediment from the upper layers. Upheaving of boulders and frost-sorting also appear necessary to produce downward fining of the openwork boulder layer and the formation of sorted patterned ground. Although this general model is widely accepted, some researchers have suggested that autochthonous and para-autochthonous blockfields and blockstreams are of polygenetic origin. In particular, it has been proposed that some plateau blockfields evolved from chemically-weathered (see CHEMICAL WEATHERING) REGOLITH mantles, of interglacial or Tertiary age, that were subsequently modified by frost action (e.g. Nesje 1989; Rea *et al.* 1996; Dredge 2000). This view is based on the location of blockfields on Tertiary erosion surfaces, and the presence in the subsurface fine fraction of clay minerals indicative of prolonged chemical weathering. On certain lithologies, however, blockfields have developed on glacially eroded bedrock since the last glacial maximum (Ballantyne 1998) implying formation under periglacial conditions alone within the last 20,000 years. Some blockfields also show evidence for modification by glacier ice or glacial meltwater (Dredge 2000).

Although there is evidence for blockfield formation during the Holocene in arctic permafrost environments (Dredge 1992), mid-latitude blockfields and blockstreams are manifestly relict. Exposed boulder surfaces have been edge-rounded by prolonged granular disintegration and many support a cover of mosses and lichens. The relationship between such relict blockfields and former ICE SHEETs has been vigorously debated. In some areas, such as western Norway and north-west Scotland, the lower limits of autochthonous blockfields descend regularly along former glacier flow-lines and have been interpreted as trimlines (see TRIMLINE, GLACIAL) marking the maximum vertical extent of the last ice sheets in these areas (Nesje 1989; Nesje and Dahl 1990; Ballantyne *et al.* 1998). Elsewhere, however, there is convincing evidence that blockfields survived the last glacial maximum under a cover of cold-based glacier ice that was frozen to the underlying substrate and hence accomplished little or no erosion (Kleman and Borgström 1990; Dredge 2000). Thus not only does the age and evolution of blockfields and blockstreams vary from area to area, but also their significance in relation to the dimensions of former ice sheets is dependent on the thermal regime of these ice masses.

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SEE ALSO: frost and frost weathering; frost heave; mechanical weathering; periglacial geomorphology

COLIN K. BALLANTYNE

BLOWHOLE

Fountains of spray are emitted through blowholes during storms and high tidal periods when large breakers surge into tunnel-like caves connected to the surface. Many blowholes develop along joint (see JOINTING) or fault-controlled shafts, but particularly spectacular examples result from marine invasion of KARST tunnels and sinkholes in limestone regions, and lava tubes or tunnels in volcanic areas. Blowholes are also common on CORAL REEFs, where encrusting coralline algae can enclose spur and groove systems and surge channels running through algal ridges.

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ALAN TRENHAILE

BLUE HOLE

Likened to sapphires set in turquoise, they are submarine, circular, steep-sided holes which occur in coral reefs.

The classic examples come from the Bahamas (Dill 1977), but other instances are known from Belize and the Great Barrier Reef of Australia (Backshall *et al.* 1979). Although volcanicity and meteorite impact have both been proposed as mechanisms of formation, the most favoured view is that they are the product of karstic processes (i.e. they are a DOLINE or CENOTE) which acted at times of low glacial sea levels when the reefs were exposed to subaerial processes. Subsequently they were submerged by the Flandrian Transgression of the Holocene.

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A.S. GOUDIE

BOLSON

Derived from the Spanish word for 'purse', bolsons are depressions with centripetal drainage that are surrounded by hills and mountains (Tight 1905). At their centre there is normally a saline playa or PAN, but if the low-lying area is drained by an ephemeral stream the basin may then be termed a 'semi-bolson' (Tolman 1909). Bolsons are a feature of semi-arid basin-and-range terrain and may contain such landform types as PEDIMENTs, ALLUVIAL FANs and BAJADAs.

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A.S. GOUDIE

BORING ORGANISM

Several life forms have evolved a means of penetrating a variety of substances for security and protection, wood and softer rocks being common subjects of such actions. The geomophological interest is largely focused on the rock-boring organisms because their activity acts as a direct erosional agent and can also weaken the rock, making it more susceptible to erosion by other means. There exist terrestrial boring organisms, mainly algae and the fungal component of lichens, and particular interest has been shown in the marine borers, especially around the intertidal zone where they can lead to the formation of an undercut notch on rocky coasts. Some erosive mechanisms appear to be mechanical but many also appear to be chemical, attacking the more soluble rocks. Thus much of the interest in boring organisms lies in the field of the production of surface textures and smaller scale landforms in terrestrially exposed limestones (Trudgill 1985, Ch. 2, 3, 4, 8; Viles 1988) and in coastal limestone geomorphology where significant features, such as undercut notches of up to a few metres in dimension, can be formed (Trudgill 1985, Ch. 9, 10).

In studies involving environmental reconstruction, the occurrence of fossil intertidal boring organisms either above or below present sea level can provide evidence of former sea levels. This is particularly the case when undercut notches are found on dry land, considerably above present sea level. In some situations, these could have been formed by river action but where there are fossil perforations made by boring organisms, this confirms a marine origin for the undercut as assemblages of boring organisms in hard rock are unusual in fresh-water situations. This is especially the case if fossil boring bivalve shells are still present and the species can be identified and confirmed as marine organisms. In some cases, the shell material can be extracted and used for dating purposes and thus if there is a sequence of raised shorelines, palaeoenvironmental reconstruction is greatly assisted. Rowland and Hopkins (1971) noted that the boring bivalve mollusc *Hiatella arctica* can be found widely in the Arctic and Atlantic oceans and also in the Pacific Ocean from Alaska to Mexico. They noted the potential for the use of its fossil shells in paleoclimatic reconstruction.

Boring algae

Boring algae may be found in the first few millimetres of very many rock surfaces, and indeed the darker colour of rock surfaces which have been exposed for any length of time is often ascribable to this algal layer. The algae are frequently found to be blue-green algae (cyanobacteria). The algae associated with rock surfaces can be described as epiliths which live on the surface or as endoliths which live below it, with a further distinction being made for the chasmoliths which exist in the interstitial spaces between the rock grains – thus only endoliths which penetrate grains, or perforants, can be regarded as borers. The presence of endolithic perforant algae in limestone leads to the formation of very fretted surfaces known as phytokarst (Viles 1988). This can give a ferociously sharp, intricate and spongy rock surface. In cave entrances, the phytokarst is directional and angled to the light source and the borings are a product of erosion by phototrophic algae.

The benefit to the algae is to have access to moisture within the rock; however they still have to photosynthesize so they are found in thin layers just below and parallel to the surface at depths where moisture is present and where light can still penetrate. This optimum depth is termed the light compensation depth or LCD. The access to moisture is especially important in harsh environments and so while endolithic algae are found very widely over the rock surfaces of the Earth, they can occur in extreme environments including Antarctica (Friedmann and Ocampo 1976) where they can be important primary producers.

In the marine environment they commonly dominate in the mid- to upper shore, beyond which (inland) it is too dry and below which (towards the sea) it is wet enough for other organisms also to occur. They provide food for a wide range of rasping molluscs which contribute to rock erosion by ingesting rock with the algae. The algae then penetrate further into the rock to achieve an optimum LCD.

Boring endolithic algae are most usually found in carbonate rocks and the mechanism by which they bore is thought to be one where the algal filaments, about 10 μm wide, release extracellular chelating or acid fluids from the terminal cell. Using a high-powered electron microscope it can be established that up to 50 per cent of a rock surface bored by algae can be void space. Such boring can give rise to an extremely fretted dissected surface.

Boring fungi and lichens

The fungal portions of lichens can penetrate into rocks, again mainly in the carbonate rocks, in both intertidal and terrestrial environments. In both cases they exploit the weakness of calcite crystal interfaces and can also make larger pits.

Boring sponges

Boring sponges, commonly of the species *Cliona* are less able to withstand desiccation than boring algae and thus their distribution is from the mid- to lower intertidal. Extensions of the sponge tissue, termed etching amoebocytes, which, using acid secretions, are able to penetrate calcite in semicircular cuttings about 60–80 μm wide. On the surface, small 'keyhole' slots of 0.5–1 mm long are visible to the naked eye.

Boring bivalves and boring barnacles

Species of bivalve molluscs produce tubular borings which may penetrate into carbonate rock by several centimetres; they may also bore into live coral, sandstone, clay, peat and wood. There is evidence for acid secretion in carbonate substrates but they can also excavate the substrate by mechanical means, combining a rocking or rotational movement, which acts to grind the substrate, and a pumping motion using muscular contractions. In tropical areas the commonest boring bivalve is *Lithophaga* and the commonest boring barnacle is *Lithotrya*. In temperate regions the boring bivalve *Hiatella arctica* is a frequent borer of limestone in low intertidal and subtidal locations (Trudgill and Crabtree 1987). Additionally there also exist boring sipunculid worms and polychaete worms which generally make much thinner borings than the boring bivalve molluscs or boring barnacles.

Boring echinoderms

In the lower intertidal, subtidal and in rock pools several species of boring echinoderms exist, in temperate regions commonly *Paracentrotus lividus* (Trudgill *et al.* 1987) and in tropical regions *Echinometra lucunter* is common. The former make semicircular pits a few centimetres in diameter and the latter effect grooves in the rock surface. It is evident that echinoderms bore for protection, such as on exposed coasts of Carboniferous Limestone in Co. Clare, Eire, *Paracentrotus lividus* bores at rates between 0.25–1.5 cm a^{-1} whereas on sheltered coasts they may exist on unbored surfaces or in shallow depressions with lower excavation rates (Trudgill *et al.* 1987).

Rates of boring

The rates of erosion by boring organisms can be measured in a linear fashion (mm a^{-1}) where there is surface retreat or a single boring, or in a cubic fashion (cm^3 a^{-1}) where there are more diffuse excavations. Published rates of the erosion of limestones by boring organisms (Spencer 1988) include the following:

Echinoderms	0.25–14.0 cm^3 a^{-1}
Sponges	1.0–1.4 cm^3 a^{-1}
Hiatella	5–10 mm a^{-1}
Lithophaga	9–15 mm a^{-1}
Lithotrya	8–9 mm a^{-1}

Given that overall surface retreat ranges from around 0.5 to 4 mm a^{-1} and commonly is 1.0 mm a^{-1}, it can be seen that boring organisms are highly significant erosive organisms. Indeed, where boring organisms are present, this leads to the formation of a horizontal undercut or notch in the coastline, not only through the direct action of the organisms themselves but also through the removal of the mechanically weakened rock by wave action. The mechanical boring of hard rock by the larger shelled organisms produces significant quantities of fine carbonate sediment.

Zonation of boring organisms

The zonation of intertidal organisms is of interest to the geomorphologist if it is proposed that there is a cause and effect relationship between biological zonation and morphological zonation (Trudgill 1987). Originally it was thought that biological zonation was a response solely to the ability of different species to withstand emersion and desiccation. However, theories of intertidal distribution have become less environmentally deterministic and now involve concepts of interspecific competition and predation. Thus species distribution can vary markedly in any one tidal zone according to the presence or absence of predators and other species. This suggests that while there may be a general zonation of boring organisms and hence morphological types produced by them, the distribution of individual boring organisms is liable to vary at different locations in relation to predation and competition rather than just to tidal zonation.

In addition, the variation of the landform itself may provide different micro-habitats which afford protection or sites which are too exposed, leading to local variability and that feedback effects can occur. In particular, on flatter, near-horizontal surfaces boring can produce low-lying areas which facilitate water retention and hence survival, meaning that they then become even deeper as boring activity and number of boring organisms can intensify.

Geomorphological significance

Schneider (1976) suggests that in the Adriatic, moisture conditions provide the limiting factor on boring activities. He sees three stages:

1. The primary depressions in rock surfaces are first colonized since they retain moisture longer than their higher surroundings. Conditions for boring and grazing prevail longest here. As a result each depression becomes the site of more intense biological erosion.
2. The pools enlarge laterally and small depressions coalesce, wet areas are preferentially bioeroded and relief is thus intensified.
3. This is the stage of maximal relief. It shows maximal contrast in ecological conditions and thus in the destructive processes. The water in the depressions is changed at most high tides and thus brings fresh sea-water to organisms. Each pool may enlarge and break into the next.

The sequence may be restarted if a deeper bedding plane or other weakness is reached, thus draining the pool; alternatively the rim of the pool may be breached.

Tidal range itself and the degree of exposure are also important considerations. In areas with limited tidal range, as, say, in the Mediterranean, there is commonly a deep undercut notch formed by boring organisms limited to 15–20 cm in height in the mid-intertidal which in itself may only have an amplitude of some 30 cm. The same ratio of notch to range applies as tidal ranges expand. Where coasts are exposed to larger waves and storms, boring organisms may be present but contribute quantitatively far less to the overall erosion of the coast, mechanical erosion tends to dominate, and, correspondingly, the undercut notch is either weakly developed or absent. Here the coast has the appearance of a sloping ramp rather than of a vertical cliff with recess or undercut notch as tends to be the case in sheltered locations.

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STEVE TRUDGILL

BORNHARDT

Bornhardts are dome-shaped, steep-sided hills, usually built of massive igneous rocks such as granite or ryolite, with bare convex slopes covered with very little talus and flattened summit surface. Bornhardts form due to differential weathering and erosion, in the course of which the surrounding less massive rock is eroded away leaving massive, sparsely jointed compartments. Many bornhardts have probably formed through selective DEEP WEATHERING followed by stripping of the SAPROLITE, but they can also emerge through gradual lowering of the surrounding terrain in the absence of deep weathering. Bornhardts are structure-controlled landforms and occur in every climatic zone; the existence of massive rock compartments is the necessary factor.

Characteristic features of bornhardts are slope-parallel joints called sheeting joints. They tend to be considered as resultant from unloading and would develop at shallow depth and after exposure, although it is also argued that sheeting develops in deeper parts of the crust, in response to compressional stress (Vidal Romani and Twidale 1999). Gradual opening of joints promotes slope instability, therefore rock slides and falls involving large masses of rock are common on bornhardts. Consequently, whereas upper slopes are bare and talus-free, footslopes may be covered by big blocks derived from upslope.

One of the persistent problems in the literature is the distinction between a bornhardt and an INSELBERG. The term 'bornhardt' was used by B. Willis in the 1930s to honour a German explorer from the turn of the nineteenth century, W. Bornhardt, who had introduced the name 'inselberg', but primarily to emphasize a special category of massive, dome-shaped inselbergs. The term subsequently evolved to describe monolithic domes regardless of their degree of isolation in the landscape. Therefore, although the terms are occasionally used as synonyms, these two categories of hills should not be confused. Nor is it justified to restrict bornhardts to granite lithology. Whereas 'inselberg' emphasizes isolation in space, bornhardts need to have distinctive domed shapes. Hence there is only partial overlap between the two and there exist bornhardts which are not inselbergs, and vice versa.

Classic examples of bornhardts include domes in Rio de Janeiro, Half Dome in Yosemite Valley and Ayers Rock in Australia. They are also abundant within African shields.

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PIOTR MIGOŃ

BOULDER PAVEMENT

Striated boulder pavements form on intertidal surfaces affected by floating ice (Hansom 1983; Forbes and Taylor 1994), in ice-affected fluvial environments (Mackay and Mackay 1977) and at the base of glaciers or grounded ice sheets (Eyles 1988). Pavements deposited subglacially are the result of accretion of boulders around an obstacle and carry striations that are largely unidirectional, similar to fluvially derived striations produced by debris-charged floating ice. Striations on intertidal pavements are controlled by the direction of grounding of floating ice together with rotational striations imparted on stranding. Intertidal boulder pavements are composed of smoothed and highly polished boulders, often up to 1 m in diameter, tightly packed together as an undulating mosaic. They are often interrupted by bedrock outcrops together with furrows and polygonal depressions up to 5 m across. The main process seems to be the bulldozing and packing of loose boulders in the intertidal zone of a low-gradient boulder-strewn shore and the abrasion and striation of boulder surfaces by rock-shod floating ice. Prerequisites for their development appear to be a boulder source, frequent onshore movement of floating ice and a low-gradient intertidal zone. The degree of development is controlled by the frequency of onshore ice movement, well-formed pavements occurring in environments subject to high frequencies of freely moving ice.

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JIM HANSOM

BOUNDARY LAYER

Emergence of boundary layer theory

The German engineer Ludwig Prandtl (1875–1953) presented a seminal paper to the 1904 Mathematical Congress in Heidelberg entitled 'Fluid Motion with very Small Shear' (Schlichting 1968). Prandtl showed that, with the aid of theoretical considerations and simple experiments, fluid flow over or around a solid body such as a sphere, cylinder or flat plate could be divided into two distinct regions. One region is relatively close to the body (or boundary), is relatively thin, and is characterized by large velocity gradients and viscous shear stresses. That is, fluid friction plays an important role in determining the physical characteristics of the layer. The second region is relatively far away from the boundary, and it is characterized by small velocity gradients and viscous shear stresses. That is, fluid friction may be neglected. This conceptualization termed boundary layer theory, as presented by Prandtl and further expanded by Geoffrey I. Taylor (1886–1975) and Theodor von Kármán (1881–1963), proved to become the foundation for modern fluid mechanics (Schlichting 1968).

A boundary layer can be defined as that part of the flow markedly affected by the presence of the boundary (Middleton and Wilcock 1994). Here, a flow refers to the motion of almost any kind of Newtonian fluid. Most real flows of geomorphic interest, such as flowing water in a river or blowing wind over a sand dune, are considered boundary layers because much of the flow is strongly affected by the boundary.

Laminar and turbulent flow

Osborne Reynolds (1842–1912) was the first to distinguish between two types of flow regime: laminar and turbulent (Schlichting 1968; Tritton 1988). In the laminar regime, the entire flow region appears to be divided into a series of fluid layers, each layer bounded by stream surfaces conforming to the boundary. In the case of two-dimensional flows, the traces of these surfaces on the flow plane are called streamlines (Figure 13). The rate of flow between two adjacent streamlines remains constant, although their spacing and orientation may vary, and velocity at-a-point does not vary or fluctuate in time. The transfer of fluid momentum, which results from the acceleration of slower fluid layers by faster moving layers, occurs at the molecular scale (Schlichting 1968).

In a turbulent flow, fluid particle paths are sinuous, intertwining and disordered. Fluid mixing occurs at both molecular and macroscopic scales. At this larger scale, fluid mixing commonly involves three-dimensional flow structures called eddies or vortices (turbulence), which are hairpin or horseshoe shaped rotating parcels of fluid moving away from or toward a boundary (Smith 1996). Because these vortices are relatively large and energetic, the time and length scales of turbulence are large and hence the turbulent transfer of fluid momentum is

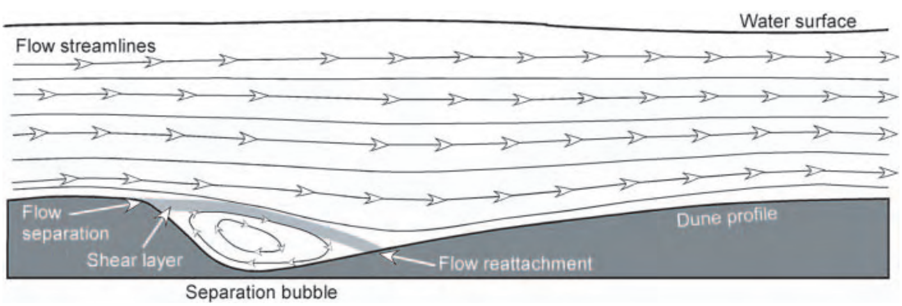

Figure 13 Time-averaged flow over a dune bedform (from Bennett and Best 1995). Flow is from left to right

large compared to that of molecular diffusion. Within a turbulent flow, velocity at-a-point can fluctuate greatly as a result of passing vortices. Turbulence intensity typically is a measure of the magnitude of the velocity fluctuations compared to the time-average value at-a-point.

The boundary Reynolds number

Reynolds found both experimentally and through dimensional analysis that the transition between laminar flow and turbulent flow occurs when the ratio of the inertia fluid forces is significantly larger than the viscous or frictional fluid forces (Tritton 1988). The inertia forces can be defined as the product ρud where ρ is fluid density, u is the mean flow velocity and d is the mean flow depth. The frictional forces can be characterized by the molecular viscosity of the flow, μ. This dimensionless ratio is called the boundary Reynolds number, Re. Flows are considered laminar when $Re < 500$, turbulent when $Re > 2000$, and transitional when $500 < Re < 2000$ (Tritton 1988). When a flow is laminar, any small disturbance to the flow, such as a protruding particle or a small change in bed topography, will not cause a change to the flow path lines or velocity and the disturbance will be damped by viscous forces. When a flow is turbulent, all disturbances to the flow will produce an effect throughout the boundary layer. When a flow is transitional, only select disturbances will affect the flow.

Few natural flows are laminar because most are deep and fast enough for the boundary Reynolds number to be very large. For example, the Mississippi River near its mouth has a Reynolds number near 10^7.

Flow separation and reattachment

As a flow moves along a curved boundary or over an obstacle, the boundary layer may separate and move away from the wall. Separation occurs when the pressure gradient in the downstream direction is adverse or unfavourable (Tritton 1988). A pressure gradient is considered adverse when a flow is expanding, diverging and decelerating in the longitudinal direction, and the pressure acting on the boundary is increasing, such as on the rear part of a streamlined pier. Both laminar and turbulent boundary layers can separate. Laminar flows usually require only a relatively short region of adverse pressure gradient to produce separation, whereas turbulent flows separate less readily.

Reattachment is the opposite of separation. There is a tendency for separation to be followed by reattachment unless the adverse pressure gradient continues long enough to prevent it (Tritton 1988). This separation–reattachment phenomenon is associated with a separation bubble or a region of flow recirculation. A typical example of flow separation can be found downstream of a ripple or a dune (Figure 13; Bennett and Best 1995). A recirculation bubble extends from the ripple or dune crest point to a distance downstream of about 5 to 7 times the bedform height, where flow reattaches to the boundary.

Downstream of the line of separation, there is a region of intense shear between the faster-moving outer part and the slower-moving or counter-rotating inner part of the boundary layer. Consequently, the flow along this shear layer is unstable, and turbulence is produced (Figure 13; Tritton 1988; Middleton and Wilcock 1994). Turbulent wakes, or regions of high turbulence, are typical of flow past any kind of obstruction at a high Reynolds number. The mixing layer present above a ripple or dune just downstream of the bedform brink is dominated by shear layer turbulence.

Structure of turbulent boundary layers

A turbulent boundary layer can be subdivided into three distinct zones: an inner layer, an outer layer and a wake region (Schlichting 1968). The inner region of a turbulent boundary layer is composed of a viscous sublayer (up to $y^+ = yu_*/\nu = 10$, where y is distance from the boundary, u_* is shear velocity, $u_* = (\tau/\rho)^{0.5}$, τ is bed shear stress, and ν is the kinematic viscosity of the flow) and a buffer layer (from $10 < y^+ < 40$). In the viscous sublayer, viscous forces dominate, yet very weak turbulent motions occur. These motions, called viscous sublayer streaks, are longitudinally oriented rotating tubes of alternating high- and low-speed fluid (Smith 1996). In some flows, sand streaks can be observed on the bed surface and these demarcate the location of low-speed streaks that tend to accumulate sand. Such sand streaks have been observed in the rock record and are called parting or current lineations.

The buffer layer is where the turbulent bursting process takes place. Low-speed streaks, in the general shape of a hairpin vortex, are lifted from the boundary and into the buffer region.

This lifted low-speed streak creates a thin shear layer that becomes unstable, oscillates, and is energetically ejected into the outer region of the flow (called a burst or an ejection event). Immediately following an ejection event, high-speed fluid from the outer region rushes in to replace the ejected fluid, impinging the bed (called a sweep event; see Smith 1996). This two-stage phenomenon is called the bursting process and can account for 70 per cent or more of all turbulence production within a boundary layer. Turbulent bursts or ejections are energetic enough to suspend sediment from the bed in river and airflows. Sweep events, with their high instantaneous drag forces, can entrain sediment particles resting on a bed surface.

In the outer region, representing the lower 10 to 20 per cent of the flow depth, there is a region where the velocity distribution varies logarithmically with distance from the bed, thus termed the logarithmic zone. Prandtl first conceptualized this velocity distribution in his 1925 mixing length theory (Schlichting 1968). Prandtl visualized a simple mechanism of fluid motion where parcels of fluid would move upwards or downwards, accelerating or decelerating the surrounding fluid. The distance over which the fluid is mixed is called the mixing length. Von Kármán expanded this theory by assuming that the mixing length varies as a simple function of distance from the bed multiplied by a dimensionless, universal coefficient (von Kármán's coefficient, $\kappa \sim 0.41$; Schlichting 1968). The final result is the Kármán–Prandtl law of the wall, $u/u_* = 1/\kappa \ln(y/y_0)$, where u is the velocity at a distance y from the wall and y_0 is the roughness height where velocity goes to zero. This velocity distribution has been shown applicable to a wide variety of flows such as pipes, rivers, near-shore environments, aeolian environments and in the near-bed region of gravity currents. Common uses of the law of the wall are the determinations of bed shear stress, roughness height and the turbulent mixing characteristics of the flow.

Finally, the velocity distribution in the outer 80 per cent of the turbulent boundary layer deviates from the logarithmic law. Here the velocity distribution is similar in shape to the velocity-defect profile in wakes (law of the wake; Coles 1956). For many straight rivers with relatively flat beds, the law of the wall is applicable over the entire flow depth (Middleton and Wilcock 1994). However, the presence of bedforms will significantly increase roughness length scales and velocity distributions.

Turbulent boundary layers are further qualified based on the roughness of the bed surface. This roughness parameter is called a grain Reynolds number Re_G, and it is defined as $Re_G = u_* k_s/\nu$, where k_s is the equivalent sand roughness height, which is approximately equal to the bed grain size (Schlichting 1968; Bridge and Bennett 1992). If the bed sediment is relatively small or absent, such that the grains are completely immersed in the viscous sublayer, then $Re_G < 11$, the sediment particles are subjected to viscous fluid forces only, and the boundary is considered hydraulically smooth. If the bed sediment is relatively large, such that the grains are larger than the viscous sublayer, then $Re_G > 70$, the sediment particles are subjected to turbulent fluid forces, and the boundary is considered hydraulically rough. Turbulent boundary layers are considered hydraulically transitional if some but not all grains are immersed in the viscous sublayer and $11 < Re_G < 70$. There are slightly different versions of the law of the wall and the determination of the equivalent sand roughness height depending on the roughness of the turbulent boundary layer. In general, beds that are composed of larger grains have relatively higher turbulent intensities and greater flow resistance.

The shape of the oft-used Shields curve for the dimensionless threshold of particle entrainment reflects this effect of grain Reynolds number on boundary layer characteristics (Bridge and Bennett 1992). Very small grains ($Re_G < 10$; hydraulically smooth flows) immersed in the viscous sublayer require higher dimensionless shear stresses for particle entrainment than larger grains that protrude higher in the turbulent boundary layer ($Re_G > 100$; hydraulically rough flows).

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SEAN J. BENNETT

BOUNDING SURFACE

Bounding surfaces represent discontinuities in sedimentation. Surfaces occur within all environments, and form an integral part of the geomorphic landscape and the rock record.

For example, migrating aeolian dunes produce three bedform-scale bounding surfaces. Reactivation surfaces form when the lee faces of dunes are eroded, such as when the dune reverses. Where associated with seasonal winds, these surfaces define cycles within the dune cross-strata. Superposition surfaces occur where smaller dunes migrate over the lee face of the main bedform. The surface is produced by scour associated with the passage of the INTERDUNE troughs of the superimposed dunes. Interdune surfaces begin with deflation of the stoss (windward) slopes of migrating dunes and culminate at the interdune floor.

At the dunefield scale, sequence or super surfaces form when accumulation in the field ceases. Examples include stabilization of the field by vegetation, and deflation to a planar surface defined by the water table.

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SEE ALSO: interdune

GARY KOCUREK

BOWEN'S REACTION SERIES

In the early twentieth century, N.L. Bowen (1928) developed an idealized model, now called Bowen's Reaction Series, to describe the evolution or differentiation of igneous rocks. Recognizing that the types of minerals that form, and the sequence in which they crystallize, depend on the chemical composition of the magma and the temperature and pressure range over which the magma crystallizes, Bowen described two separate reaction sequences at high temperatures that eventually merge into a single series at cooler temperatures (see Figure 14).

The discontinuous series (left-hand side), involves the formation of chemically unique minerals at discrete temperature intervals from iron and magnesium-rich mafic magma. The first rocks to form are composed primarily of the mineral olivine. Continued temperature decreases, and fractionation of the magma (the early formed minerals are removed from the liquid by gravity), change the dominant minerals which form from pyroxene, to amphibole, and then to biotite.

The continuous series (right-hand side), involves the mineral plagioclase feldspar. At high temperatures, these minerals are dominated with calcium. With continued cooling, calcium and aluminum are exchanged for sodium and silicon. The convergence of both series occurs with a continued drop in magma temperature. Crystallizing rocks become richer in potassium and silica. The last mineral to crystallize in the Bowen's Reaction Series is quartz.

Examples of complete igneous sequences from basalt to granite are rare and other mechanisms are now known to produce differentiation sequences. Bowen himself acknowledged that the series was a simplification of very complex reactions and could be misleading if taken at face value. The reaction series also is used to explain susceptibility of minerals to weathering (see GOLDICH WEATHERING SERIES).

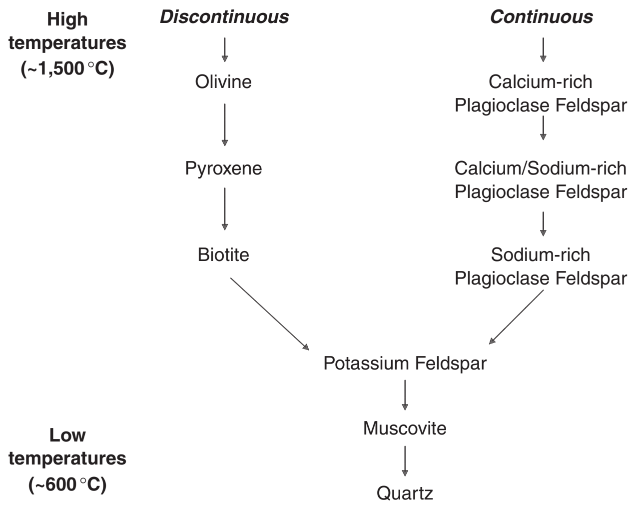

Figure 14 The Bowen's Reaction Series

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SEE ALSO: Goldich weathering series; chemical weathering

CATHERINE SOUCH

BOX VALLEY

A box valley has a broad flat floor, bounded by steep slopes which form a sharp piedmont angle. They are common in periglacial areas (see PERIGLACIAL GEOMORPHOLOGY), where they are formed by rapid lateral migration of braided channels, and by MECHANICAL WEATHERING by an 'ice rind' beneath the floodplain. Box valleys in mid-latitudes have been interpreted as relics of former cold climates. However, they also occur in tropical and arid lands, where they are formed by intense weathering or sheet wash.

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R.W. YOUNG

BRAIDED RIVER

Phenomenology of braiding

The hallmark of braided rivers is the presence of multiple active channels that divide and rejoin to form a pattern of gently curved channel segments separated by exposed bars (Plate 18). Braided rivers are marked equally by temporal dynamism: gradients in sediment flux associated with the complex spatial topography change local slopes, leading the flow to continually adjust its path as

Plate 18 The braided Rakaia River, New Zealand

it picks its way through the network. Even when external conditions are constant, the braided pattern is continually changing, yet statistically consistent: a true dynamic equilibrium.

Braided rivers are known from around the world, but they are most common today at high latitudes. Often, braided rivers are GRAVEL-BED RIVERS, but prominent exceptions include some of the largest braided rivers in the world, such as braided sections of the Huang He and Ganges–Brahmaputra rivers. It has been suggested that braiding was the dominant river pattern on Earth before the first appearance of land plants in late Silurian time (Schumm 1968). Braided patterns have been observed on Mars, and vegetal patterns that resemble braiding are known from some bogs. However, although many morphological features of rivers are reproduced under oceans and lakes by the action of density currents, braiding appears to be rare or absent in the subaqueous realm.

It is worth distinguishing braided rivers from anastomosing channels, the other main type of anabranching channel (see ANABRANCHING AND ANASTOMOSING RIVER). In anastomosing channel networks, the typical width of the channels is much smaller than that of the bars, whereas in braided rivers these two length scales are comparable. Thus, the braided channel pattern is more space-filling than the anastomosed pattern. It is also worth distinguishing two somewhat different ways in which braided stream patterns can develop. In one case, 'confined braiding', there is a well-defined channelway that fills with water during floods and develops a pattern of submerged bars. As stage decreases, the bar tops emerge, producing a braided channel pattern. In 'free braiding', the braiding develops on an effectively unconfined plain. As discharge increases, an increasing number of channels is occupied, but the braid plain is never completely submerged. The relation between these two types of braiding is still not clear.

The dynamic character of braided networks owes much to interplay of their three basic elements (Ashmore 1991b): channel segments (anabranches), confluences and in-channel bars. Generally, these morphological elements are associated with locally parallel, converging and diverging bank geometries respectively. Channel segments may be straight or gently sinuous. Curved channel segments increase their SINUOSITY through erosion of their outer bank much as MEANDERING channels do, though braid anabranches often widen as they do so. Confluences are associated with channel narrowing, elevated velocities and local scour (Best 1988; Roy and Bergeron 1990). Bars are associated with channel expansion and widening, deposition (mainly on the bar periphery) and eventual splitting of the flow via scour along the bar sides. The dynamics of stream braiding largely results from the strongly nonlinear relation between flow strength and sediment flux. Confluences become scour sites because narrowing and acceleration of the flow increase its capacity for sediment transport. The scour further accentuates the narrowing and acceleration – an example of positive feedback. The converse is true in divergences. This tendency of the bed to accentuate local variability in the flow means the system never develops a static, steady-state configuration, even if water and sediment are supplied at a constant rate. This 'dynamic equilibrium' applies also to the flow of sediment through the braided network. As bars grow and are then incised by channels, sediment is impounded and released, producing highly variable sediment flow even if external conditions are steady (Ashmore 1991a).

Why do rivers braid?

Conditions commonly associated with the occurrence of braided rivers in nature include steep slopes, variable water discharge, coarse grain size and high rates of sediment supply. Empirically, we can identify sets of variables that discriminate braided from straight or meandering rivers. The most common of these discriminant plots is slope versus discharge, in which braided rivers appear at higher slopes for the same discharge than meandering rivers do. Empirical relations such as these provide hints as to the important variables, but little physical insight into the actual cause of braiding.

Historically, a major step in analysis of the causes of river patterns like braiding came with the application of stability analysis to the problem (Fredsoe 1978; Parker 1976). In stability analysis, one asks mathematically how a system responds to small perturbations. In analysing river planform the starting system is a straight channel, referred to as the 'base state'. Then one adds perturbations to the bed (and in some cases the banks), generally represented as one- or two-dimensional sine waves of infinitesimally small amplitude, and investigates how the perturbations change the flow and sediment-transport fields. If any of these perturbations changes the system in such a way as to produce its own growth, we have positive feedback and the system is unstable. This approach is based on the idea that natural systems are constantly being 'probed' by random disturbances – a tree falls in the river, for instance – that include a wide spectrum of wavelengths. A system that could not recover from such a disturbance would not last long in the real world.

In the case of rivers, the main control on planform stability turns out to be the channel aspect (width:depth) ratio. Channels narrower than about 20 times the depth tend to remain straight; those with widths roughly between 15 and 150 depths develop alternate bars, presumed to lead to meandering; and channels wider than about 150 depths develop multiple bars that are interpreted as leading to braiding.

Stability analysis was a major advance in that it provided a mechanistic foundation for understanding the origin of braiding and meandering. It also raised a number of new questions. For most rivers, the channel aspect ratio is not imposed from outside but is set by the dynamics of the channel itself. Unfortunately, the dynamics of channel width remains one of the fundamental unsolved problems of fluvial geomorphology. But it does seem clear that one of the strongest controls on width is the total effective sediment discharge (i.e. excluding the washload, suitably defined). Thus, high effective sediment loads are critical to braiding in two ways. First, high effective loads directly increase the width, directly increasing the aspect (width:depth) ratio. Second, for a given water supply, increasing the ratio of sediment to water discharge increases the slope, leading to smaller depths and thus further increasing the aspect ratio. This analysis helps explain why plots using slope and water discharge can discriminate a braiding 'regime' but suggests that neither variable is the fundamental control per se.

Chaos, complexity and braiding

The core idea of *chaos* in the scientific sense is that a fully deterministic system nonetheless can be effectively unpredictable. Surprisingly little has been done to analyse braided rivers formally as chaotic systems. It is clear that they are governed by a set of reasonably well-known deterministic equations, and they certainly appear to be unpredictable to any level of detail on timescales much longer than that required for migration of an anabranch or bar a significant fraction of its width. One especially fruitful line of analysis (Foufoula-Georgiou and Sapozhnikov 1998; Sapozhnikov and Foufoula-Georgiou 1996) has shown that braided-river plan patterns are fractal (specifically, *self-affine* fractals). The self-similarity or self-affinity that defines a pattern as fractal can occur either within one river (a small part of the river looks like a larger part), or between two different rivers (a small river looks like a larger river). An easily seen manifestation of similarity between large and small rivers is that the braided patterns one might see around town or on the beach share many basic dynamical characteristics with full-scale braided rivers. The similarity of large and small braided rivers makes braided rivers accessible to experimental study (Ashmore 1982).

Braided rivers also show a time–space scaling according to which the time evolution of a small part of a braided channel system is statistically indistinguishable from that of a larger part of the system, provided time is scaled (imagine speeding up or slowing down a film) according to a power of the ratio of the two areas being compared (Sapozhnikov and Foufoula-Georgiou 1997). These scaling results are not chaos per se, but power-law scaling of this kind is a common by-product of chaotic dynamics. The time–space scaling also implies that braided rivers may be self-organized critical systems.

Chaos theory arose from the study of atmospheric convection, and turbulent fluid flow remains one of the archetypes of chaotic behaviour. Braiding as a phenomenon seems analogous to turbulence in some respects (Paola 1996). In effect, a braided channel pattern is to a straight channel as turbulent flow is to laminar flow. Increasing the Reynolds number of laminar shear flow increases the momentum flux, which produces unstable high velocity gradients. The instability leads to a new, chaotic state that is more efficient at transferring momentum than the original laminar flow. In a straight river channel, increasing the sediment flux increases the width and (indirectly) decreases the depth, leading to unstable high channel aspect ratios. This instability leads to a new, chaotic state (braiding) that is more efficient at transferring sediment than a straight channel. (The nonlinearity of sediment flux as a function of flow velocity means that a flow system with high-speed and low-speed regions transports more sediment on average than a uniform stream with the same mean velocity.) The main stability parameter in braiding, and hence the equivalent of the Reynolds number, is the width:depth (aspect) ratio.

These observations help us understand aspects of the phenomenon of braiding not well captured in either empirical analyses or stability theory. The appearance of alternate bars in stability analyses is generally interpreted as implying a meandering plan pattern. But experimentally, meandering in channels without cohesive sediment is a transient phenomenon. Left to its own devices, a channel with alternate bars and noncohesive banks eventually evolves into a braided pattern with a low braid index. Channelized flow over noncohesive sediment cannot produce fully developed meandering, regardless of the channel aspect ratio. Evidently, just as pipe flow has two fundamental states (laminar and turbulent), channel flow in noncohesive sediment has two fundamental states: straight and braided. A fully realized meandering state requires that channels with alternate bars be stabilized, for example by cohesive sediment or vegetation.

We seem to be on the threshold of major advances in the theoretical modelling of braided rivers (see MODELS). The new field of complexity theory, which seeks unifying theoretical ideas and common behavioural patterns across a range of nonlinear systems, may help us to develop better theories of stream braiding. It is clear at this point that some aspects of the phenomenon of braiding can be captured in models that greatly simplify the detailed mechanics of flow and sediment transport (Murray and Paola 1994), while other aspects cannot. An insightful synthetic approach based on abstracting the detailed mechanics, perhaps ordered in the kind of hierarchical structure that has been used to study other complex systems, may be the most effective way of modelling braided rivers. It also may be that the best approach will simply be to develop numerical tools to solve the governing flow and sediment-flux equations on a sufficiently fine and adaptable mesh to allow for detailed simulation of the complex physics of braiding. Either way, newly emerging 'synoptic' field and laboratory data sets that capture the co-evolution of flow and topography over a whole river reach rather than a small area will be the standard against which new theoretical ideas will be tested. Of the main river types, braided rivers have proved to be the most challenging to analyse formally. The next edition of this encyclopedia will no doubt show dramatic results from some of the simulation efforts now under way.

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CHRIS PAOLA

BROUSSE TIGRÉE

One of the most striking forms of PATTERNED GROUND is the brousse tigrée as identified from aerial photographs in West Africa (Clos-Arceduc 1956). This pattern is composed of alternating bands of vegetation and bare grounds aligned at the contour. From the air, these bands or arcs form a distinctive pattern similar to the pelt of a tiger.

Similar patterns have been recognized from aerial photos from many parts of the world. They were called *mulga groves* in Australia and *mogote* in Mexico. Ground truth may differ since banded vegetation can consist either of grass (Mauritania, Somalia, Sudan), shrubs (Australia, Mexico), or trees (Australia, Mali, Niger). They occur only where the co-occurrence of several critical conditions is met: low annual rainfall (75–650 mm), gentle and uniform slope (0.2–2 per cent) and crusting soils. These factors favour water runoff sufficient to produce sheet OVERLAND FLOW over a distance of a few tens of metres but insufficient to trigger the concentration of runoff into RILLS. In flatter landscapes, the vegetation is no longer banded but spotted because of the nondirectional runoff pattern. Slope also controls the wavelength (band plus interband width) of the pattern even at a local scale. The wavelength decreases exponentially with increasing slope gradient. Differences observed in the soils of bands and associated interbands are a consequence rather than a cause of banded ground (Bromley *et al*. 1997).

For a given slope, the mean annual rainfall determines the ratio between the width of the vegetation bands of arcs and the width of the bare bands. The bands accumulate runoff water and function as if they were in a higher rainfall climatic regime. The optimal rainfall for band development increases with increasing percentage of high rainfall event and decreasing duration of the rainy season. This optimal annual rainfall increases from 250 mm in central Australia to 550 mm in south-west Niger.

These banded patterns are natural examples demonstrating the principles of water, soil and nutrients conservation in space and time. Although the role of wind cannot be overlooked in certain circumstances, surface hydrological processes are critical to the ongoing functioning of banded landscapes. Three main processes are involved: differential infiltration, obstruction to overland flow, and efficient nutrient cycling. Soil crusts dominate in the interbands, resulting in low infiltration, whereas vegetation, litter and bioturbation effects facilitate high infiltration rates in the bands and arcs. The banded patterns act as a natural water harvesting system, the overland flow produced from the bare and impermeable interbands running onto the bands. Vegetation bands tend to obstruct or regulate sheet flow so that sediments and organic matter are continually being deposited and conserved within the bands, forming a natural bench structure that limits soil erosion. Due to the rainwater redistribution, the bands receive from two (in south-eastern Australia) to four times (in south-west Niger) the rainfall at the site. The centre of the bands has abundant biopores enabling effective water capture from the interband. The soils in the bands also concentrate more soil nutrients and organic matter than the adjacent interbands. This resource concentration enables the formation of a forest system, the productivity of which equals and can even double that of adjacent non-banded landscapes.

These systems can persist in the face of severe drought by adjusting the proportion of runoff and runon areas. They can also resist the stress and disturbance caused by moderate land use. The earliest indicator of deterioration is the decline in the contrast between the two mosaic phases. The late stage in degradation is characterized by disruption of the band pattern. Overgrazing is considered to be the prime cause of deterioration of banded landscapes in Australia. Firewood and timber harvesting threaten the brousse tigrée in West Africa.

Models have demonstrated that these patterns may result either from landscape degradation or rehabilitation, but the natural initiation of banded landscapes has never been observed. The slow upslope migration of the bands is also a debated topic. It is linked to the runoff/runon theory that underpins the basic functioning of banded vegetation. The obstruction of overland flow by the bands would favour the upslope germination of pioneer plants in this upslope edge and the decline of vegetation due to resource shortage at the downslope edge. This notion of upslope band migration is strongly supported by an array of arguments such as the seedling concentration on the upslope edge of the band, the decaying vegetation in the downslope edge, the sequence of soil crust types across the interbands, and the marked gradient in soil organic matter. The migration 'velocity' of bands has been assessed using a variety of methods including field monitoring with benchmarks, digitized aerial photographs, age distribution of trees with dendrochronology, and TRACERS (residual ^{137}Caesium) distribution in the soil, under a wide range of climatic and topographic conditions. The fastest observed migration was $1.5\,\mathrm{m\,yr^{-1}}$ for grass bands and $0.8\,\mathrm{m\,yr^{-1}}$ for trees and shrubs. Because of some stationary systems, the migration of vegetation bands cannot be regarded as an invariable property of the banded systems.

In the arid and semi-arid environment, the banded patterns are clear examples of heterogeneous landscapes that are more sustainable than homogeneous systems. The lessons drawn from them lead to the recognition of the ecological value of water harvesting and runoff farming.

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SEE ALSO: crusting of soil

CHRISTIAN VALENTIN

BRUUN RULE

The prospect of accelerated sea-level rise as a consequence of global warming has renewed interest in models that link sea-level rise and coastal change, such as the shoreline translation model of Cowell and Thom (1994). But to date, no model is better known or more widely accepted than that of Per Bruun.

In 1962 Bruun (1962) proposed that with a rise in sea level, the profile of a beach and its nearshore zone would move landward and upward, and that the quantity of sediment eroded from the upper part of the profile would be transported seaward to build up the adjacent seafloor by an amount equivalent to the sea-level rise (Figure 15). In this model the retreat of the beach (R) is given as:

$$R = XS/Y,$$

where R is the difference in distance between the initial sea level–profile intercept and the intercept after sea-level rise, X is the horizontal length from shore to the limiting depth, S is the sea-level rise, and Y is the vertical dimension of the profile, which is the sum of the limiting depth below sea level and the top of the foredune above sea level.

Early testing of the model in a small-scale laboratory wave-table experiment, as well as sequential field measurements of shore profiles around Cape Cod, Chesapeake Bay and Lake Michigan during and after episodes of rising water level, tended to support the basic tenets of the model, such that in the 1970s it became known as the Bruun Rule, after temporarily being declared a 'theory' by Schwartz (1967). However, such status has not been without its critics, for rarely are the model's basic assumptions satisfied in the real world of multidimensional coastal morphodynamics (Healy 1991).

The Bruun Rule is a two-dimensional cross-shore model applicable to long straight sandy shorelines that adjust to a rise in sea level over decadal to centennial timescales. The original model assumes that the initial and final profiles are in equilibrium, that a constant profile shape is preserved over the period considered, that the total quantity of sediment in the cross-section is conserved, and that a constant water depth is maintained in the offshore zone as sea level rises. Clearly, these assumptions are unrealistic. For instance, given the fact that beach erosion is already widespread, it is unlikely that it would be possible to determine an EQUILIBRIUM SHORELINE, or that the shape of the profile would not change through time. It is also unrealistic to expect a beach profile to conserve sediment and

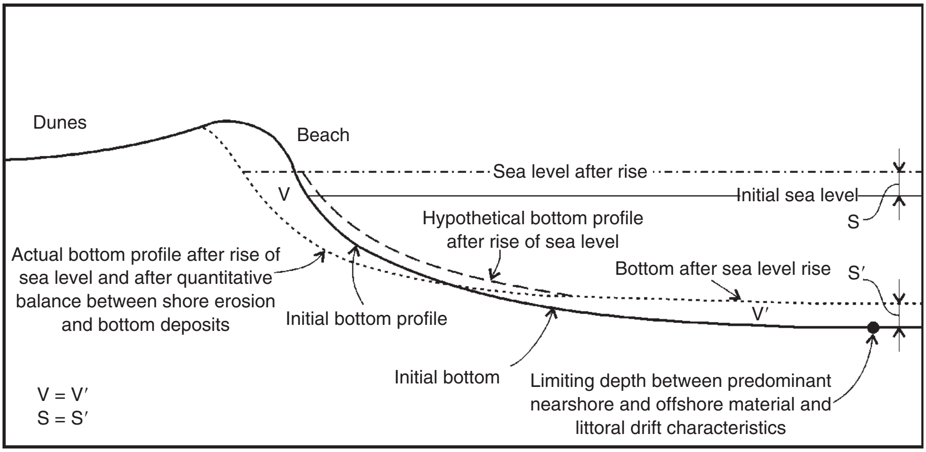

Figure 15 The Bruun Rule implies that the sediment volume removed from the beach and nearshore (V) must equal the sediment volume on the lower shoreface (V′) and that the lower shoreface aggrades in direct proportion (S′) to the rise in sea level (S). (After Bruun 1962 and Dubois 2002)

not to have sediment leakage, either as gains or losses resulting from LONGSHORE (LITTORAL) DRIFT, which is such a common process on sandy beaches. Similarly, wind erosion is not included in the Bruun Rule even though it can result in significant profile change through deflation of the beach and accumulation on a foredune at the landward end of a profile. Determination of the seaward end of a profile (closure depth) is equally problematical and maintenance of a constant water depth as implied in the Bruun Rule would result in a bathymetric or sediment discontinuity that in reality is difficult to define.

In spite of such difficulties, the Bruun Rule remains particularly attractive for several reasons. First, it is simple in concept and intuitively attractive given the fact that over the past hundred years or so global sea level has been rising at rates of around 1–2 mm per year, and that during that time approximately 70 per cent of the world's sandy shorelines have been eroding. Second, it can give quantitative results such that shore retreat will be 50 to 100 times the rise in sea level. For instance, a rise in mean sea level of 50 cm would result in beach recession of 25 to 50 m. And, third, the Bruun Rule has proven flexible enough to spawn a number of derivative models. Some refinements were proposed by Bruun (1983, 1988) himself, others by Dean and Maurmeyer (1983) who upscaled the concept to account for the landward and upward migration of an entire barrier island system. But the most persistent alternative model builder has been Dubois (1992, 2002).

Because the Bruun concept is not dependent on the shape of the shore profile, more complex topographies than in the original figure (Figure 15) can be incorporated (Dubois 1992). In Figure 16 zones of erosion and deposition are identified associated with onshore bar migration after shoreward displacement of the whole profile resulting from sea-level rise. Moreover, in this alternative model the eroded beach material is not only displaced seaward (as in the original Bruun model) but is also moved in a landward direction and washed or deflated on to the dune face or into a backing swale or lagoon. While overall conservation of sediment should be maintained within the whole profile, fine suspended sediment can be carried seaward and deposited in deeper water such that the lower shoreface and ramp do not accrete following the rise in sea level, but are simply abandoned by wave action (Figure 16).

In a comprehensive review of the response of beaches to sea-level changes, Working Group 89 of the Scientific Committee for Ocean Research (SCOR 1991) concluded that the quantitative predictions of shore change based on the Bruun

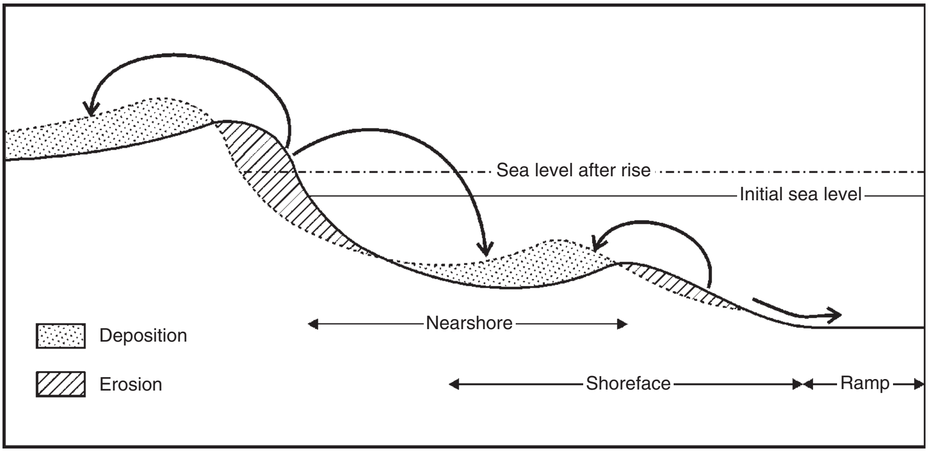

Figure 16 A two-dimensional model of a shore profile responding to a rise in sea level. Arrows show potential directions of sediment transport. Note that Bruun's Rule is embedded in this model, although it contributes only a small amount to the total shore erosion caused by rising sea level. (After Dubois 1992)

model are dependent on a number of parameters that are difficult to define, that there may be a significant time lag of beach response to sea-level rise, though the principal hindrance in achieving acceptable predictions is that the model does not include other sediment budget components that can result in either coastal accretion or erosion. Nevertheless, the SCOR group did suggest that the Bruun Rule could be used, though only for order-of-magnitude estimates of potential shore recession rates in appropriate coastal settings.

There is little doubt that globally sea level is rising and that sandy shores around the world are continuing to erode, including barrier beaches and barrier islands. The Bruun Rule gives an insight into how sea-level rise and coastal erosion are coupled, though we also know there are a host of other factors that contribute to coastal erosion independent of sea level.

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SEE ALSO: barrier and barrier island; beach–dune interaction; equilibrium shoreline; longshore (littoral) drift

ROGER F. McLEAN

BUBNOFF UNIT

Unit providing a useful means to quantify the rate of operation of diverse geomorphological processes as a rate of ground loss (perpendicular to the surface) or slope retreat. A unit equals 1 mm per 1,000 years, equivalent to $1\,m^3\,km^{-2}\,a^{-1}$ (Fischer 1969).

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A.S. GOUDIE

BURIED VALLEY

A buried valley is the bedrock expression of a valley buried by more recent deposits. These features are surprisingly common but are not well known as they have no surface expression. They are usually identified following borehole information or other sub-surface investigations employing geophysical techniques. The identification of these features is often an important element in the reconstruction of the geomorphological history of an area.

A number of types of buried valley can be identified. First, there are those buried valleys which are the result of glaciation. These can be sub-aerially eroded valleys buried by deposits of glacial origin. A good example of this type is afforded in the English Midlands by the Proto-Soar valley. This broad sub-drift valley has its head between Stratford-on-Avon and Warwick and heads northeastwards towards Leicester where it underlies the contemporary river Soar, a major tributary of the Trent. Before the glacial event which buried this surface, the main watershed of England between Avon/Severn drainage to the west and Soar/Trent drainage to the east lay some 30 km at least to the west of its present position which is on top of a thick plug of glacial deposits. Elsewhere, buried valleys have been described which themselves have been created by subglacial processes and then subsequently been buried. In interpreting the buried valleys identified widely in East Anglia, Woodland (1970) drew attention to the 'tunnel valleys' of Denmark and northern Germany. In East Anglia, many of the buried valleys appear to be quite narrow, up to 500 m wide and often 100 m deep, whereas the features in Denmark are broader and shallower. Woodland, however, attributes a similar origin to these features – subglacial erosion beneath an ice sheet. In the case of the East Anglian examples, they have been infilled by a wide range of often complex sediments which mask the former topography. Other authors have preferred to explain the excavation of these tunnel valleys through glacial modification of existing valleys (West and Whiteman 1986).

Second, there are buried valleys resulting from changes in SEA LEVEL following the last glacial event. These are widespread around many coastlines where marine transgressions, estuarine deposition and alluvial fill have buried a landscape graded to a lower sea level. Thus under many existing rivers can be found buried valleys that represent the valley of the former river draining into a sea that might have been 100 m or so lower. Boreholes can reveal that the essentially level surface of the contemporary alluvium conceals an irregular surface comprising valley forms which often, but not always, parallel the existing drainage.

Finally, mention must be made of the numerous buried valleys, usually in urban areas, where human activity has been responsible for modification of the valley topography. In some cities, rubble has been used to infill minor valleys to create flat land for urban land uses.

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TERRY DOUGLAS

BUTTE

Butte is a small steep-sided and flat-topped hill, built of flat-lying soft rocks capped by a more resistant layer of sedimentary rock, lava flow or duricrust, surrounded by a plain. Butte is smaller than MESA and may be considered as a more advanced stage of mesa degradation, although there are no formal criteria to distinguish between the two. Together with mesas, buttes are outliers, indicative of long-term scarp retreat. They occur in front of CUESTAs and plateau margins, their morphology being best pronounced in arid and semi-arid regions.

PIOTR MIGOŃ

C

CALANQUE

Coastal inlets (such as those to the east of Marseilles) which tend to be of a gorge-like form. They are widespread around the Mediterranean Sea and may be karstic dry valleys which have been partially drowned, as a result of the Flandrian transgression of the Holocene. Their positions may be fault controlled. In Mallorca they are called *calas*.

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A.S. GOUDIE

CALCRETE

A term, proposed by Lamplugh (1902), to describe a terrestrial near-surface accumulation of predominantly calcium carbonate ($CaCO_3$) which occurs in a variety of forms ranging from powdery to nodular to highly indurated. It results from low temperature physico-chemical processes operating within the zone of WEATHERING which lead to the displacive and/or replacive introduction of $CaCO_3$ into a soil profile, sediment, rock or weathered material. Calcretes develop as a result of carbonates in solution moving laterally and vertically through vadose and shallow phreatic groundwater systems until they become, over time, saturated with respect to $CaCO_3$ and precipitate as calcite crystals (Wright and Tucker 1991). Calcretes often occur within soil profiles, where they may form single or multiple horizons, but they are not a type of soil. The term is synonymous with CALICHE (SODIUM NITRATE) and kunkur but distinct from other $CaCO_3$ cemented materials such as cave SPELEOTHEMs, lacustrine algal STROMATOLITEs (STROMATOLITHs), TUFA AND TRAVERTINE, BEACH ROCK or AEOLIANITE.

Calcretes are estimated to underlie 13 per cent of the Earth's land surface and are most widespread in semi-arid regions. They form an important component of many contemporary dryland landscapes, and, where well indurated, may act as a threshold (see THRESHOLD, GEOMORPHIC) to erosion. Important areas of occurrence include the High Plains of the USA (e.g. Gile *et al*. 1966; Machette 1985), Africa north of the Sahara (e.g. Goudie 1973), the Kalahari of southern Africa (e.g. Watts 1980; Netterberg 1980), central and western Australia (e.g. Mann and Horwitz 1979; Milnes and Hutton 1983), and parts of southern Europe (e.g. Nash and Smith 1998). The close association between calcrete distribution and present-day dryland regions has led to the widespread use of calcretes in the geological record as indicators of past aridity. However, it is critical that the mode of origin of any calcrete is identified before it can be interpreted in this way. Whilst carbonate accumulation within soil may require a semi-arid climate, calcretes developed by other mechanisms may form under much wetter conditions. Non-pedogenic Holocene calcretes have, for example, been found in temperate locations such as the UK (Strong *et al*. 1992). Furthermore, calcrete accumulation is closely controlled by carbonate supply.

Calcretes are highly variable in appearance and range from thin rock coatings to massive horizons. Thickness varies with the mode of origin and stage of development, with laminar calcretes rarely exceeding 0.25 m whilst multiple pedogenic and groundwater profiles may reach tens of metres thickness. Most calcretes are white, cream or grey in colour, though mottling and banding is common. Calcretes are predominantly cemented by calcite with some $CaMg(CO_3)_2$ (dolomite) often present. The size and shape of calcite crystals is dependent upon the composition of the host material, the duration of wetting and the influence of biological mechanisms. Cements are typically dominated by microcrystalline carbonate (or micrite) although larger crystals of sparry calcite may be present. If significant biological fixation of carbonate occurred during development, the calcrete is likely to exhibit a complex beta fabric dominated by organic structures when viewed in microscopic thin-section as opposed to simpler alpha fabrics developed by inorganic mechanisms (Wright and Tucker 1991). The mean global chemical composition of calcrete is *c.*78 per cent $CaCO_3$, 12 per cent SiO_2, 3 per cent MgO, 2 per cent Fe_2O_3 and 2 per cent Al_2O_3 (Goudie 1973), although variations occur dependent upon the host material chemistry, cement type, presence of authigenic silica and silicates, mode of origin, and stage of development.

There are a range of classification schemes for calcrete, of which the most widely employed use morphological criteria. Netterberg (1980), for example, recognized a range of forms including calcareous and calcified soils, powder, nodular, honeycomb, hardpan, laminar and boulder calcretes, a sequence which also reflects the phases of development of many calcretes. Gile *et al.* (1966) and Machette (1985) have proposed a scheme to assist the identification of calcretes at different stages of development, with stage I–III calcretes consisting of morphologically simple carbonate accumulations within soils progressing to more mature horizons by stages IV–VI. Calcretes have also been classified on the basis of their hydrological setting, with vadose, capillary fringe and

Table 5 A genetic classification of calcrete types

Environment of formation	Calcrete type	Incorporated calcrete types	Mode of formation
Pedogenic	Pedogenic calcrete	Caliche; kunkar; nari; petrocalcic horizons	Developed by vertical redistribution of calcium carbonate within a soil profile
Non-pedogenic	Non-pedogenic superficial calcrete	Laminar crusts; case hardening; gully bed cementation	Formed by surficial transport of calcium carbonate
Non-pedogenic	Non-pedogenic gravitational zone calcrete	Gravitational zone calcrete	Formed by downward accumulation of calcium carbonate in irregular permeability channels
Non-pedogenic	Non-pedogenic groundwater calcrete	Valley calcrete; channel calcrete; deltaic calcrete; lake margin calcrete; alluvial fan calcrete; fault trace and other groundwater calcretes	Formed by lateral transport of calcium carbonate
Non-pedogenic	Detrital and reconstituted calcrete	Recemented transported calcrete; calcretes which are brecciated and recemented *in situ*	Formed by recementation of existing fragmented or brecciated calcrete

Source: After Carlisle (1983)

phreatic types identified. Other schemes have subdivided calcretes by their dolomite content (Netterberg 1980) or by the relative abundance of alpha and beta cements (Wright and Tucker 1991). However, none of these classifications completely distinguishes between calcretes formed by different mechanisms. As such, the most helpful scheme is Carlisle's (1983) genetic classification (Table 5) which subdivides calcretes into pedogenic and non-pedogenic forms using geomorphological, chemical, macro- and micromorphological criteria.

Pedogenic calcretes develop near the land surface, usually in areas of low slope angle, through the mobilization, redistribution and relative accumulation of $CaCO_3$ within a soil profile. Formation may also involve some absolute accumulation of $CaCO_3$ if there are additional carbonate inputs to the profile. Such calcretes commonly show enrichment in $CaCO_3$ up-profile and consist of a powdery or nodular basal section overlain by a more massive hardpan which may, in turn, be capped by a laminar crust. Cements are usually dominated by micrite and, because of the mechanisms by which they develop, exhibit a complex micromorphology. Non-pedogenic calcretes encompass a wide variety of types, ranging from laminar crusts developed on rock or other calcrete surfaces by evaporation and/or biological fixing of $CaCO_3$, to detrital and reconstituted calcretes formed by the cementation of pre-existing fragmented crusts. By far the largest group are the groundwater calcretes. These are calcretes developed in channel, valley, alluvial fan, delta and lake marginal sediments, usually in the absence of soil-forming processes and sometimes at depths of tens of metres beneath the land surface. They can be distinguished from pedogenic calcretes by their lack of profile development, normally simple micromorphology and the presence of more crystalline calcite cements, especially where formation occurred at or below the water table (Nash and Smith 1998).

Despite the wide range of mechanisms by which they can develop, all calcretes result from the solution, movement and subsequent precipitation of $CaCO_3$, described by the following chemical reaction:

$$CO_2 + H_2O + CaCO_3 \leftrightarrow Ca^{2+} + 2HCO_3^-$$

Calcretes require a carbonate source, usually released as a result of $CaCO_3$ dissolution (Goudie 1983). $CaCO_3$ solubility is closely linked to environmental pH, with solubility rapidly increasing below pH 9.0. Mechanisms which lower pH and drive the reaction to the right, such as the introduction of weak carbonic acid ($CO_2 + H_2O$ in the equation) or an increase in soil CO_2 partial pressure, will trigger dissolution. Sources of carbonate can be distant or local to the site of formation, and include weathered bedrock, volcanic (and other) dust and organic remains. Once carbonate is in solution, it may be moved laterally and/or vertically to the site of formation. Lateral transfer mechanisms may include transport in solution via ephemeral or perennial rivers as well as in shallow or deep groundwater systems. Vertical transfers include percolation of surface water or capillary rise from the water table. Carbonate precipitation (where the above reaction proceeds to the left) may be triggered by a variety of factors which lead to the concentration of carbonate-rich solutions and/or cause environmental pH to increase. Foremost amongst these are evapotranspiration, biological processes, decreases in CO_2 partial pressure, CO_2 degassing, and the common ion effect (Goudie 1983; Salomons and Mook 1986).

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SEE ALSO: duricrust; silcrete

DAVID J. NASH

CALDERA

Calderas are large circular or elliptical volcanic depressions whose diameter (typically several or several tens of kilometres) greatly exceeds those of any included vents. They are formed by the evacuation of a magma chamber within the crust, and subsidence of the overlying rocks. Calderas can form on volcanoes of different magma composition, from low silica (mafic) to high silica (silicic), though the mechanisms vary. Though there are no hard and fast distinctions, the term crater tends to be reserved for smaller features created as a result of excavation of rock during explosive eruptions or smaller-scale collapse (collapse pits). The highest magnitude explosive eruptions on Earth, sometimes called super-eruptions, have generated the largest calderas. Volcanoes that have experienced more than one super-eruption are colloquially known as super-volcanoes. The geometries and structures of the resulting nested and overlapping calderas can be complex, and obscured by post-collapse uplift, volcanism and erosion.

Origins and development

Super-eruptions involve magma volumes of several thousand km^3 (masses up to 10^{16} kg). The rapid removal of such an amount of material from a crustal magma chamber invariably induces failure of the overlying rocks. There is a rough correspondence between the volume of magma erupted and that of the hole left in the ground by the caldera collapse. The largest known Quaternary eruption occurred about 74,000 years ago, and expelled an estimated 7×10^{15} kg (2,800 km^3) of silicic magma, making a significant contribution to the 100 km $\times$ 30 km caldera complex occupied today by Lake Toba in northern Sumatra (Oppenheimer 2002). Toba can certainly be classed as a super-volcano, since at least two similar events occurred around 840,000 and 500,000 years ago, as can Yellowstone (USA), whose last super-eruption took place about 600,000 years ago.

Important insights into caldera evolution associated with explosive eruptions have been gained from detailed investigations of a number of much smaller historic and prehistoric calderas, for example at Crater Lake (Oregon, USA; Bacon 1983), and Santorini (Greece (Plate 19); Druitt *et al.* 1999), and also ancient examples such as Scafell caldera of the English Lake District, Ordovician in age but revealing much of its structure thanks to erosion (Branney and Kokelaar 1994). Several subsidence processes have been recognized (Lipman 2000). Larger calderas tend to involve piston-like (plate) collapse, where the floor remains largely undeformed. The collapse occurs along steep ring faults, with vertical displacements of about 1 km. In contrast, downsag subsidence does not preserve the coherence of the developing caldera floor, which is instead tilted and flexed. Intermediate between these two is trapdoor subsidence, which occurs when the caldera floor remains hinged along part of its length but elsewhere has subsided in plate fashion. Geometrically complex systems of arcuate faults and subsiding blocks reflect, in some cases, the breakup of the floor during eruption but prior to ring-fault subsidence, and are referred to as piecemeal calderas (Branney and Kokelaar 1994). Relatively small-scale collapses resulting from modest explosive eruptions from a central vent, such as that of Pinatubo (Philippines) in 1991, are sometimes referred to as funnel calderas. These lack a bounding ring fault or coherent subsided plate.

Calderas associated with low silica (mafic) volcanoes such as those found in Hawai'i have somewhat different origins to their silicic counterparts. Some interpret their development as a late stage in the growth of Hawaiian shields; others see them as a recurrent process. Clearly, those on

Plate 19 Santorini volcano, Greece. The last major caldera-forming eruption occurred in the mid-seventeenth century BC but is only the most recent in a series of eruptions exceeding 1,014 kg in magnitude (Druitt *et al.* 1999). The caldera rim is partly submerged such that the caldera is open to the sea

Mauna Loa and Kīlauea are very young features suggesting that they are rejuvenated at intervals of a few centuries, rather than the many millennia that can separate subsidence events on silicic volcanoes. Again, unlike silicic systems, the volumes of caldera subsidence on mafic volcanoes do not bear any obvious relationship with erupted volumes; the volumes of the young Hawaiian calderas are larger than any known Hawaiian eruptions. While Mauna Loa and Kīlauea appear superficially to be well-defined piston subsidence structures, Walker (1988) has suggested on the basis of mapping of the 2 Myr old Koolau volcano on Oahu, where erosion has exposed its 1 km depth roots, that downsagging may prevail in the central parts of funnel-shaped subsidence structures underlying the calderas. He suggests that, rather than simple piston subsidence into an evacuated magma chamber, the Hawaiian calderas develop as the dense intrusive rocks associated with the magma chambers sink into the warm lithosphere beneath.

Geomorphology

Calderas are enclosed by a topographic rim that is simply the head of the escarpment bounding the caldera. The inner walls can form cliffs in young calderas but retreat through time as a result of landslides. In map view, most large calderas reveal bites in the rim due to larger slope failures. The rock redistributed by landslides may form a collapse collar on the caldera floor. The arcuate bounding faults (ring faults) are sometimes exposed in deeply eroded calderas, and where observed, generally dip near-vertically or steeply inwards.

The largest calderas often enclose a central elevated massif. The vertical extent of the upheaval in these resurgent calderas can be 1 km or more. Toba is a good example; a lake occupies much of the caldera but an island – Samosir Island – occupies much of the lake. The island is composed substantially of the pyroclastic intracaldera fill from the Younger Toba Tuff eruption. Lacustrine sediments can be found several hundred metres above the present lake level and testify to substantial post-caldera uplift. Yellowstone is another example of a resurgent caldera. The mechanisms of resurgence, however, are not well understood. Refilling of the magma chamber is one possibility, along with the exsolution of remaining volatiles in the caldera eruption chamber and bubble formation (vesiculation) causing an increase in volume, expressed at the surface in the form of uplift (Marsh 1984).

Erosion rates of the pyroclastic deposits of caldera-forming eruptions can be rapid, especially for non-welded portions. Where developed, columnar jointing, akin to that observed in mafic lava plateaux (see LAVA LANDFORMS), can strongly influence the drainage fabric. In arid environments, the outflow sheets of ash flow deposits (i.e. those outside the caldera) may experience rapid aeolian erosion. This is evident in the wind-sculpted morphology of the Central Andean ignimbrites, which are adorned with YARDANGS and deflationary hollows. Wigwams or tent rocks are another common feature of the pyroclastic deposits of caldera-forming eruptions. These may develop by the intersection of drainage channels, or more commonly by the action of a resistant block of lava within the deposit, protecting the underlying material from erosion.

Hazards and climate change

Along with bolide impacts, large caldera-forming eruptions are the most catastrophic geologic events that affect the Earth's surface. Compared in terms of energy release, they are more frequent than bolides, however. A super-eruption today would have devastating impacts on regional populations and the global economy. The massive quantities of sulphur gases that can be released in super-eruptions strongly perturb atmospheric chemistry and radiation, with potentially global-scale climatic consequences (Rampino and Self 1993). Currently, little is known about the precursory events that might lead up to a super-eruption but numerous calderas worldwide have shown signs of unrest in the historic period (Newhall and Dzurisin 1988).

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SEE ALSO: lava landform; volcano

CLIVE OPPENHEIMER

CALICHE (SODIUM NITRATE)

The term has been used for both CALCRETE and for sodium nitrate deposits. The most famous and important deposits of the latter in the world occur in the Atacama Desert (Ericksen 1981), though others are known in California and Antarctica. Chile possesses a band of nitrate-containing terrain up to 30 km wide and 700 km long. Much of the Coastal Range is mantled with nitrate-bearing saline-cemented regolith, commonly ranging from a few tens of centimetres to a few metres in thickness. The petrography of the deposits, which cause extreme bedrock disintegration, is described by Searl and Rankin (1993). Most of the deposits lie at altitudes below 2,000 m, though some occur as high as 4,000 m. Low-grade deposits are also known in the coastal desert of Peru several hundred kilometres north of the Chilean border. Many of the deposits were extensively worked in the late nineteenth and early twentieth centuries. The landscape is now scarred with the pits from which nitrate was dug and is dotted with waste heaps.

The reason for the localization of the nitrate deposits appears to be the extreme aridity of the area, for sodium nitrate is more soluble in water than most common crust materials. The Atacama is among the driest and oldest of the world's deserts, and the average annual rainfall is less than 1 mm in the areas where the nitrate deposits are most prevalent. In any given part of the desert measurable rainfall (1 mm or more) may be as infrequent as once every 5–20 years. Heavy rainfall of a few centimetres or more may occur only a few times each century.

The Chilean nitrate fields occur typically in areas of low relief characterized by rounded hills and ridges and by broad, shallow debris-filled valleys. Significantly for their origin, they occur in all topographic positions from tops of hills and ridges to the centres of the broad valleys, though the richest deposits that have been worked most extensively tend to be on the lower slopes of hills.

Such a catholicity of geomorphological siting tends to imply that the nitrates have been derived as atmospheric inputs from the sea or from volcanic emissions, a mechanism supported by the fact that the nitrates occur on all rock and sediment types (Ericksen 1981). The model proposed by Ericksen (1981: 32) is that there has been long-term accumulation of atmospherically derived saline material for perhaps 10–15 million years (i.e. since the Mid-Miocene, under conditions of general extreme aridity). The sources of the material would include sea spray, volcanic emanations, photochemical reactions and dust from *salars* (salt lakes). The ore-grade nitrate deposits are formed by accumulation of saline materials on very old, flat to gently inclined or undulating landsurfaces, where rainwater dissolved the more soluble component and redeposited them at deeper soil levels. Ericksen's model of nitrate bedformation as a result of long-term atmospheric deposition has been confirmed by recent stable isotope studies (Böhlke *et al.* 1997).

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A.S. GOUDIE

CALVING GLACIER

Calving GLACIERs terminate in water and lose mass by calving, the process whereby masses of ice break off to form ICEBERGs. Since they may be temperate or polar (see glaciers), grounded or floating, and may flow into the sea or into lakes, many types exist. They are widely distributed, but while lake-calving glaciers may exist in any glacierized mountain range, tidewater glaciers are currently confined to latitudes higher than 45°. Typically calving glaciers are fast flowing and characterized by extensional (stretching) flow near their termini, resulting in profuse crevassing. They terminate at near-vertical ice cliffs up to 80 m high. Calving activity above the waterline comprises a continuum from small fragments of ice to pillars the full height of the cliff. Below the waterline much ice may be lost through melting, but in deep water buoyancy causes infrequent but high magnitude calving events. In lakes, thermal erosion (melting) at the waterline can cause calving by undercutting the cliff. Calving permits much larger volumes of ice to be lost over a given time than melting.

Glaciers calve faster in deeper water. This correlation between calving rate (u_c in metres per annum) and water depth (h_w) is linear, and can be simply expressed as $u_c = ch_w$. The value of the coefficient c varies greatly in different settings, being highest for temperate glaciers and lowest for polar glaciers. Also, for any given water depth, calving is an order of magnitude faster in FJORDs than in lakes. The established correlation between calving rate and water depth may or may not imply that faster calving is *caused* by deeper water.

Calving glaciers are significant for three main reasons:

1 *Glacier dynamics* Calving glaciers comprise the most dynamic elements of many of the world's ice masses, and calving is the major means of ice loss from the two continental ICE SHEETs of Antarctica and Greenland. During the waning stages of Quaternary glacials (see ICE AGES (INTERGLACIALS, INTERSTADIALS AND STADIALS)), calving was the dominant process of mass loss around the mid-latitude ice sheets; the efficiency of calving helps to explain the catastrophic rates of ice sheet disintegration. Armadas of icebergs discharged during ice sheet collapses are believed to have caused global climate change by altering oceanic circulation.

2 *Non-climatic behaviour* Calving glacier fluctuations are highly sensitive to topographic controls (see PINNING POINT). Some tidewater glaciers fluctuate cyclically, in ways unrelated to climate, over distances of tens of kilometres and over timescales of centuries to millennia. Therefore neither the contemporary behaviour of calving glaciers nor the geomorphological records from past fluctuations (see MORAINE) are reliable indicators of climatic change.

3 *Socio-economic impacts* Calving glaciers constitute significant resources and GEOMORPHOLOGICAL HAZARDs for society. Resources include tourism and the potential for

harnessing Antarctic tabular icebergs as a source of freshwater, while hazards include icebergs and OUTBURST FLOODS.

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CHARLES WARREN

CAMBERING AND VALLEY BULGING

Cambering occurs where large-scale valley incision has exposed gently dipping bedrock in which competent strata overlie less competent strata. Cambered strata consist of an attenuating drape of competent caprock extending down the valley sides. This drape of CAPROCK shows evidence of extension, accommodated by deep fractures termed *gulls* (Figure 17). Gulls run parallel to the contours and separate intact caprock blocks. These blocks tend to tilt forward, increasing apparent dip and producing the widely reported 'dip-and-fault structure' (Figure 17). Cambering is often associated with the development of anticlinal deformation within the less competent strata beneath the valley axis, the resulting structure being termed *valley bulging*. Cambering and valley bulge structures are thought to have formed during the Quaternary Period. They reflect ice segregation processes within the less competent strata (usually clays) during PERMAFROST aggradation at depth, and subsequent thaw consolidation processes caused by thawing at the base of the permafrost (underthaw) during permafrost degradation in the transition from cold glacial to warm interglacial stages. Hutchinson (1991) has provided a detailed review of processes and structures involved in cambering and valley bulge development.

Classic descriptions of cambering and valley bulging come from the Jurassic Limestones and underlying Upper Lias clay in England (Chandler *et al.* 1976; Hollingworth *et al.* 1945; Horswill and Horton 1976). At the Empingham Reservoir, the cambered limestone strata are draped across much of the valley side. Gulls separate cambered blocks, which display classic dip-and-fault structure. Valley bulging is highlighted by disturbance of marker horizons within the Upper Lias clays. Disturbed brecciated clay extends to a depth of 25 to 30 m, the base lying parallel to the present ground surface, suggesting that the phase of brecciation occurred after the main valley relief was established. The disturbed Lias is underlain by a sheared plane of decollement at a depth of approximately 60 m under the valley crest and 30 m under the valley bottom. It is likely that the brecciated clay fabrics reflect the combined effects of ice segregation, creep and shear.

Hutchinson (1991: Figure 5.5) provides a model for cambering and valley bulging based on estimated displacements in the Lias at

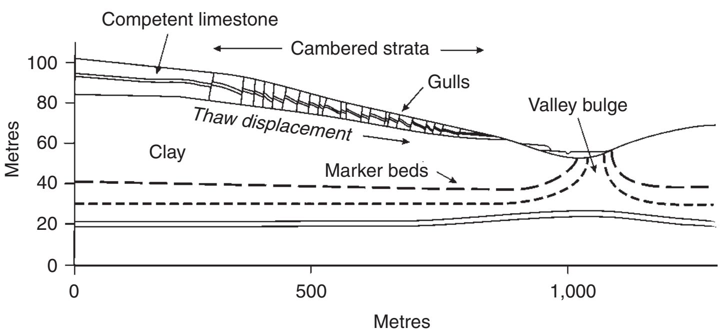

Figure 17 Cambering and valley bulging, based on the Gwash Valley, Lincolnshire, England (Horswill and Horton 1976)

Empingham (Vaughan 1976). The following stages are envisaged:

1 Initial incision of the river leading to unloading and valley rebound.
2 Permafrost develops, with ice segregation processes increasing ice contents of the frozen clay.
3 Valley-ward creep of the frozen clay occurs in response to lateral stresses. This causes extension and initial cambering of caprocks.
4 Permafrost degradation due to large-scale climate warming leads to thaw from the surface downwards, enhanced surface solifluction, and displacement of caprock down the valley sides over the thaw-softened clays beneath.
5 Thawing of the permafrost base due to geothermal heat flux is associated with thaw consolidation and high pore pressures within an effectively confined thawed stratum.
6 Lateral extrusion of the thawed clay at the permafrost base towards the valley bottom leads to compression along the valley axis and pronounced increase in the valley bulge structure.

Cambering and valley bulging represent some of the largest structures attributed to permafrost. Their presence as Quaternary relict features may be of considerable significance to engineering works such as the design of foundations for buildings which are particularly affected by the presence of gulls, and dam construction, where voids may increase water seepage, and deep-seated shearing may affect foundations.

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CHARLES HARRIS

CANYON

A canyon is a long, deep, relatively narrow, steep-sided valley, often cut through bedrock which forms precipitous cliffs along the valley walls. The word comes from the Spanish cañon. Canyons are formed by running water. The term is typically used for such features in arid and semi-arid regions, such as the western United States (e.g. the Grand Canyon in Arizona, USA). Canyons are similar to gorges (see GORGE AND RAVINE), but the side-walls are usually not as steep, and canyons are typically larger than gorges (e.g. the Grand Canyon contains an 'Inner Gorge' through which the Colorado River runs). Canyons are typical of mountainous regions, but are also found cutting high-elevation plateau (e.g. the Black Canyon of the Gunnison on the Colorado Plateau in Colorado, USA). They occur where stream erosion significantly outpaces weathering. Streams in canyons frequently flow through BEDROCK CHANNELs.

JUDY EHLEN

CAPROCK

Geomorphologically resistant lithological units, which protect underlying less resistant rocks from erosion and denudation, are called caprocks. Tablelands, CUESTAs, MESAs, buttes and hogbacks are examples of landforms which are composed of a backslope (dipslope), supported by a competent caprock, and a bipartite scarp slope. The scarp slope consists of an upper slope in the caprock and a moderately inclined lower slope in the less resistant rock below. It originates from fluvial down-cutting or fault scarp development. In composite landforms like canyons and stepped cuesta landscapes whole sequences of caprocks and soft rocks can be found as in the Grand Canyon, in the Giant Staircase of southern Utah and northern Arizona or in the scarplands of southwestern Germany.

There is no specific method for defining caprock resistance on a metric scale, but there have been attempts to describe the resistance on an ordinal scale (Schmidt 1991). A scarp-forming rock must be relatively more resistant than the underlying soft rock. Caprock resistance can be connected with lithological attributes such as: mechanical hardness protecting the rock against the direct effects of weathering and erosion; porosity and JOINTING resulting in greater water permeability. The attributes of the more resistant scarp-forming rock are most effective when the less resistant rock possesses contrasting characteristics such as mechanical weakness, easy disintegration and low permeability (Ahnert 1998: 239). The effects of different resistance are most visible in dry climates with selective weathering and erosion, where even minor lithological variations are reflected in slope geometry.

In most cases, especially in climates with greater surface water availability, permeability is more important for determining caprock resistance than mechanical strength. Due to their perviousness caprock outcrops are generally characterized by a lack of surface water and low values of drainage density. Water infiltrates and percolates through the caprock body until it reaches the impermeable lower slope rock. At the caprock/soft rock interface it reappears in springs and seepage zones, sometimes connected with sapping processes and slope undercutting. The missing to low activity of surface water erosion on the caprock-protected backslopes reduces mechanisms of denudational downwearing.

Most caprocks are sedimentary rocks like sandstones, conglomerates and limestones. The properties of the cementing materials (carbonates, iron oxides, clay minerals) control the erosion and weathering susceptibility of the sandstones and conglomerates. Karst processes are effective in carbonate caprocks. Joints and fissures are widened by solution resulting in increased permeability. In the scarplands the softer rocks below the caprocks are often clays, marls and fine, densely layered sandstones.

Volcanic rocks can also act as caprocks. This especially occurs in the case of lava flows which moved down former valley floors and covered older sedimentary rocks. The valley slopes in sedimentary material at the sides of the lava flows were subsequently removed by erosion and denudation, and the lava flows, due to their resistance, survived as caprocks of residual hills. This process is called relief inversion (see INVERTED RELIEF). Examples of this geomorphological process combination can be found in Tertiary volcanics in the Ore Mountains in Saxony and in Cenozoic volcanics at the margins of the Colorado Plateau in Utah and Arizona.

Scarp formation is also possible with DURICRUSTs as caprocks. The resistant crusts develop in homogeneous lithological units by weathering and soil-forming processes. Scarps of this kind are called homolithic scarps and are most frequently found in semi-arid areas (calcretes as caprocks) and in the tropics (silcretes and ferricretes). Some rock types can only have the function of caprocks under specific climatic conditions. Gypsum, for instance, can be a caprock in arid regions (e.g. southern Morocco), but in more humid climates rapid sulphate dissolution makes it an incompetent lithological unit.

The geomorphology of the upper scarp slope and the backslope is controlled by the attributes of the caprock. The upper scarp slope has a steep to cliff-like morphology. Its upper end forms a sharp crest (Trauf), especially in horizontally bedded caprocks with vertical joints. Caprock slopes, which are controlled by mass movement activity, also have a sharp crest at their top. In poorly cemented sandstones (e.g. Navajo Sandstone on the Colorado Plateau) rounded upper slopes are developed, which can also be found in more humid and colder areas, where sheet wash or solifluction processes are active on the upper slope.

The backslope begins at the scarp crest and follows the direction of caprock dip. In dry regions with highly selective weathering and erosion the inclination of the backslope is the same as the dip angle. Here the backslopes are stripped surfaces. The overlying less resistant rocks have been removed by erosion and denudation; there is a close conformity of topographic and structural relief. In the scarplands of more humid climates (e.g. Central Europe) the inclination of the backslope is generally less than the dip. The overlying strata are truncated at a low angle in the distal parts of the backslope. It resembles an inclined planation surface, but it is a caprock-controlled structural relief element of the cuesta landscape.

Especially in dry climates there is no mechanism of denudational downwearing of the caprocks. Their erosion is accomplished by aquatic and gravitational processes, which work on the scarp slope and result in lateral backwearing by parallel scarp recession. The area of the caprock outcrop is consumed by scarp retreat, at the margins of residual outliers by circumdenudation. The rate of retreat is controlled by caprock lithology and thickness (Schmidt 1989). It is not surprising that the backslope length is also controlled by these variables, and additionally by structural dip and the lithology of the overlying rocks (Schmidt 1991).

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SEE ALSO: butte; cuesta; hogback; mesa; structural landform

KARL-HEINZ SCHMIDT

CASE HARDENING

Case hardening describes rocks with outer shells more resistant to erosion than interior material. This hardening is sometimes called induration. Although case hardening is occasionally used as a synonym for DURICRUST, case hardening most frequently refers to differential weathering of the same rock type – often associated with intricate weathering features such as tafoni (Campbell 1999).

Two general types of processes create the appearance of case hardening: core softening of the interior; and case hardening of the exterior. James Conca proposed a lithologically based explanation. Crystalline rocks such as granite tend to core soften, whereas clastic rocks such as sandstone tend to case harden. The dichotomy has to do with the way the minerals bond together. Since sandstone grains are held together by cementing agents, a greater accumulation of cements at the surface causes case hardening. In contrast, the greatest change in hardness in a crystalline rock takes place when bonds are broken by CHEMICAL WEATHERING. Core-softened boulders are seen in locales of most intense chemical weathering.

The early literature advocates the view that hardening occurs by solutions that are mobilized from the rock's WEATHERING rind, drawn out by evaporative stresses, and then reprecipitated in the rock's outer shell. A growing body of evidence indicates that external agents also penetrate into the outer shell of the host rock, hardening the surface. A variety of different hardening agents have been found within host rocks lacking these agents. Amorphous silica, calcite, calcium borate, clay minerals such as kaolinite, iron hydroxides, oxalate minerals, rock varnish and other internally and externally derived agents penetrate about a millimetre to harden the very surface of the rock.

Case hardening, by definition, is not a ROCK COATING. However, a wide variety of rock coatings can act as case-hardening agents. Glazes of mostly silica and aluminum with some iron, only 20–30 μm thick, impede erosion of greenschist in southern England (Mottershead and Pye 1994). The role of silica glaze can be striking for temperate sandstones: '[o]ne of the most important characteristics of many porous sandstones is their tendency to case-harden owing to the development of a surface crust or rind' (Robinson and Williams 1994: 382). In Antarctica, iron-stained silica glaze reduces permeability and channels moisture towards uncoated rock surfaces. Thus, rock weathering is concentrated away from the rock coatings (Conca and Astor 1987). Lichen-generated oxalates protect sandstone surfaces in the Roman Theatre of Petra. Dark coatings of silica, oxides of iron/manganese, and charcoal case hardened rock faces at Yarwondutta Rock, Australia (Twidale 1982). While lichens are usually erosional agents, these epilithic (rock surface) organisms sometimes protect the underlying rock from erosion.

Although case hardening is most commonly noted in warm deserts where little soil covers rock surfaces, case-hardened rocks occur in all terrestrial weathering environments. In the wet tropics, for example, case hardening is frequently seen on bedrock along rivers at stages only reached by wet-season floods. In alpine settings, case hardening helps preserve glacial polish. Silica glaze is an important case-hardening agent in temperate (Mottershead and Pye 1994; Robinson and Williams 1994) and Antarctic areas (Conca and Astor 1987). Iron films can be seen splitting apart and also holding together weathering rinds in northern Scandinavia (Dixon *et al.* 2002).

Case hardening (Plate 20) often has subsurface origins in JOINTING. Mottershead and Pye (1994), for example, discerned a three-stage process. First, the host rock hardens along joint faces within the subsurface. Silica, aluminum and some iron comprise the bulk of the case-hardening agent. Second, DENUDATION of the land surface exposes joint faces at the surface. Third, erosion of rock underneath the case-hardened surface creates cavities called TAFONI, that highlight the case hardening. Rock engravings (petroglyphs) also emphasize planar JOINTING surfaces that were case hardened while in the subsurface. Road cuts of granitic rocks that weather to GRUS often reveal case-hardened subsurface joints. Geothermal and other DIASTROPHISM processes often leave behind case-hardened joints.

Considerable disagreement exists over how long it takes case hardening to form, with assertions in the literature ranging from months to thousands of years. James Conca studied rates of hardening in the Mono Basin of eastern California, finding that changes take place on the timescale of thousands to tens of thousands of years. Rates of hardening, however, vary with climate and the particular hardening process.

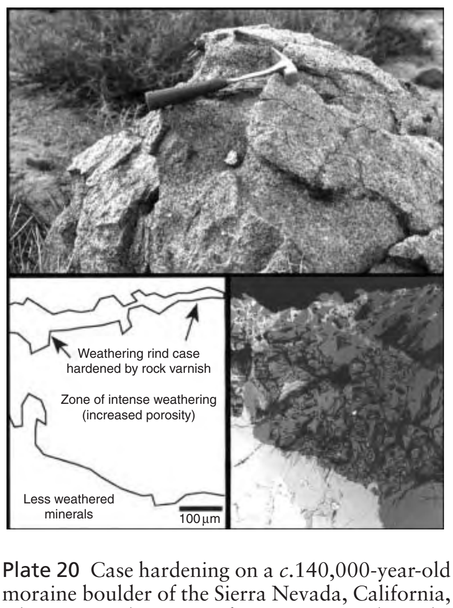

Plate 20 Case hardening on a *c*.140,000-year-old moraine boulder of the Sierra Nevada, California, where a combination of processes produce the differential weathering seen in the top image. Core softening of the host granodiorite boulder is the most important process. The electron microscope image and corresponding map shows a close-up of the area around the tip of the rock hammer. Some softening comes from chemical weathering and some hardening takes place as a result of the penetration of desert varnish into the weathering rind

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SEE ALSO: chemical weathering; denudation; grus; jointing; rock coating; tafoni; weathering

RONALD I. DORN

CATACLASIS

A natural process whereby a faulted rock is deformed as a result of mechanical forces in the crust, such as fracturing, shearing, grain boundary sliding, and granulation. Cataclasis transforms a simple fault into a zone of fracturing and deformation without chemical alteration, and causes a decrease in the porosity of the rock alongside rock volume. Cataclasis takes place at low temperature–low pressure conditions, and high strain rates. The product of cataclasis in sediment is a cataclasite, a metamorphic rock composed of angular fragments (e.g. tectonic breccia) and a structureless rock powder fabric. When considered on a regional scale, cataclasis has also been interpreted as a flow mechanism.

STEVE WARD

CATACLINAL

A dip stream or a valley that runs in the same direction as the dip of the surrounding rock strata. Cataclinal slopes can be further classified into overdip slopes, underdip slopes, and dip slopes (steeper than, shallower than, and following the dip of surrounding strata, respectively). They may follow an individual rock layer from the base of a mountain to its peak (e.g. Mount Rundle, Canadian Rockies) (Cruden 2000). In contrast, anaclinal slopes dip in the opposite direction to the surrounding strata.

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STEVE WARD

CATASTROPHISM

Catastrophes in common parlance are unexpected major events with negative outcomes affecting large numbers of people. Large earthquakes, floods, hurricanes and wars are good examples of catastrophes, not all of which have geomorphic relevance. In the slightly more technical language of natural hazards, catastrophes are high magnitude, low frequency geophysical events which impact the socio-economic environment negatively. At a third level of technicality, catastrophes can be defined in the context of a mathematical approach to the analysis of inherently unstable Earth surface systems. This mathematical approach is called bifurcation theory and a special branch of bifurcation theory has been called catastrophe theory (Thom 1975). The essential concept is that many Earth systems are inherently unstable and are called dissipative structures. Such systems are characterized by a series of thresholds, below which the solutions to equations governing system dynamics are unique; beyond the threshold (or bifurcation point) the system loses its structural stability and undergoes a sudden or catastrophic change to a new form. Examples in the geomorphic literature include stream junctions in the Henry Mountains, Utah (Graf 1979), sediment transport processes in rivers (Thornes 1983) and accretion–degradation processes on a beach (Chappell 1978).

Catastrophism is a mode of thought which was common in the eighteenth and early nineteenth centuries and had as its basic premise that Earth history consisted of a series of high magnitude events separated by periods of quiescence. This mode of thought contrasts with gradualism in which small-scale, commonly acting processes are thought to be the dominant mode of geomorphic evolution.

The origins of catastrophism have often been traced to Baron Georges Cuvier (1769–1832) but, as we shall show below, his was only one brand of catastrophism and there were many catastrophist predecessors. Cuvier was the father of comparative anatomy and proposed that the Earth had suffered many catastrophes in the form of global earthquakes, each of which had changed the global landscape and annihilated almost all the flora and fauna. After each catastrophe, a new set of flora and fauna appeared: ‘Thus we shall seek in vain among the various forces which still operate on the surface of our Earth, for causes competent to the production of those revolutions and catastrophes of which its external crust exhibits so many traces’ (Cuvier 1817: 36–37). Huggett (1990) summarizes the issue by saying that there is either abrupt and violent change (catastrophism) or gradual and gentle change (gradualism). But the issue is more complex. There are different styles of catastrophism, and the style is determined by the premises of the author with respect to the following dichotomies: actualism v. non-actualism; directionalism v. steady state and, in dealing with organic change, internalism v. externalism. Although dichotomies tend to oversimplify the situation and in reality theorists of the Earth occupy positions along a spectrum of ideas, some simplification is necessary within the word constraints of this encyclopedia entry. The issue between actualists and non-actualists is whether past processes have differed in kind from those now in operation; non-actualists took the view that past processes could differ in kind from those presently in operation. The issue between directionalists and steady statists has been eloquently discussed by Gould (1987) in terms of time’s arrow versus time’s cycle; directionalists took the view that monotonic change, whether progress or regress, could be detected in Earth history. The issue between internalists and externalists, which arose only in the context of organic change, was whether the motor of organic change was an inner drive or external environmental factors; externalists argued that the environment forced change. Using these dichotomies as a basis for classification of styles of catastrophism, Huggett (1990) came up with eight categories of catastrophism which could be distinguished in the early history of environmental science. Six of these categories are reproduced here:

1 *Actualistic directional catastrophism* The Wernerian system of Earth history, following Abraham Werner (1749–1817), is a classic example of this kind of catastrophism. He envisaged five periods of Earth history, punctuated by intermittent and catastrophic ocean subsidence and precipitation of crustal rocks. The five periods were consecutive and demonstrated directional change.
2 *Non-actualistic directional catastrophism* René Descartes (1596–1650) described the origin of the Earth as an incandescent ball, followed by collapse of the Earth's outer crust and the release of massive volumes of water. His system involved directional evolution from original chaos created out of nothing by God through to an ordered universe which evolved according to natural laws invested in the original particles by God. The so-called Scriptural geologists, such as William Buckland (for most of his active career), Adam Sedgwick (1785–1873), William Conybeare (1787–1857) and Robert Murchison (1792–1871) all fell into this category in the sense that they believed in God's special intervention in the regular course of Nature through geological catastrophes and the sudden rise of species.
3 *Non-actualistic steady-state catastrophism* Baron Georges Cuvier (1769–1832) was the leading protagonist of this school of thought. He saw in the fossil and stratigraphic record evidence of catastrophic changes too great to be explained by the ordinary, slow-acting processes on the Earth's surface. Each catastrophe had changed the global landscape and a new set of plants and animals had appeared, with no particular connection with the previous flora and fauna.
4 *Inner-driven directional catastrophism* Louis Agassiz (1807–1873) espoused this position for most of his career. The underlying premise is that organisms have an immanent quality leading to progressive but discontinuous change. The progression of life was the unfolding of God's plan through catastrophes in the inorganic world and punctuations in the organic world, but with no causal connection between the two.
5 *Environmentally driven directional catastrophism* The mature William Buckland was a proponent of this position, that the Earth had suffered a series of catastrophes and that a new set of species was created after each mass extinction. Each new creation was an improvement on the previous one and the improvements placed the organisms in better harmony with the changed environment.
6 *Environmentally driven steady-state catastrophism* Baron Georges Cuvier was the leading advocate of this position. He could not accept a progression of organisms. He recognized four chief branches of animals, the members in each of which were fixed and designed to meet all environmental conditions. His catastrophes were essentially sudden environmental changes in the distribution of land and sea. The motor of biotic change is sudden environmental change.

The other two of Huggett's categories, the actualistic and inner-driven steady-state catastrophisms, were rarely expressed positions in the early nineteenth century but became more popular in the late twentieth century with the rise of neocatastrophism.

Since the 1960s, one of the key assumptions of catastrophism has been making a strong comeback in the form of neocatastrophism. This key assumption is that high magnitude, low frequency events are cumulatively more important in Earth history than low magnitude, high frequency events. Some reasons for this changed perspective are:

(1) Improved precision in geochronology has demonstrated unexpectedly rapid past changes.
(2) The exploration of mass extinctions in the past has intensified.
(3) Some geomorphological features, such as the channelled scablands of eastern Washington, are more amenable to explanation by low frequency, high magnitude events than by gradual, semi-continuous processes.
(4) Space exploration has generated a strong interest in galactic-scale events.
(5) Interest in global environmental change has provided evidence of rapid past changes, such as found in the polar ice caps and the oceanic deep sediments.
(6) The rise of non-linear dynamics and chaos theory is beginning to provide ways of synthesizing gradualism and catastrophism.

It seems self-evident that Earth history contains a combination of catastrophic and gradual events; accepting the occurrence of catastrophes is not to deny the effectiveness of gradual processes. The main reason for concerns about catastrophism expressed by Earth scientists since the mid-nineteenth century has been a fear of the reintroduction of religious beliefs into the canon of modern science because of the long-held views about the Noachian flood in western thinking. Largely for this reason, the geological establishment of the day refused to countenance the work of Harlen Bretz (1923) in his account of the origins of the channelled scablands of eastern Washington. He suggested that these SCABLANDs (massively and regionally gullied lands) could be explained best by the action of a single large flood over a period of only a few days. The fact that they have been shown subsequently to be caused by a succession of pulses associated with the draining of glacial Lake Missoula (Baker 1973) was complete vindication for Bretz. It is important to recognize that in no sense did the catastrophists violate the principle of uniformity of law (i.e. that natural laws are invariant in time and space). On the other hand, the non-actualists did violate the principle of simplicity, a principle which states that no unknown causes should be invoked if available processes are adequate. This guideline is known as the uniformity of process.

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SEE ALSO: actualism; neocatastrophism; uniformitarianism

OLAV SLAYMAKER

CATENA

The concept of a catena is one of linkage – catena being the Latin for chain. So, by analogy individual elements are linked in some way and have something in common. In the case of a soil catena, the soils are linked by having the same parent material and age but are differentiated by their position on a slope which alone gives them different characteristics, especially in relation to drainage. Thus a formal definition is: 'A sequence of soils of about the same age, derived from similar parent material, and occurring under similar climatic conditions, but having different characteristics because of variations in relief and in drainage.'

A more generalized term which can be used is the toposequence, which refers to any sequence of soils which varies with topography. A more extensive, related concept is soil association, which is any repeated pattern in the landscape which may not necessarily be repeated with topography, as with a catena, but could be linked to geology, geomorphology and indicated by vegetation in any geomorphic element of the landscape.

The term soil catena was first proposed by Milne (1935a,b; 1936) for topographically linked soils in East Africa. On the sides of large valleys different soils were found on the upslope crest, the lower downslope and the footslope. Milne felt that the profiles of the soil types changed character downslope according to both drainage and the past history of the land surface. Here, there is an essentially uniform lithology throughout the slope and the differences derive from the shedding of moisture upslope and wetter conditions downslope as well as the movement of solutes in that water and the physical movement of eroded particles and their accumulation downslope.

Milne felt that the upper soils might also be older and more remnant and that the downslope soils might be younger, with a fresher accumulation of deposits, and also that the rocks might actually differ downslope. Ruhe (1960) further expanded on this by differentiating between the 'classic' catena on similar parent materials and a sequence which could be formed on two or more geological formations where the lithology actually varied. Here the downslope series is still linked by drainage and transport of material and shares a common physiographic history and geomorphic evolution despite being on different parent materials.

The catena concept is useful in soil mapping because it implies a regularly occurring relationship of the soils with topography, giving an expectation or prediction of what is liable to be present. An example of a general schema is in Table 6. With mechanical movement of particles, the upslope soils might be more coarse grained, having lost fines downslope, while the downslope soils could be more alluvial, with the accumulation of fines and/or with much coarser particles accumulating, according to the amount of mechanical action on the slope. Generally the midslope soils tend to be prone to erosion and are much thinner than the slope foot, where accumulation occurs.

The more complex and nutrient-rich montmorillonite clays may be found where nutrients accumulate at the base of the slope. In warmer areas, montmorillonite clays also survive on the soil crest and plateau but in wetter, cooler areas, the simpler kaolinite clays are found in these positions due to greater leaching. In hot, humid or semi-humid regions, the upslope leached soils are often red due to OXIDATION and at the slope foot where drainage is impeded the colour changes downslope through yellow to grey. The effect of topography can thus be expressed in the tropics with high rainfall where red soils with kaolonitic clays are found in the better drained upslope areas with montmorillonite and black, organic soils in the less well-drained depressions. On a larger scale, which includes tropical mountains, in areas of highland surrounded by lower arid land, the sequence with decreasing height is one of upland podzols and brown earths to chernozems to desert and semi-desert soils; for uplands surrounded by lower land with high rainfall the downslope sequence is one of podzols and brown earths to yellow, red and black soils of the humid lowlands.

The significance of a catenary sequence for agriculture and land-use planning is again predictive in that more drought resistant crops can be grown on the upper slopes and moisture-tolerant crops on the lower slope where the aerated zone is nearer the surface.

Hillslope hydrologists have also used the catena concept when devising predictive models for hillslope runoff generation. Here hydrological processes, such as infiltration, overland flow and throughflow, can be predicted in relation to soil profile properties which can be predicted to vary systematically downslope (McCaig 1985). Anderson (1985) also uses the concept of catena in predicting the mechanical, load-bearing properties of soils. Regression modelling has been attempted to quantify the relationship between slope position and soil properties with varying degrees of success (Furley 1971).

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Table 6 The relationship of soils and topography

	Wetter, cooler climates	Wet climates	Drier climates
Upslope/ slope crest	Peat, podzolization	Podzolization	Leached soil
Midslope	Brown earth	Brown earth	Non-calcareous
Downslope	Peat, gley	Gley	Calcareous soil

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STEVE TRUDGILL

CAVE

The standard definition of a cave is 'an underground opening large enough for human entry' (Oxford English Dictionary and others). As such, natural caves occur in most consolidated rocks or compacted sediments and in most geomorphic settings, created by a variety of processes. This entry focuses upon KARST caves, which are the most important in terms of their magnitude, frequency, diversity of form, and role in general geomorphology. They are formed where dissolution is either the quantitatively predominant or essential trigger process of rock removal. Karst caves thus are restricted to the comparatively soluble rocks. In descending order of solubility, these are salt, gypsum and anhydrite, limestone and dolostone and (to a much lesser extent) quartzites, calcareous and siliceous sandstones. Limestone caves display the greatest range in size and form. Although 'enterable by humans' is the criterion, initial solution caves are much smaller (~1 cm diameter is a reasonable minimum) and the rules of genesis do not change significantly as they are enlarged to enterable size.

Caves created with little or no dissolution are PSEUDOKARSTic. CAVERNOUS WEATHERING voids can be considered transitional. Many pseudokarst caves are created by mechanical processes, including piping (see PIPE AND PIPING), THERMOKARST collapse, frost riving, wave action and CAVITATION along coasts. Such caves are rarely longer than a few tens of metres; most do not pass out of the range of daylight. Stream melting and sublimation in GLACIER ice and firn creates greater caves that are nearly identical in form and scale to the vadose shafts, canyons and simple phreatic passages found in many dissolutional caves. The lengthiest pseudokarst caves are lava tubes (see LAVA LANDFORM), formed by channelled discharge of still-molten lava within consolidating flows; single tubes, dendritic networks and anastomosing mazes are known, some extending for 10 km or more (Gillieson 1996).

Karst caves

Modern classification (Klimchouk *et al.* 2000) recognizes three principal genetic settings: (1) coastal caves in young carbonate rocks; (2) hypogene caves, formed by waters ascending out of artesian traps ('confined groundwaters') in any soluble rock; (3) unconfined meteoric water caves in soluble rocks, the most abundant and significant class. Figure 18 shows a basic range of plan patterns in these caves, relating them to type of recharge and the most effective (transmissive) porosity existing at the onset of dissolution. In most known hypogene and unconfined caves intergranular ('matrix') porosity is low, (<5 per cent), solvent water being transmitted via penetrable bedding planes, joints and faults which control the loci of the solutional passages. The relevant chemistry and kinetics of aqueous dissolution are discussed in Ford and Williams (1989: 42–126); Klimchouk *et al.* (2000: 124–223).

EOGENETIC COASTAL CAVES

'Eogenetic' describes very young limestone and dolostone accumulations where consolidation by compaction and interstitial cementation (i.e. diagenesis) is still limited, with the consequence that intergranular porosity offers principal or, at least, significant routes for solvent water flow. Such rocks are found in tropical/subtropical coastal settings today, e.g. Florida, Yucatan, Bahamas, many Pacific atolls, etc., and are chiefly Pleistocene in age. The matrix porosity yields cave patterns similar to those of cavernous weathering in non-karst strata.

'Syngenetic' caves (Jennings 1985) form in calcareous sand dunes when surficial sand becomes case hardened by cementation (i.e. earliest diagenesis), following which storm waves or surface streams breach the casing and wash out the non-cemented sand behind it, creating cavities sometimes many metres in length or height. This is one end of the spectrum of karst caves, where mechanical washout (piping) is quantitatively predominant but dissolution and cementation must precede it.

Much more widespread are caves formed where fresh and salt waters mix along the water table at the coast itself (flank margin caves) or at the halocline beneath the freshwater lens further inland (Klimchouk *et al.* 2000: 226–233). Flank margin caves display large entrance chambers dividing and tapering to blind endings a few metres or tens

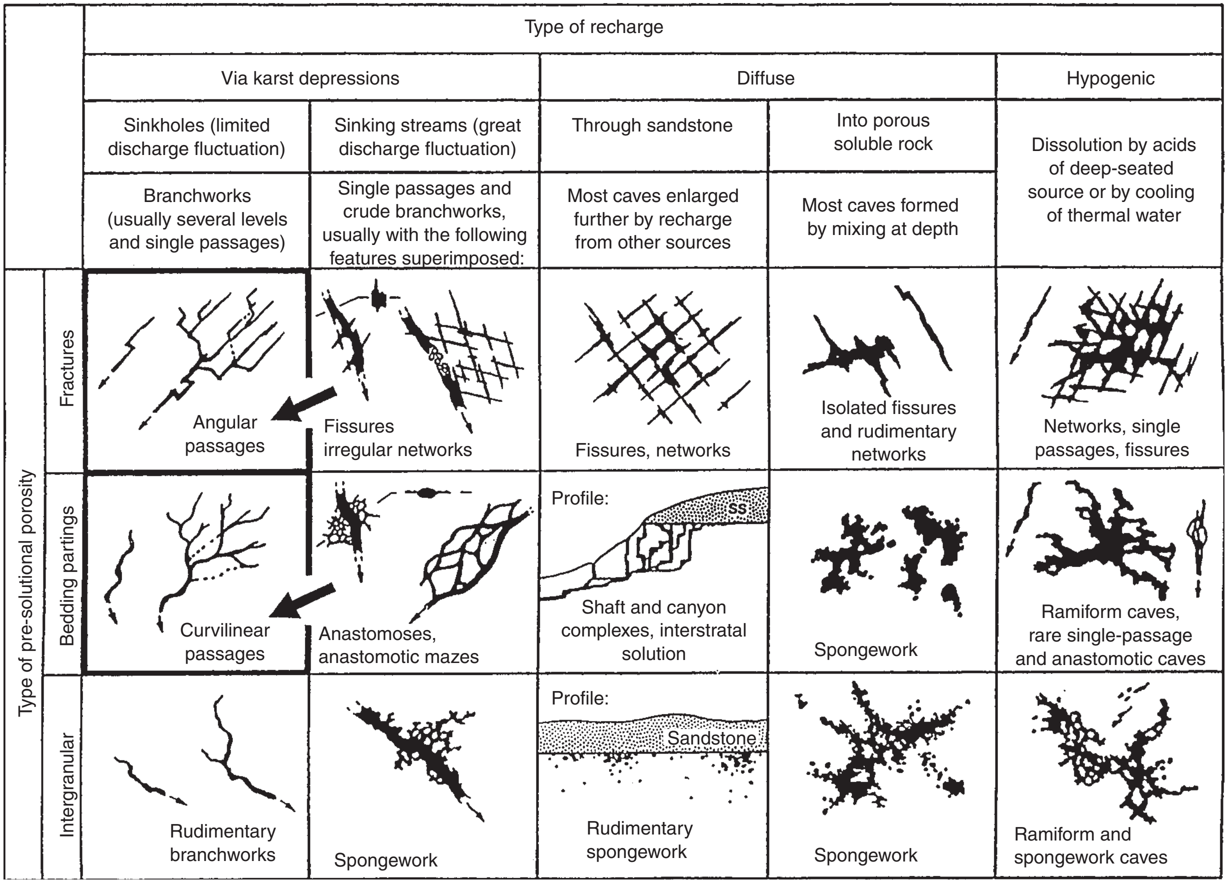

Figure 18 Basic plan patterns of caves shown in relation to types of pre-solutional porosity and conditions of recharge; slightly modified from Palmer (1991)

of metres inland. Halocline caves have more complex spongework patterns, in part due to the shifting of salt/fresh mixing zones as Quaternary sea levels moved up and down; aggregate passage lengths as great as 1 km are rare.

Cave systems many tens of km in length and extending 5 km or more inland are being explored along the Caribbean coast of the Yucatan Peninsula; although in young limestones, they are of unconfined meteoric origin, modified in form by the salt waters that now inundate them.

HYPOGENE CAVES

In hypogene caves the waters may have circulated deeply within karst strata alone (due to synclinal or graben-type structural traps), or be ascending into the karst rocks from underlying, non-karst aquifers (interstratal flow). They may be thermal (>4 °C warmer than mean temperatures in the rock to be dissolved) or ambient.

A majority of hypogene caves are excavated under phreatic conditions, i.e. beneath any water table. The most simple form is a vertical/near-vertical chimney on a fracture, up which the water flows to discharge at a surface spring or into overlying, more porous, strata. Active instances include thermal springs in Mexico from shafts 40+ m in diameter and plumbed to −300 m. Deeper examples are known, but drained by uplift and erosion; some contain economic minerals precipitated on the walls, e.g Tyuya Muyun, Kazakhstan. More complex in form are arborescent chimney caves, branching upwards from basal reservoir chambers; Satorkopuszta Cave, Hungary, is a spectacular instance with later spheroidal rooms of condensation corrosion origin branching from an original phreatic shaft (Klimchouk *et al*. 2000: 292–303).

Fracture-guided network caves (Figure 18) are common. In western Ukraine local meteoric waters passing up through a ~14 m stratum of gypsum from an underlying sand aquifer created joint-guided mazes with intersections every 2–5 m; 212 km of contiguous passages are mapped in Opitimists' Cave. More complex are multistorey rectilinear mazes in the Black Hills, South Dakota, where thermal waters converged on Carboniferous palaeokarst (see PALAEOKARST AND RELICT KARST) preserved under clastic strata. The waters discharge through weaknesses in the clastics today, enlarging their routes and lowering the water table in Wind Cave 14 km distant at 40 cm k yr^{-1}. Jewel Cave, ~40 km distant, is fully drained; its 200 km of mapped passages are crusted with 10–20 cm of calcite spar deposited as the waters declined. Large-scale groundwater invasion and dissolution of limestones, gypsum and salt such as this, but deeply buried under later rocks, is associated with formation of solution breccias hosting oil and gas, lead/zinc and other mineral deposits, or creating breccia pipes that can stope upwards through 1,000+ m of overlying rocks. The greatest reported hypogene cavity is in Archean-Proterozoic marbles of the Rhodope Mountains, Bulgaria. It has an estimated volume of 237 million m^3 and a roof-to-floor depth believed to exceed 1,340 m (Klimchouk *et al*. 2000: 304–306).

A very distinctive type is the cave formed by sulphuric acid from H_2S. In most known instances the gas migrated from adjoining coal or oil basins where it was produced by bacterial reduction of gypsum. Reaching carbonate rocks it oxidized to H_2SO_4 at and just below the water table. Small H_2S caves tend to be linear outlets to springs. Large caves ramify about the gas inlets (Figure 18). Big chambers are created by lateral corrosion at the water table plus condensation corrosion above it that may convert limestone walls to gypsum to depths as great as one metre. Shift of inlet points and lowering of springs leads to multilevel development of spectacular systems such as Lechuguilla Cave, New Mexico (172 km, ~480 m in depth; Widmer 1998).

Unconfined caves

These are the principal type known to explorers and geomorphologists. The caves extend from surface water input points such as KARRENfields, DOLINEs, river sinks or POLJEs at the karst margins, to springs that are lower in elevation. Although it is rare for cavers to be able to follow the water all the way, dye tracing and other analysis invariably confirms that dissolutional conduits are continuous between sink and spring and regulate the flow. The most simple caves are single dissolutional pipes between sinks and springs. There are many instances in underground meander cut-offs or river short-cuts across narrow horsts or anticlines. However, the majority of caves have multiple inlets that, in plan view, link up to form crudely dendritic patterns. These are angular where joints are dominant controls and sinuous in bedding planes; many caves exhibit mixtures of the two. Joint mazes and bedding plane anastomoses

(Figure 18) are usually subsidiary components in the dendritic plans, formed where there is rapid flooding at stream sinkpoints or underground blockages. The published plans of many cave systems appear more complicated than these simple combinations because the systems are multiphase; relict passages from higher levels cross those that are still active Modelling of plan pattern genesis is quite advanced, (see Ford and Williams 1989: 249–261; Klimchouk *et al.* 2000: 175–223).

Cave morphology in long section (i.e. length × depth) is closely related to geologic structure and groundwater zonation. In young mountain terrains if karst aquifers are thick and rapid uplift opens vertical fractures widely, caves are sequences of shafts down the steepest fractures, linked together by short, sinuous, stream canyons. Gravitational control of flow is predominant. A majority of the world's caves deeper than 1 km gain most of their depth in this upper vadose zone. Voronja Cave, Caucasus – currently the explored depth record holder at −1,710 m – is a fine example.

Where the karst formations are relatively thin and/or were little stressed during uplift, such conditions may never have existed. Instead, the uppermost zone is waterfilled initially but drains progressively as caves propagate through it and become enlarged – the 'drawdown vadose zone'. The caves display initial phreatic features such as elliptical cross sections in bedding planes, but with subsequent gravitational entrenchments beneath them. This combination can be found locally in young mountains also where passages are perched on shale bands or other obstructions. Entire cave systems, from sinkpoints to springs, can develop wholly within these two vadose settings, especially where karst strata are perched on insoluble rocks above regional base levels. There can be deep gravitational entrenchment into the insolubles; e.g. some 'contact caves' in Greenbrier County, West Virginia, have 90+ per cent of their volume in erodible shales beneath limestones that hosted the initial passages.

Most extensive cave systems have substantial water table or phreatic (sub-watertable) sections, however. Their length is usually greater than that of the vadose parts, and the geometry more complex and varied. There are four basic possibilities ('four state model'; Ford and Williams 1989: 261–274). If the density of penetrable, interconnected fissures is very low, geologic structure may compel the conduits to follow courses below the elevation of the springs or water table ('phreatic loops'). A State 1 system passes from the vadose zone to the spring in one loop. State 2 is a sequence of loops whose crests fix local elevations of the water table. Where fissure frequency and interconnection are greater, caves display mixtures of loops with gently graded passages at the water table (State 3). Very high frequency permits continuous development along the water table (State 4), similar to flank caves in young, porous limestones. There is probably phreatic looping to depths of 1,000+ m in some State 1 caves. Individual loops greater than 250 m deep are common in State 2. In regions subject to large magnitude, abrupt flooding such as alpine mountains, overflow ('epiphreatic') passages develop above the low stage conduits (Audra 1995).

Multi-level (multi-phase) caves

Most extensive caves have two or more 'levels' that developed to drain to successively lower springs – 'level' denoting the historical succession but not implying that the galleries must be near-horizontal; often, there is State 2 or 3 geometry. In vadose caves the lower levels may be simple entrenchments beneath the older passages. Where there is water table or phreatic development, the new springs are usually offset laterally tens to hundreds of metres or more and may have distributaries. The new springs steepen the hydraulic gradient in the downstream section of the old cave, which then adjusts to its new 'level' in a sequence of breakthrough undercaptures (French – *soutirages*) that move the hydraulic steepening progressively upstream like a river knickpoint. Portions of individual capture links are incorporated into the final dendritic pattern of the new level but others are left redundant, becoming drained relicts or silted backwaters.

Superimposition of successive levels, redundant links and invasion vadose caves from new sinkpoints, make maps of great systems such as Mammoth Cave, Kentucky (556 km – the longest mapped) more complex in appearance than almost any other geomorphic or hydrogeologic phenomena.

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DEREK C. FORD

CAVERNOUS WEATHERING

Cavernous weathering is a process which causes the hollowing-out of rock outcrops and boulders on vertical and near-vertical faces. The hollows, or caverns, may take one of two forms. The first are known as 'tafoni' (singular: tafone), a term derived from the Sicilian word meaning 'windows'. TAFONI typically conform to a spherical or elliptical shape, have arched-shaped entrances, concave inner walls, overhanging visors and gently sloping debris-covered floors. Tafoni range in size from several centimetres to several metres in diameter and depth, may coalesce or interconnect and second order tafoni may develop on the backwalls of larger forms. The second form caverns may take is commonly called 'alveoli' (singular: alveole). Alveoli are formed by similar processes, termed HONEYCOMB WEATHERING, which involves the progressive development of closely spaced cavities in rock faces. These small-scale caverns are separated by narrow, intricate walls, creating a surface reminiscent of honeycomb. Alveoli are usually several centimetres in diameter and rarely are larger than one metre wide. The relationship between alveoli and tafoni has not been clearly defined and the distinction is therefore one of size and shape. Although cavernous and honeycomb weathering frequently occur together, their independent existence and differences in form have led some geomorphologists to contend that they are not genetically related, but rather derive from different modes of origin (Mustoe 1982). Cavernous weathering cannot be defined on the basis of geographical, lithological or climatological occurrence, as caverns may be found in many environments and on most rock types.

Caverns cannot be categorized on the basis of their occurrence on specific rock types or in specific climate regimes, as they occur under cold, temperate, hot, humid and arid environments, and are found on a variety of rock types. Cavernous weathering was previously considered to be a diagnostic feature of arid environments (Blackwelder 1929). Tafoni are most prolific in salt-rich environments, and have been documented most often in deserts (McGreevy 1985) and coastal zones (Mottershead and Pye 1994). Common factors in this disparate range of environments are high salt concentrations and frequent or occasional desiccating conditions. The occurrence of tafoni on sandstone surfaces (Young and Young 1992), and on granite surfaces (Dragovich 1969) has frequently been documented. Aside from these siliceous rock types, tafoni have also been recognized on tonalite, dolerite, lacustrine silts and conglomerates. A range of weathering processes may be responsible for the occurrence of tafoni, and there is clearly convergence of form. Considerable literature has accumulated on the nature of both tafoni and alveoli, but as more information has been presented, their possible origins, rather than being clarified, seem to have become more confused.

There are two types of tafoni: 'basal' and 'sidewall' tafoni. Basal tafoni are often found, as the name suggests, on outcrops and boulders at ground level. Sidewall tafoni are present on vertical and near-vertical outcrop surfaces where strong rock discontinuities are not present. Tafoni which develop along discontinuities, above ground level, may also be considered to be basal tafoni, and are characterized by a higher rate of back weathering than upwards progression.

Early studies of cavernous weathering suggested that caverns were created by the action of the wind, which was also responsible for the removal of weathering products. It is now widely accepted, however, that aeolian deflation is not responsible for the hollowing out of boulders or pitting of rock faces. Disintegration of cavern walls generally proceeds by flaking and granular disintegration, and numerous weathering processes have been invoked, including insolation weathering, WETTING AND DRYING WEATHERING, frost weathering (see FROST AND FROST WEATHERING), solution and chemical alteration of rock minerals, and SALT WEATHERING.

CHEMICAL WEATHERING processes are considered to be important in the development of tafoni in some circumstances. Tafoni in sandstone appear to be partly the result of the reaction of water and organic acids with iron and silica. Caverns in limestone are produced by a solution of calcium and magnesium carbonates; chemical weathering of dolerite and tonalite has been identified in caverns. Other chemical weathering processes which may have contributed to cavernous weathering include solution, HYDROLYSIS and hydration reactions, induced by microclimatological differences between caverns and exposed rock surfaces.

Of all the processes which may create caverns, salt weathering is the most commonly invoked. The importance of salt weathering is indicated by the presence of salt crystals on walls of coastal and desert tafoni and alveoli. Salts are evident in seepage, in granular debris and flakes being detached from cavern walls, in floor sediments and in crevices within caverns in many locations, testifying to the role of salt crystal growth in cavern development. Salt crystallization, hydration and thermal expansion may contribute to disaggregation, but given the wide geographical, climatological and lithological range of tafoni and alveoli, it is more likely that several weathering processes are involved in cavernous weathering.

A common feature of cavernously weathered surfaces is a case-hardened layer on exposed rock surfaces, penetrated by tafoni. Some researchers conclude that the presence of a hardened crust and weakened interior is a fundamental reason for tafoni occurrence (Conca and Rossman 1982). Conca and Rossman (1985) postulated that the presence of a case-hardening cement is of secondary importance compared to core softening, a result of differential weathering rates between the rock interior and exterior. Many other examples of tafoni in the absence of case hardening have been presented, however, so it appears that a hardened outer crust is not a prerequisite for tafoni formation.

A problem that has beset studies of tafoni weathering has been the tendency to relate it to one single formative process, whereas in many cases cavern development can only be satisfactorily explained by invoking the operation of a range of weathering mechanisms. There may be processes which are active in all cases, but it is likely that the relative importance of physical and chemical weathering processes will vary with different environmental conditions, which may operate under different catalytic conditions, and may act synergistically.

The relative significance of lithological controls is unclear in the context of tafoni origin. Caverns may be initiated along pre-existing joints or bedding planes, or may be distributed randomly across rock surfaces. This random (or pseudo-random) distribution may reflect points of mineralogical weakness on the rock surface, but this idea cannot be tested as the initial surface has been lost in the creation of the hollow.

Once initiated, the sheltering afforded by caverns may provide temperature ranges which are less extreme than on rock surfaces exposed to direct insolation, and higher relative humidities, a significant factor influencing the deposition, absorption and evaporation of moisture on rock surfaces. Microclimatological differences created in shadow zones of tafoni may accelerate the effects of weathering processes. Conversely, the exposed surfaces may shed moisture and solutes more rapidly, creating a negative, or self-regulating, feedback in the weathering system. Cavernous weathering may therefore represent the response of the weathering system to dynamical instabilities, where the positive feedback produced by material loss encourages accelerated weathering within caverns and the system responds by divergent evolution of the rock surface into hollows and exposed stable surfaces.

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SEE ALSO: case hardening; weathering

ALICE TURKINGTON

CAVITATION

A form of erosion that can occur in rapidly moving water. Local areas of low pressure may be created in the water and as Drewry (1986: 68) has explained:

> If the pressure falls as low as the vapour pressure of the water at bulk temperature, macroscopic bubbles of vapour (cavities) will form. The cavitation bubbles grow and are moved along in the fluid flow until they reach a region of slightly higher local pressure where they will suddenly collapse. If cavity collapse is adjacent to the channel wall localized but very high impact forces are produced against the rock. This action may give rise to mechanical failure of the channel.

The destructive action of cavitation is probably due to the shock waves created when the bubbles collapse. Cavitation is of great significance in the malfunction of hydraulic machinery (e.g. turbine blades, ships' propellors, etc.) but its geomorphological effects may also be substantial (Barnes 1956). Sufficiently high velocities for its operation can occur in such situations as waterfalls, rapids, bedrock channels, beneath glaciers and on TSUNAMI scoured surfaces (Aalto *et al*. 1999). Mean flow velocities necessary to initiate the process are generally higher than about $10\,m\,s^{-1}$ for flow depths greater than about 4 m. Cavitation may contribute to the fluting and potholing of massive, unjointed rocks in bedrock channels (Whipple *et al*. 2000).

The term cavitation has a second and unrelated meaning in geomorphology, namely the formation of cavity at the bed or a sliding glacier. Their formation can enhance basal sliding (Lliboutry 1968).

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A.S. GOUDIE

CAY

Cays are the general terms for islands which develop on CORAL REEFs but can be divided for geomorphological purposes into true cays and *motu* (Nunn 1994). Both types are dominated by clastic materials scooped off the front of a coral reef, particularly off reef-talus slopes by large-amplitude waves, and dumped on reef surfaces. Such deposits have been observed to migrate across reef surfaces away from the oceans until they reach a point where they accumulate.

True cays are more transient, sometimes in existence for less than one year, compared to *motu*, some of which are 3,000–4,000 years old. In general, true cays are confined to narrow reef flats, commonly in either high-energy wave environments and/or in places affected annually by storm surges. *Motu* tend to be confined to broader reef surfaces, typically those outside the hurricane (tropical-cyclone) belt, where Holocene sea level exceeded its present level around 4,000 cal. yr BP.

True cays

Being ephemeral and transient reef islands, true cays are generally distinguished by being bare of vegetation and regularly overtopped by waves. They are also characterized by the absence of cemented deposits which renders them vulnerable to obliteration by large waves.

Although we know that true cays usually form as a result of storm surge (or tsunami) deposition of reef-front sediments, we do not clearly understand why this happens only in some instances while in others cays can be removed. It is likely that cay formation occurs when waves are coming from a direction where, in running up the submarine island slope, they can (and do) pick up a lot of material which is carried on to the reef surface. Cay removal may occur at higher velocities but may also be when the wave picks up little material during run-up so that the main outcome of its passing across a reef surface is erosion rather than deposition. But there are other factors involved, particularly to do with reef morphology, sediment character and wave aspect which may lend a cyclical dimension to cay formation and removal. Studies of cays on Ontong Java Atoll in the western Pacific were an important step towards understanding this process (Bayliss-Smith 1988).

One characteristic of cays (and to a lesser extent of *motu*) is their mobility across reef flats. Historical data show that cays change shape regularly and even migrate across reef flats with erosion along the windward side commonly being compensated by progradation along the leeward side. A good example is that of Sand Island off St Croix in the Caribbean (Gerhard 1981). Such movements are the bane of cay-based tourist resorts.

Many cays endure for more than a few decades because they grow large enough to become vegetated and are in appropriate locations to develop beachrock. Such cays are better referred to as *motu*.

Motu

The main way in which *motu* can be distinguished from true cays is by the inclusion of shingle ridges within their fabric (Steers and Stoddart 1977). Such shingle ridges tend to be the residuals of rubble banks thrown up on the outer (ocean) sides of reefs by storm surges. As these ridges migrate landwards or lagoonwards, so the finer material is removed by wave wash so only the coarser fractions remain. Since they are so difficult to shift, particularly when located on the least vulnerable parts of a reef, these ridges often form the core of an atoll *motu*. The migration of the rubble bank thrown up on the Funafuti (Tuvalu) reef during Tropical Cyclone Bebe in 1972 was monitored by Baines and McLean (1976). Later work on the other atolls of Tuvalu demonstrated that such coarse shingle banks were integral parts of *motu*, particularly along windward reefs (McLean and Hosking 1991).

Motu also persist longer than true cays because they develop various forms of physical protection against wave erosion. These include emerged reef, BEACH ROCK, conglomerate platforms (*pakakota*) and phosphate rock, all of which greatly increase the resistance of *motu* against wave and/or precipitation attack, particularly during storms.

Those CORAL REEFs which were able to 'keep up' with Holocene sea-level rise grew to levels of 1–1.5 m above present levels in most parts of the tropics (except apparently the Caribbean) around 4,000 cal. yr BP (Nunn 2001). When the sea level fell by this amount in the later Holocene, the surface of these reefs was exposed and died. Subaerial erosion reduced them and wave erosion trimmed them, but many remained to act as foci for the accumulation of reef detritus. These fossil-reef cores underlie many *motu* today in the central Pacific, for example, and explain their persistence and suitability for human habitation.

Beach rock forms in a variety of ways beneath the surface of sandy beaches within the regularly inundated zone. For beach rock to form also usually requires a critical mass of sediment (equated with minimum *motu* size) so that ground water can flow through the beach sand.

Conglomerate platforms or breccia ramparts are cemented features of many *motu* thought to have formed at present sea level. Although there is clearly some unexplored genetic diversity amongst these features, most are believed to have originated as rubble banks which were subsequently cemented and planed down (McLean and Hosking 1991).

Phosphate rock also forms through the lithification of unconsolidated sediments, but on those *motu* where large numbers of seabirds roost (Stoddart and Scoffin 1983).

The future of cays

Many cays (including *motu*) have experienced apparently unprecedented morphological changes during the twentieth century, many of which can be attributed to the sea-level rise of ~15 cm. It is projected that twenty-first century sea-level rise may be 3–4 times as much, which has resulted in many gloomy prognoses for the future of cays (Roy and Connell 1989).

Should projections of sea-level rise prove correct, then it is likely that the numbers of cays worldwide will decrease hugely. First they may decrease because sea-level rise will, through the Bruun Effect, cause the erosion of sandy shorelines. It may become more common to see lines of beach rock exposed across reef flats marking places where cays once existed. Also, because of the rise of mean sea level and the likely inability of most oceanic reefs to respond immediately (despite some optimistic forecasts), it is probable that sediment of every grade presently lying on reef surface will become more mobile.

For many people occupying cays, particularly in independent countries like the Maldives (Indian Ocean) and Kiribati, Marshall Islands, Tokelau and Tuvalu (Pacific Ocean), it is unlikely that they will be able to continue living in such environments and will become 'environmental refugees'. Questions about national sovereignty and whether or not the Exclusive Economic Zones (EEZs) of these countries will be redrawn as a consequence are exercising the minds of many decision-makers.

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SEE ALSO: coral reef

PATRICK D. NUNN

CENOTE

A cenote is a distinctive type of DOLINE or sinkhole, formed by the dissolution of limestone or other soluble rocks in subdued KARST plains (Plate 21). The type example occurs in the northern Yucatan Peninsula of Mexico, where the Mayan word dzonot, from which cenote is derived, means 'water cave'. Cenotes also occur in south-eastern Australia, in Africa, in Papua New Guinea, in Florida and in north-western Canada (Marker 1976).

The typical Yucatan cenote is a near-circular, water-filled shaft with vertical or overhanging walls extending up to 100 m downward from the ground surface (Pearse *et al*. 1936; Corbel 1959; Gerstenhauer 1968; Doering and Butler 1974). Some Yucatan cenotes resemble cylindrical shafts, but others are flooded bell-shaped chambers with bulbous bases, relatively small surface openings and thin roofs (Reddell 1977). Some have horizontal cave passages leading off from the walls, although these are often blocked by fallen rock (breakdown). The upper portions of cenote walls generally are pitted by dissolution, but the lower walls are blocky and overhanging, suggesting collapse (Whitaker 1998).

The development of cenotes is incompletely understood. Earlier hypotheses suggested that they developed through the local focusing of downward surface dissolution, but it now appears more likely that they have evolved through localized upward dissolution along fractures intersecting groundwater-filled caves or by stoping of cave ceilings, ultimately leading to surface collapse. Global sea-level oscillations have probably played a significant role too, since most cenotes are developed in Tertiary or younger reef limestones in low-lying coastal areas (Marker 1976). Sea-level lowering would have encouraged collapse by reducing the buoyant support of cenote rock walls and ceilings. Cenote-like flooded shafts, known as BLUE HOLES, occur in

Plate 21 A steep-sided cenote formed in dolomitic limestones at Otjikoto near Tsumeb, in northern Namibia

offshore reefs in the Caribbean and Australia (Mylroie *et al.* 1995). These may be drowned cenotes, although some of them have other origins (Ford and Williams 1989).

The distribution of cenotes may be unrelated to other karst landforms, but they are often distributed in a linear pattern that may reflect underground fracture patterns or the paths of major cave passages. It has been suggested that the arcuate pattern of the Yucatan cenotes represents fracturing around the perimeter of the Chicxulub impact crater, which formed at the end of the Cretaceous period, some 65 million years ago (Hildebrand *et al.* 1995).

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MICHAEL J. DAY

CHANNEL, ALLUVIAL

Unconsolidated sediment deposited by rivers in subaerial settings is called ALLUVIUM and river channels formed of alluvium usually have a mobile boundary and are self-adjusting in response to changing conditions. An alluvial channel is commonly parabolic or trapezoid in cross section with adjacent roughly horizontal FLOODPLAINs that are inundated when the channel exceeds *bankfull capacity* (see BANKFULL DISCHARGE). Due largely to tributary contributions, channels generally increase in size and discharge downstream.

Alluvial channel morphology is the product of complex fluid mechanical processes. It was not until the late nineteenth and early twentieth centuries that river channels received widespread and detailed investigation when research was undertaken into partially self-adjusting canals built by the colonial British on the Indian subcontinent. The most important subsequent advances in understanding natural river-channel form and process originated from research in the USA by L.B. Leopold, M.G. Wolman, J.M. Miller, S.A. Schumm and their associates in the 1950s and 1960s.

Channel gradient and knickpoints

Channel gradients are the result of two broad controls. An *independent* gradient is imposed on the stream by antecedent VALLEY forms, products of geological and hydrological history. However, an adjustable and therefore *dependent* component develops from the interaction of channel discharge, width, depth, velocity, sediment size, sediment load, boundary roughness and path sinuosity. Mackin (1948) stated that a *graded* (or equilibrium) stream is one in which, over a period of years, the slope is delicately adjusted to provide, with available discharge and prevailing channel characteristics, just the velocity required for the transportation of the material supplied from upstream. Due to bedrock constraints (see BEDROCK CHANNEL), confined upland channels are generally not at equilibrium gradient, whereas in the middle or lower reaches the valley is wider and a channel can more readily adjust gradient by altering sinuosity and hence path length. Following this original emphasis on gradient, later work has shown that slope provides adjustments in concert with a variety of other morphological and hydraulic parameters (Leopold *et al.* 1964).

For three reasons the long profiles of natural rivers show a strong tendency for upwards concavity. First, downstream discharge increases as a cubic function whereas the resisting channel boundary increases only as a squared function, so if gradient did not decline the growing imbalance between impelling and resisting forces downstream would cause flow to accelerate rapidly. Second, because grain size commonly declines downstream, the equilibrium gradient required for sediment transport must also decline. Third, antecedent relief and potential energy conditions along a river from headwaters to mouth cause random-walk models to develop concavity as the *most probable* profile-shorter streams tend to have less concave profiles and streams with greater relief exhibit greater concavity.

Marked downstream increases in channel gradient are termed KNICKPOINTs and may reflect changes in bedrock erosional resistance, changes in sediment load from tributaries, tectonic activity, meander cutoffs, removal of LARGE WOODY DEBRIS (LWD), or base-level changes in the past. In confined valleys, knickpoints as concentrated zones of erosion can migrate considerable distances upstream. Unconfined channels, however, can adjust more readily by increasing sinuosity and thereby locally reducing knickpoint gradients.

Channel equilibrium and threshold conditions

Because alluvial channels are open systems with mobile and deformable boundaries, they have the ability to self-regulate to the imposed flow and sediment load. This reflects DYNAMIC EQUILIBRIUM, a condition first described for rivers in the nineteenth century by G.K. Gilbert. If one variable is altered, the others adjust in a way that minimizes the effect of the change and the system will return to something like its original condition (*homeostasis*).

While rivers in dynamic equilibrium generally resist change, Schumm (1973) has shown that an *extrinsic* THRESHOLD can be reached when a progressive change in an external variable triggers a sudden change in the system's response. At an imposed critical change of slope or sediment load, a meandering channel can change abruptly to a braided channel (Schumm and Kahn 1972). Similarly, a gradual and progressive increase in the flow velocity may suddenly achieve the threshold for sediment entrainment, after which the whole channel bed becomes mobile. Changes can be initiated intrinsically when, with no external change, one of the variables reaches a critical condition (an *intrinsic threshold*). A meander cutoff is an example where gradual, ongoing adjustments to equilibrium conditions prevail in a channel until an intrinsic threshold is reached.

Biota, soils and channel form

Prior to the middle Palaeozoic, subaerial erosion was dominantly physical and produced abundant coarse material forming mostly braided river channels. In the late Silurian and Devonian the evolution of terrestrial plant communities and associated soils greatly enhanced chemical weathering and the production of clays. The development of cohesive banks of muddy sediment and stabilizing root systems must have changed river channels dramatically. There is a growing appreciation of the importance and complexity of river–vegetation interactions. Particular attention has been given to the influence of within-channel vegetation and large woody debris (LWD) on flow resistance, and of bankline vegetation on bank strength and channel morphology.

The evolution of animals, including dinosaurs in the Mesozoic, has undoubtedly played a part in channel formation. Large mammals (e.g. American buffalo, African hippopotamus and domestic cattle) as well as smaller mammals (such as beavers) have been documented trampling sediments, creating paths down banks and damming channels, leading to channel avulsion and initiation.

Hydraulic geometry, regime theory and dominant discharge

Acceleration due to gravity acts to move water and sediment downslope while flow resistance opposes such motion. The interaction of these two forces ultimately determines the ability of flow to erode and transport sediment and to shape the boundary of an alluvial channel.

Flow velocity is usually fastest at or just below the surface near the centre of a straight channel and declines towards the bed and banks, the flow field deforming through river bends (Figure 19). A narrow deep channel usually exhibits a relatively gentle velocity gradient towards a fine-grained erodible bed, and directs relatively steep gradients to banks that are often cohesive, well vegetated and therefore erosion resistant. Wide shallow channels tend to exhibit erodible banks and coarse and/or abundant bedload that requires high shear stress for transport, braided rivers being a classic type.

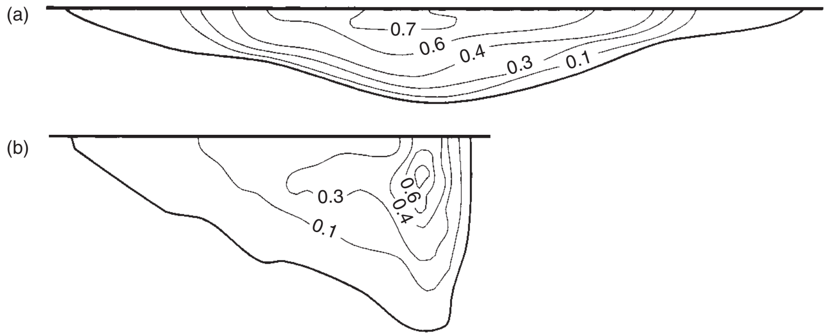

Figure 19 Velocity fields in cross sections of: (a) a wide shallow channel (note the steep velocity gradient to the bed); and (b) a narrow deep channel bend viewed downstream and curving to the left (note the steep velocity gradient against the outer cutbank)

Channel geometry is the cross-sectional form of a stream channel (width, depth, cross-sectional area) fashioned over a period of time in response to formative discharges and sediment characteristics. Because the above three geometric parameters and the additional four flow parameters (velocity, water-surface slope, flow resistance and sediment concentration) vary with discharge, the term HYDRAULIC GEOMETRY or *regime theory* is used to describe the relationships of all seven parameters to discharge as the independent variable (Figure 20). Consequently, stable alluvial rivers exhibiting consistent and predictable hydraulic geometries are said to be *in equilibrium* or *in regime*.

Discharge changes can be measured increasing at-a-station as the channel fills during a flood, or at bankfull in the downstream direction. There are significantly different relationships for *at-a-station* and *downstream* hydraulic geometry (Figure 20). Holding discharge constant, at-a-station hydraulic geometry is controlled mostly by variations in bank strength and available sediment load. Channels with low sediment loads and cohesive or well-vegetated banks tend to be relatively narrow and deep whereas those with abundant loads and weak banks tend to be wide and shallow. However, because bank strength has only a moderate range but river discharges vary by many orders of magnitude, hydraulic geometry is remarkably consistent across the full range of river discharges (Figure 20). Furthermore, because channel depth is greatly restricted by the limited strength of alluvial banks, rivers increase in width relative to their depth as their size and discharge increases – a prominent downstream tendency (Church 1992).

Hydraulic geometry shows that river channel dimensions are closely adjusted to water discharge. However, discharge varies from perhaps no flow in droughts through to catastrophic flood events, so which discharge(s) define a channel's characteristics? Leopold *et al.* (1964) showed that, in the USA, bankfull flows occur with the surprising regularity of about once every 1–2 years across a diverse range of rivers, something that would be an extraordinary coincidence if bankfull flows did not in themselves play a large part in determining channel dimensions. It has also been shown that with increasing at-a-station discharge, flow velocity tends to increase until near bankfull flow conditions and then stabilizes at higher discharges because of a marked increase in roughness near the bank crests and over the floodplains. In other words, most flows beyond bankfull are not notably more effective in altering the channel and transporting sediment than is bankfull flow. Furthermore, while in some cases exceptional floods may undertake significant work in the form of sediment transport and channel reconstruction, they are sufficiently rare that on an average annual basis, they usually achieve far less than do smaller but more frequent events of about bankfull stage (Figure 21) (Wolman and Miller 1960).

In some environments, particularly in confined alluvial settings, high-velocity events can cause considerable channel enlargement, followed by

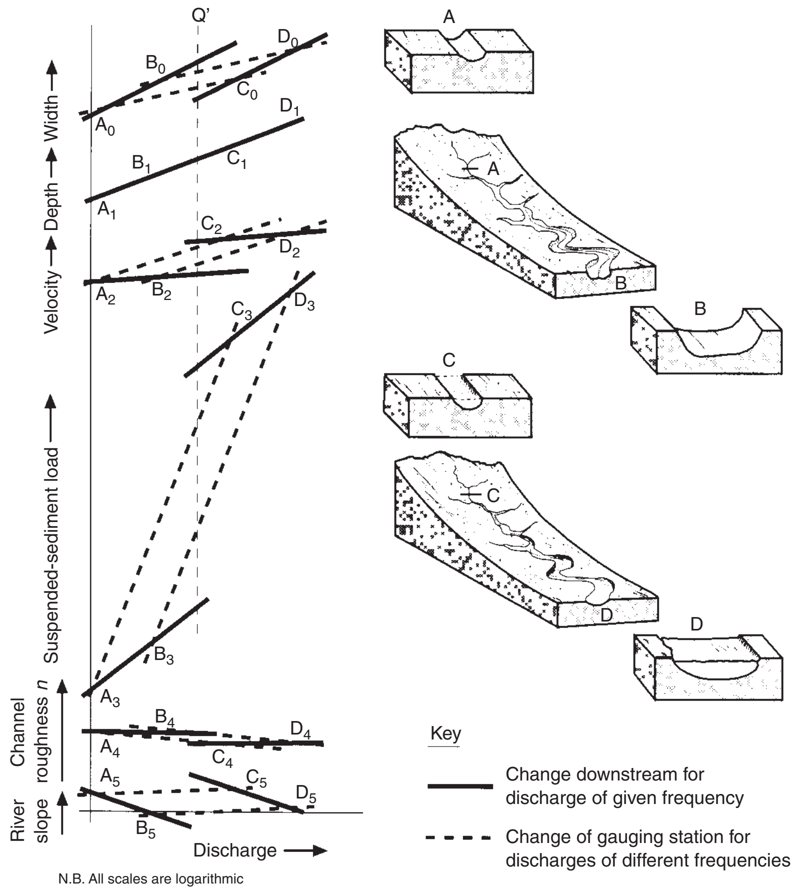

Figure 20 Hydraulic geometry relationships of river channels, comparing variations in width, depth, velocity, suspended load, toughness and slope to variations in discharge, both at-a-station and downstream (after Leopold *et al.* 1964)

a long period of 'recovery' from smaller flows. Thus, channel dimensions at a given time in such an environment may reflect considerable 'memory' of the last extreme event.

While empirical research into stochastic relationships has shown that, as flows vary, rivers construct highly predictable channel forms and sedimentary structures, a truly rational or deterministic explanation for such consistency has not been obtained. A lack of mathematical closure results from there being four flow variables (width, depth, velocity and slope) but only three determining equations (continuity, resistance and sediment transport). As a consequence, solutions have been sought by adopting *extremal hypotheses*, such as maximum sediment transport rate or minimum stream power. In a recent reassessment of some of these approaches, Huang and Nanson (2000) have demonstrated mathematically that straight reaches of alluvial rivers appear to operate at MAXIMUM FLOW EFFICIENCY (MFE) and illustrate the basic physical LEAST ACTION PRINCIPLE. However, although research into this principle is ongoing, such theoretical proposals remain contentious and are not uniformly accepted.

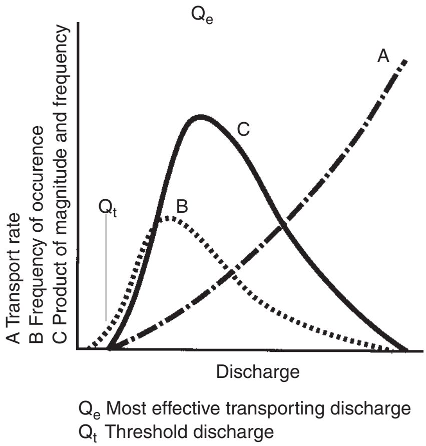

Figure 21 Dominant discharge. Curve A is the transport rate of suspended load rising with discharge. Curve B is the frequency of the full range of possible discharges. Curve C is the product of curves A and B and shows that the most effective discharges for transporting a river's load are moderate floods, generally occurring about once every 1 to 5 years (after Wolman and Miller 1960)

Channel patterns

River channels respond to imposed discharges and sediment loads by adjusting pattern or planform in conjunction with their hydraulic geometry. Because channel patterns are so easily recognized on air photos and maps they have become a primary basis for river classification from which it is possible to generalize less obvious channel characteristics such as lateral stability, sediment load, sediment size, bed/suspended load ratio and width/depth ratio.

Leopold *et al.* (1964) proposed the first widely adopted geomorphological classification with their concept of a continuum of river channel patterns from *straight* to MEANDERING to BRAIDED RIVERS, although a significant problem is that these are not mutually exclusive. For example, meanders sometimes also braid. Both meandering and braiding patterns appear to reflect a need to consume excess energy where the valley slope is greater than that required for an equilibrium channel slope (Bettess and White 1983; Schumm and Kahn 1972).

The term *anabranching* describes rivers that flow in multiple channels separated by stable, vegetated, alluvial islands that divide flows up to bankfull, regardless of their energy or sediment size, with *anastomosing* rivers simply a low-energy fine-grained type (see ANABRANCHING AND ANASTOMOSING RIVERS) (Nanson and Knighton 1996). Importantly, neither of these terms is now used as a synonym for braided rivers in which the flow is divided by unstable braid bars overtopped below bankfull.

Straight rivers consist of a single channel with a sinuosity of <1.1 (sinuosity is the ratio of channel length to valley length), a condition rarely persisting in an alluvial reach for distances of more than 7–10 channel widths. Consequently, straight reaches are classed as those without significant bends for more than this distance. Flume experiments suggest that straight channels form at very low gradients (Schumm and Kahn 1972). Where naturally sinuous channels in readily erodible material have been artificially straightened, alternate bars usually form rapidly and subsequent bank erosion leads to the development of a meandering pattern.

Braided rivers are relatively high-energy systems with large width–depth ratios and at low stage have multiple channels that divide and rejoin around alluvial bars. They tend to occur in settings with steep gradients, weakly cohesive banks, abundant coarse sediment (usually gravel and sand), and variable discharge (Leopold *et al.* 1964; Knighton 1998). Leopold *et al.* (1964) argued that braiding is an equilibrium adjustment to erodible banks and excessive load whereas Bettess and White (1983) see it as a pattern consuming energy in excessively steep valleys. Both explanations probably apply under different circumstances.

Meandering rivers consist of a single channel of moderate to low gradient with a sinuosity >1.3 and moderate width–depth ratios. Point bars commonly develop on the convex bank of a bend whereas the concave bank is typically erosional and adjacent to a pool. The locus of the lowest point in the channel (the thalweg) regularly oscillates laterally, switching channel sides at the riffle (crossover), a shallow zone in the long profile between each bend (pool). While there is no general agreement as to exactly how or why streams meander, they are self-similar over a wide range of scales. Width and wavelength in particular can be related to channel discharge (see Knighton 1998, Table 5.9). Using 'probability theory', Langbein and Leopold (1966) proposed that meanders reduce stream gradients to an equilibrium slope for the transport of an imposed sediment load (see also Bettess and White 1983), producing a longer path length with minimum variance and minimum total work. Meandering rivers tend to have cohesive and/or well-vegetated banks, mixed loads of sand (sometimes gravel) and mud, and commonly perennial flow. Channel lateral migration rates are most rapid where bends have a radius of curvature to channel width ratios of about 2 to 3 (Hickin and Nanson 1975).

Anabranching rivers are a system of multiple channels characterized by vegetated, stable alluvial islands which are either excised by avulsion from a previously continuous floodplain, or formed by the accretion of sediment in a previously wide channel. The islands divide flow up to bankfull (Nanson and Knighton 1996). Anabranching rivers include a wide range of subarctic, alpine, temperate, wet tropical and semiarid settings and individual channels can be straight, meandering or braided. *Anastomosing* channels are low gradient, laterally stable, straight (most common) to highly sinuous variants, with low width–depth ratio and well-vegetated or highly cohesive banks. Anabranching rivers can confine flow and maintain an equilibrium bedload flux over low gradients, however, they can also distribute sediment over wide floodplains in disequilibrium, vertically accreting locations.

Church (1992) noted the problem of including rivers from mountains to basins within one classification scheme. He divided the full range of alluvial and non-alluvial channels into small, intermediate and large categories based not on channel dimensions but on the relationship between grain diameter (D) and depth (d). This approach offers opportunities for better classifying aquatic habitats but is less visually appealing.

Sediment transport and channel sedimentation

River channels transport their sediment load in essentially four ways; *bedload (traction load), saltating load, suspended load and dissolved load.* BEDLOAD is the coarsest fraction and moves short distances during relatively infrequent, high magnitude flows. It is commonly the smallest proportion of transported sediment (often <5 per cent of the total load), yet is of great geomorphic importance. It is largely bedload that controls channel configuration because its transport is a function of shear stress acting on the channel bed, and this is controlled by channel gradient (adjustable with sinuosity) and channel geometry. The capacity of alluvial rivers with unconstrained mobile boundaries to transport bedload is usually hydraulically defined, but few flows reach their capacity for transporting suspended load that is determined largely by the rate of supply. In other words, the character of a river is first determined by its imposed bedload, with suspended load and vegetation influencing bank cohesion and form. Because sediment character has a profound influence on river-channel morphology, Schumm (1960) developed a highly influential classification of rivers based on bedload, mixed load and suspended load systems, with width–depth ratios of >40, 10–40 and <10, respectively.

Alluvium results from fluvial sedimentation. This takes place both inbank and overbank as velocity wanes locally. The coarsest fractions are deposited first and as a result, sediment sizes are sorted vertically and laterally within the channel and floodplain. In laterally migrating meandering channels, upward fining successions within point bar and floodplain deposits result from flow velocities that decline from near the deepest part of channel (the thalweg) and adjacent point bar (depositing gravel or coarse sand), to the upper point bar and floodplain surface (depositing fine sand and mud). In braided rivers, coarse braid bars characterize the lowermost deposits while braid-channel and braid-bar migration or abandonment, overbank fines and channel fills characterize the uppermost deposits. Adjacent to laterally stable channels, or on floodplains away from the zone of active channels, floodplain strata broadly fine upward as each successive stratum makes the surface higher and less accessible to channel flows.

Secondary currents play a major role in producing the broad spatial variations of sediments in channel bends and bars, as well as numerous smaller flow structures. In gravel streams, prolonged flows near critical entrainment conditions can winnow fines and armour the surface, thereby lifting the threshold of bed motion during the next flood.

Conclusion

Alluvial channels represent continuum of forms that are classifiable on the basis of their cross-sectional shape, planform and associated processes. Whereas early research focused on stochastic relationships between channel form and process, there is a growing appreciation of rational explanations based on mechanics and accepted physical theory. Research into the operation and maintenance of alluvial channels remains a major area of pure and applied research within fluvial geomorphology.

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SEE ALSO: armouring; bank erosion; channelization; confluence, channel and river junction; gravel-bed river; levee; long profile, river; models overflow channel; riparian geomorphology; river continuum; roughness; sediment load and yield; suspended load

GERALD C. NANSON AND MARTIN GIBLING

CHANNELIZATION

Channelization is the term used to describe the modification of river channels (usually alluvial channels, see CHANNEL, ALLUVIAL) by engineering. The aim is to provide flood control, improve land drainage, reduce erosion of channel banks and river beds, improve and maintain river navigation and to relocate channels in situations such as highway construction (see Brookes 1988 for a detailed text on channelized rivers). Some river channels have also been altered to float logs out from forests. River engineering can also create impounded rivers through the construction of DAMs (Petts 1984). Whilst the term channelization is extensively used, there are some equivalent terms used for the same group of engineering methods. These include 'kanalisation' in Germany, 'chenalisation' in France and 'canalization' in the UK.

River channelization has a long history and a large geographical coverage. Its origins can be traced to Mesopotamia and Egypt where there is evidence of river channelization for flood control and water supply as early as the sixth millennium BC. Indeed, most early civilizations constructed flood embankments. By 600 BC, reaches of the Huanghe (Yellow River) in China were embanked, and in Britain, the Romans constructed embankments to provide flood protection in the Fens and Somerset Levels.

Not surprisingly, the highest density of channelization occurs in developed countries associated with industrialization, urbanization and the intensification of agriculture. In the USA, 65 per cent of all channel alteration work is concentrated in Illinois, Indiana, North Dakota, Ohio and Kansas, with 51 per cent of all levee (embankment) work in California, Illinois and Florida (Brookes 1988: 10). In England and Wales, Brookes *et al.* (1983) estimated that for the period 1930–1980, 8,500 km of main rivers underwent major structural river engineering and a further 35,500 km were regularly maintained by dredging and weed cutting. And in Denmark, it is estimated (Brookes 1987) that 97.8 per cent of all streams were straightened by 1987. This is equivalent to a density of modified watercourses of 0.9 $\mathrm{km\,km^{-2}}$ and compares with a density of channelized rivers in England and Wales of 0.06 $\mathrm{km\,km^{-2}}$ (Brookes *et al.* 1983) and 0.003 $\mathrm{km\,km^{-2}}$ for the USA (Leopold 1977). Thus, Denmark has a density of channelized river fifteen times greater than

England and Wales and 300 times greater than the USA, reflecting its intensity of land use.

Engineering techniques for flood control aim to prevent flood discharges overtopping the channel banks and spilling out onto the surrounding floodplain. Channels are designed and engineered to carry a design flood, which has a particular magnitude and frequency. If the 100-year event is selected as the design flood, the river channel will be engineered to contain and transmit a peak flow that will occur on average once every 100 years.

A range of 'structural' or 'hard' river engineering techniques are employed in channelization (Wharton 2000: 24–34) and many projects are comprehensive or composite in nature in that they employ more than one of the following engineering techniques.

Resectioning increases the cross-sectional area of the channel through widening and/or deepening. This allows flood flows that would have previously spread onto the floodplain to be contained and to flow through the channel at a lower and safer level. By combining with a process known as regrading (smoothing out the river bed by removing features such as depositional bars and pool-riffle sequences) the flow velocities are increased and flood levels are further reduced in the engineered reach.

Embankments, also known as levees, floodbanks and stopbanks, are structures built alongside a river to increase the bank height and prevent flooding onto the floodplain. They are normally constructed from material excavated from the channel or from a borrow pit in the floodplain, although imported materials are sometimes used. Detailed design specifications exist for embankments but a major consideration is the height, determined by the design flood.

Lining of channels in artificial materials is undertaken for both flood control and channel stability. Lined channels are common in urban areas and are normally rectangular in cross section with a straightened planform.

Realignment or straightening aims to improve the ease with which water flows through a river reach. The techniques range from removing deposited sediment by dredging (regrading), for example 'rock raking' carried out on gravel-bed rivers in New Zealand, to the removal of meander bends through cutoff programmes, for example the Middle Yangtze and Huanghe rivers in China and the Lower Mississippi river in the USA. River straightening also improves river navigation.

Diversion channels are relief channels constructed to divert flood flows away from an area requiring protection. The Jubilee River (completed in 2002) is a diversion or bypass channel providing flood relief for part of the River Thames catchment, UK. The newly engineered Jubilee River has a maximum capacity of $215\,m^3\,s^{-1}$ and the main channel of the River Thames and existing right bank flood channels can carry up to $300\,m^3\,s^{-1}$. It is predicted that the overall system capacity of $515\,m^3\,s^{-1}$ will protect up to the 1 in 65-year return period flood. For environmental reasons the Jubilee River maintains a flow of $10\,m^3\,s^{-1}$ at all times.

Culverts are structures that encase watercourses to provide flood protection. They may be masonry arches or large-diameter concrete or metal pipes. In many towns and cities, culverted streams flow beneath the streets, for example the rivers Fleet, Westbourne and Tyburn in London (UK).

Bank protection methods and river training works are engineering techniques for controlling river channel adjustments that could threaten settlements and agricultural land and have an impact on river navigation. Deposits from eroded riverbanks can also impede the river flow and increase the risk of flooding. Riverbanks have traditionally been protected by riprap (quarried stone), gabions (rock-filled wire baskets) and revetments (coverings of resistant materials such as concrete, steel or plastic sheeting). Although riprap is usually the preferred option, gabions do have an advantage in that the wire mesh allows the rocks to change position (caused by unstable ground or scouring of the riverbank) without failure. River training works are structures built to extend from the channel banks into the river and provide bank protection by deflecting erosive river flows away from vulnerable areas along the channel banks. The most common structures are groynes (also known as deflectors or dikes). Flows can be deflected onto channel deposits that pose problems for navigation or flood control to promote their removal through the natural process of scouring. Groynes have been used in this way on the Mississippi River to maintain a navigation channel. River training works can also be used to promote sediment trapping and deposition in areas that have suffered erosion. For example, a series of permeable groynes will allow water to pass through the structures but induce deposition of fine suspended sediment between the groynes, whereas impermeable groynes will deflect river flows and promote the trapping of larger bed material.

Dredging, weed cutting, clearing and snagging (collectively known as channel maintenance activities) are routinely undertaken on many rivers to improve the efficiency of water flow through the channel and reduce the flood risk. The removal of 'obstructions' to flow, reduces channel roughness, increases river flow velocity and lowers the flood height for a given discharge. Dredging may simply involve breaking up and loosening material for the river to transport downstream. In contrast, sediment may be removed by mechanical diggers, pumped onto the floodplain or be discharged into river barges before being dumped at selected locations. Weed cutting is practised in many streams, especially nutrient-rich chalk streams, to control the annual growth of submerged and emergent aquatic plants. In addition to physically reducing the capacity of the channel, plants also increase flood risk by increasing the resistance to flow and reducing water velocities. This further promotes the accumulation of sediment within and around the plants. Aquatic vegetation is traditionally controlled by mechanical cutting, but herbicides and grazing fish (such as carp) have also been used. Clearing and snagging refers to the removal of fallen trees and debris dams from the river and the harvesting of timber from the channel banks and floodplains, respectively.

A number of concerns surround river channelization. First, channelization, is unable to provide complete protection against flooding and its associated channel form adjustments. It is simply not possible or economically viable to control the very rare, high-magnitude flood events. To achieve this, all rivers would need to be channelized and all flood defences would need to be designed and constructed to convey a correctly estimated maximum possible flood. Second, there is evidence from developed countries with a long history of channelization that the financial costs of floods are continuing to rise despite ever-increasing expenditure on structural flood defences. This has been attributed, at least in part, to the false sense of security created by flood defences that encourages further floodplain development. And third, river channelization has resulted in changes to the river, many of which were not anticipated at the design stage. These changes can have a damaging impact on the river environment and also necessitate costly maintenance activities to keep the structures operating at their design specifications. Brookes (1988: 81–185) provides a comprehensive review of the main impacts of river channelization. Included in this third set of concerns are fears that river engineering may have worsened flooding on some rivers. Whilst channelization can reduce flood risk in the engineered reach, the reverse may be true downstream.

Brookes (1988) describes the primary impact of channelization as the physical alteration to the river (i.e. its width, depth, slope and planform) by the engineering procedure. These changes then result in secondary effects that encompass changes to the river channel morphology, hydrology, water quality and ecology. Importantly, these impacts are transmitted beyond the channelized river section to downstream and upstream reaches and even along tributary streams. Post-engineering adjustments demonstrate the need for long-term and often costly maintenance operations and also have implications for structures built adjacent to, or across, river channels. For example, bridges may have to be reinforced or even replaced if bank erosion causes the river to migrate and enlarge.

The reporting of channelization impacts and the appraisal of channelization schemes will lead to improved understanding of the various changes that river engineering may cause. Greater recognition of the undesirable consequences of channelization has led to calls for a 'reverence for rivers' (Leopold 1977) with attempts to design with nature (after McHarg 1969) and develop 'geomorphic engineering' (Coates 1976). This has translated into a variety of revised construction and maintenance procedures (see Brookes 1988: 189–209) and the development of more environmentally sensitive flood alleviation schemes, such as the flexible two-stage channel constructed on the River Roding, UK (Raven 1986). In this design, the additional capacity is created by excavating outside the original channel thus leaving it to transport the normal range of flows and remain as natural as possible. Growing concern over the impacts of channelization has also prompted efforts to enhance, rehabilitate and restore river systems (see RIVER RESTORATION).

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SEE ALSO: anthropogeomorphology; bankfull discharge

GERALDENE WHARTON

CHAOS THEORY

Chaos theory has been claimed by some enthusiasts as being one of the great ideas of twentieth century science which, as with relativity and quantum mechanics, has the power to transform our view of the world. As the popular book on chaos by James Gleick (1987) illustrates, chaos theory developed in a series of often unrelated spheres of science as developments in computing power permitted the increasingly sophisticated study of non-linear systems. Non-linear systems are those in which a change in one variable produces a non-linear response in another, and thus have to be represented by non-linear equations. Chaos is a property sometimes exhibited by such non-linear systems where even under simple conditions the system can tend to complex, pseudo-random behaviour. A classic paper by Edward Lorenz in 1963 illustrates the potential for chaos in relatively simple systems. Lorenz developed a simple climatic model of the atmosphere heated from below to produce convection, involving three non-linear equations. The three equations describe the change in x, y and z over time respectively, where x describes the intensity of convective motion, y the horizontal temperature variation and z the vertical temperature variation. Despite its simplicity the modelled system exhibited chaotic behaviour, indicating the unpredictable behaviour of this sort of climatic system.

Chaotic behaviour can be identified in systems through using phase diagrams, which plot the state of the system over time in terms of the system variables. In the case of the Lorenz model above, for example, the phase diagram would plot each point in time of the evolution of the system on x, y and z co-ordinates. A stable system would have a phase diagram which converged on a point, an oscillating or periodic system would have one which resembled a ring. Such shapes on a phase diagram are called attractors. Phase diagrams for chaotic systems are characterized by what are called 'strange attractors' – bifurcating, complex patterns illustrating the many different possible states of the system as it evolves over time. Lorenz's model, for example, has a strange attractor which looks like an owl mask. Strange attractors are fractals (see FRACTAL).

According to Malanson *et al.* (1990) chaos theory has three central tenets. First, that many simple deterministic systems are rarely predictable. Second, that some systems show great sensitivity to initial conditions. Tweaking an input to one equation of a system very slightly at the beginning can thus produce highly divergent outcomes. Third, that the conjunction of the first two tenets produces a seeming randomness which may be quite ordered (as illustrated by the presence of strange attractors in their phase diagrams).

Geomorphologists have been keen to investigate the utility of chaos theory ideas and methods for the study of geomorphic systems, many of which can be shown to be non-linear in nature. For example, in a series of papers Jonathan Phillips has investigated the presence of chaos in surface runoff, hillslope evolution, coastal wetlands and soil systems as reviewed in his book on Earth surface systems (Phillips 1999). Mass movement systems often behave chaotically (Qin *et al.* 2002). Increasingly, geomorphologists suspect that chaotic and self-organized behaviour (see SELF-ORGANIZED CRITICALITY) may be common within Earth surface systems, and that stable states may be relatively uncommon. However, chaotic behaviour may also be scale-dependent, and at other scales ordered behaviour may emerge. For example, at the microscale turbulence (a classic manifestation of chaotic behaviour) characterizes many aeolian-sediment interactions within dunefields, whereas at the larger scale ordered dune systems result. As Phillips (1999: 71) puts it 'Order is an emergent property of the unstable, chaotic system'.

Although chaos theory has undoubtedly stimulated much interesting and useful research and discussion in geomorphology, its application to geomorphic systems is not problem-free. Three key issues are discussed by Baas (2002). First, the presence of random noise within many geomorphic systems can often mask chaotic behaviour and make it almost impossible to analyse what is going on. Second, analysing chaos requires good datasets, which are not necessarily forthcoming in many areas of geomorphology, although the advent of good quality DIGITAL ELEVATION MODELS (DEMs) at a range of scales has started to help enormously in this regard. Finally, there are a range of different interpretations of chaos theory in the scientific literature, and many different methods available to analyse chaotic systems – all of which can be rather confusing to geomorphologists wishing to understand and utilize chaos theory.

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HEATHER A. VILES

CHELATION AND CHELUVIATION

Organic compounds, derived through the partial decomposition of organic matter, are important agents in weathering. Some act because they are acid and simply etch into minerals but for others, ions from the mineral actually become incorporated into the chemical structure of the organic compound and it is these compounds which are called chelates. The word is derived from the Latin *chela* and Greek *khele* which means a claw – and it can be readily imagined how the claw of, say, a crab can hold an object in the tips of its pincers and this is analogous to the way in which the chemical compound holds an atom derived from a mineral. The definition of a chelate can now be appreciated: 'a compound containing a ligand (typically organic) bonded to a central metal atom at two or more points'; where a ligand is: 'an ion or molecule attached to a metal atom by co-ordinate bonding'.

In the context of weathering, the metal ions of interest are commonly iron but can be zinc, copper, manganese, calcium or magnesium. Chelation weathering is then the process of the incorporation of these metal atoms into an organic compound derived from the decay of organic matter. The significance of this process is that many minerals are subject to weathering by chelates to a much greater degree than they are in water, even acidified water (Huang and Keller 1972; Huang and Kiang 1972).

Cheluviation is a compound word derived from chelation and eluviation (see ELUVIUM AND ELUVIATION). Since eluviation is the down-washing of material through the soil in mobile soil water, cheluviation involves the down-washing of chelates, with their associated metal cations, from the upper horizons of the soil to the lower horizons. It is in this way that iron can be moved from the upper horizons of a podzolic soil, rendering it a pale colour with an absence of reddish oxidized iron, ferric iron or Iron III. The process involves simultaneous chelation and REDUCTION of the iron to the mobile ferrous (Iron II) form. The iron then may accumulate lower down in the soil as a reddish or, because of the presence of organic matter, blackish layer. Here the reddish oxidized Iron III forms as a result of OXIDATION through a rise in pH which occurs in the lower parts of the soil profile which are less acid than the upper parts which are acidified by organic acids. The redeposition of the iron can be in the form of a hard iron pan, termed a BFe horizon, or a more diffuse reddish horizon. The latter is termed a Bs horizon as it contains sesquioxides which are defined as compounds such as Fe_2O_3 which have a ratio of metal to oxygen of $1{:}1\frac{1}{2}$.

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STEVE TRUDGILL

CHEMICAL DENUDATION

Central to any understanding of landform change through time is an understanding of DENUDATION rates (the volume of rock removed from a given area in a specified period of time). Knowledge of denudation rates is relevant also to geochemical and sediment mass balance studies, with important implications for global carbon budgets and global climate change. Denudation results from the removal of solid particles (mechanical denudation; Meybeck 1987) and dissolved material (chemical denudation). Overall, chemical denudation has received less attention than mechanical denudation and estimates of its local, regional and global significance often are subject to much uncertainty. The processes of CHEMICAL WEATHERING through which atmospheric, hydrologic and biologic agents act upon and alter mineral constituents of rocks by chemical reactions, thereby releasing dissolved material to be removed, are considered elsewhere. Here, an overview of the methods for studying chemical denudation and the variability of rates from different environments is provided. The role of relief, lithology and climate as controlling factors are discussed.

Methods for studying chemical denudation

SOLUTE YIELDS

Most frequently chemical denudation is calculated from the solute loads of rivers draining large catchments. An estimate of chemical denudation can be achieved simply by multiplying the mean solute concentration from samples of river water by mean discharge. More accurate estimates, however, take into account solute concentration relationships with discharge, particularly through floods using solute rating curves (see SOLUTE LOAD AND RATING CURVE) constructed from equations which best fit the relationship between solute concentrations and discharge. For greater accuracy these rating curves can be used for the rising and falling limbs of flood hydrographs. Solute transport rates can then be calculated by relating the rating curve to either continuous stream-flow data or flow-duration curves based on hourly, daily or even monthly data.

Given the complexity of measuring separately each dissolved constituent in stream water, electrical conductivity (specific conductance) of the water, which is more easily measured, is often used to provide an estimate of solute concentration. Although there is a strong correlation between the concentration of ionic species in solution and electrical conductivity, the exact relationship varies depending on concentrations present of particular dissolved constituents. Moreover, SiO_2, which may be a significant component of many tropical lowland rivers, is not recorded by this technique.

The most significant problem in estimating the contribution of solute transport to denudation is the separation of denudational and non-denudational contributions. Chemical weathering is not the only process affecting solute yields (Figure 22). Dissolved constituents introduced into a catchment from atmospheric wet and dry deposition must be accounted for. These atmospheric deposition fluxes can be quantified by direct measurement, though results often are highly variable with complex spatial patterns in regions with different vegetation types (Drever and Clow 1995). Global estimates of non-denudational atmospheric inputs (details in Summerfield 1991) from precipitation (oceanic salts) and atmospheric CO_2 (incorporated during weathering reactions), average approximately 40 per cent of catchment solute yields. The fraction is highest for the ions of Na, Cl and HCO_3 (50 to 70 per cent), although it is important to caution that these values vary greatly.

Further complications, depending on the timescale of study, relate to changes in the exchange pool of cations and anions in the soil and biomass (Figure 22). In the short term, changes in the soil occur as a result of precipitation events, evapotranspiration and the growth cycles of plants. As plants grow, they extract inorganic nutrients from the soil solution and incorporate them into plant tissue. When plants die and decompose, the process is reversed and the elements are returned to the soil. If an ecosystem

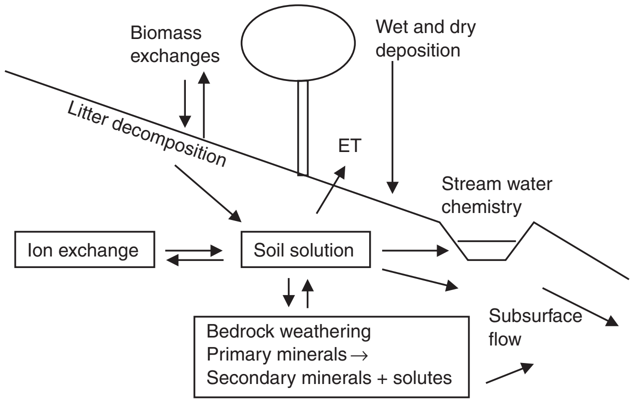

Figure 22 Schematic representation of key factors influencing solute fluxes in a catchment (adapted from Drever and Clow 1995)

is in steady state, new growth is exactly balanced by the death and decay of old vegetation and the biomass is neither a net source or sink. However, in forested catchments the biomass is rarely in steady state. For example, data of Likens *et al.* (1977) in Hubbard Brook indicated that the uptake of Ca by biomass was 45 per cent of the amount released by weathering. For K the value was 86 per cent.

The solute loads of rivers increasingly are being impacted by human inputs, especially where industrial and agricultural activities are concentrated. Elevated acid inputs increase the rate of chemical weathering in watersheds underlain by reactive rock types and cause acidification of water and soil in catchments underlain by non-reactive rocks types. Significant changes in the soil and biomass exchange pools can result, enhancing rates of solute input into stream waters. Such anthropogenic influence further complicates interpretation of solute concentrations in terms of 'natural' chemical denudation rates.

While corrections can be made to solute yields to account for atmospheric, biogenic and anthropogenic processes, the actual volume change (denudation) in the catchment cannot be determined unless the alterations in bulk density that accompany the weathering reactions releasing the solutes are known. While some chemical weathering dissolves bedrock minerals completely with all the products as dissolved species (common with limestone and quartzite) (congruent reactions), many weathering reactions produce both dissolved species and new solids (often clay minerals) with a similar volume but decreased bulk density. Moreover, even the congruent reactions and associated chemical losses may result in a substantial decrease in density as silicate rocks are altered to SAPROLITE (a weathered residuum retaining the structure and layering of the bedrock from which it forms) (density of bedrock 2.5–2.7; saprolite 1.3–1.7; soil 0.8–1.3). Thus there may be no discernible effect on the configuration of the landscape and direct conversion of dissolved loss into surface lowering is unrealistic.

SOIL PROFILE DEPLETION AND MASS BALANCE MODELLING

Soil mineralogy represents the residual product of chemical reactions which integrate the weathering rate over the entire period of soil development. Thus an alternative approach to determine long-term rates of chemical weathering and denudation is to quantify element and mineral losses in a soil profile relative to the initial or parent material. The most common approach is to define the mass ratio (enrichment) of a conservative component whose absolute mass does not change during weathering. As relatively soluble minerals in soils dissolve away, the more immobile elements in soils become increasingly enriched relative to their concentrations in the unweathered parent material. Measurement of enrichment of immobile elements, such as Zr, Ti, of rare Earth elements such as Nb, can reveal the degree of soil weathering and thus can be used to quantify the total dissolution loss from a soil (see examples in White 1995). However, there is considerable disagreement in the literature about the relative mobility of elements in different weathering regimes, and minor elements, such as Zr and Ti, are often concentrated in the small size, heavy fraction which may be subjected to significant fractionation during sediment transport and deposition. Assuming these are not major issues, the average weathering rate can be estimated by dividing the dissolution loss by the soil age. However, because non-eroding soils of known age are rare, this mass balance approach cannot be used in many environments. Riebe *et al.* (2001) show how the soil mass balance approach can be extended to measure long-term weathering rates in eroding landscapes. Physical erosion rates can be inferred from cosmogenic nuclides (see COSMOGENIC DATING), with dissolution losses inferred from rock-to-soil enrichment of insoluble elements.

Regional and global patterns

The distribution of studies of solute loads and chemical denudation is uneven. Most recent work on the rates and significance of chemical weathering in small watersheds has been driven by interest in the effects of acid deposition. Thus numerous studies have been conducted in North America and Europe, yet few data collected at the watershed-scale exist from other parts of the world. Thus estimates at continent-wide or global-scales must be extrapolated. This is usually based on empirical relationships observed between solute transport rates and the factors thought to control these rates, notably rock-type, climate and relief (see further discussion below). Moreover, records of dissolved loads tend to be short and results variable through time. Thus there are also problems extrapolating such data to the longer time periods over which ecosystems, soils, landscapes and climates evolve.

Some of the earliest regional estimates of chemical denudation were attempted by Dole and Stabler (1909) for the United States. Data compiled by Summerfield (1991) are used here to provide a range of estimates of solute load transport and equivalent rates of chemical denudation (see regional summary in Table 7). These estimates are subject to all the errors described above.

The data yield a global average for chemical denudation of 3,700 Mt a^{-1}. Reducing this value by 40 per cent, to account for non-denudational component of solute loads (see discussion above), the estimate is 2,200 Mt a^{-1} for denudational solute load. Globally, this is approximately 15 per cent of natural mechanical denudation. Chemical denudation rates although less variable than mechanical denudation rates, do vary significantly (Table 7). Reported values range from 1 mm ka^{-1} in drainage basins such as the Nile, Niger and Rio Grande to 27 mm ka^{-1} in the Chiang Jiang basin (Summerfield and Hulton 1994).

Some of the measured variability is related to lithology. Maximum yields of 6,000 t km^{-2} a^{-1} occur in rare instances (for example in areas underlain by halite). More usual maxima are 1,000 t km^{-2} a^{-1} in limestone regions. Although few studies have attempted to reconcile laboratory-based experimental studies of weathering rates and catchment scale estimates, the real-world weathering rates of different lithologies do correspond qualitatively to rates measured in the laboratory (Drever and Clow 1995). Many studies have documented a consistent positive correlation between solute load and annual runoff. This results from more water available for chemical reactions in the regolith and solute release, and greater runoff to transport these solutes. The relationship with temperature tends to be very weak (overwhelmed by other variables, especially precipitation and local relief). Relief influences a number of factors which impact the rate of surface runoff, rate of subsurface drainage and therefore rate of leaching of soluble constituents, and rate of erosion of weathered products and thereby rate of exposure of fresh mineral surfaces. In the Amazon basin, a relationship between relief and chemical weathering exists: ~86 per cent of the solutes delivered by the Amazon to the Atlantic come from the Andes mountains (~12 per cent of the area) (Gibbs 1967). The problem, however, is that for the Amazon relief and lithology are highly correlated; outcrops of limestone and evaporates are common in the Andes, whereas most of the remainder of the basin is underlain by silicate rocks. Based on data for externally draining basins exceeding 5×10^5 km^2 in area, Summerfield and Hulton (1994) conclude that chemical denudation rates are more strongly associated with relief than climatic factors. This supports the idea that the efficient removal of bedrock in the weathering zone is the critical determinant of the rate of chemical weathering.

Table 7 Solute denudational loads of major rivers in relation to climate and relief

Climate and relief zone	Denudational solute load ($t\,km^{-2}\,a^{-1}$)	Total denudation ($mm\,ka^{-1}$)	Typical solute load as % of total
Mountainous			
High precipitation	70–350	95–740	10
Low precipitation	10–60	45–370	10
Moderate relief			
Temperate or Tropical climate	25–60	30–110	35
Low relief			
Dry climate	3–10	5–35	10
Temperate climate	12–50	15–30	65
Subarctic climate	5–35	5–15	80
Tropical climate	2–15	1.5–10	50

Sources: Adapted from Summerfield (1991) based on Meybeck (1976)

Overall, high rates of chemical denudation are found in humid mountainous regions, where high relief is coupled with high runoff (Table 7). Minimum rates tend to be recorded in semi-arid regions where runoff is very low (although concentrations of dissolved load may be high), and in high latitude lowland terrains where runoff and solute concentrations are low. In some basins, especially those in a predominantly humid lowland environment, chemical denudation exceeds mechanical denudation. The other extreme are basins where extremely high sediment yields mean that chemical denudation represents less than 5 per cent of total denudation. Proportionally chemical denudation tends to become lower in drainage basins experiencing higher total denudation rates (Table 7).

Chemical denudation and global climate change

Given chemical denudation is an important control on the biogeochemistry of ecosystems, its study has implications not only for landform development but also global environmental change, notably issues of water quality, watershed acidification, nutrient cycling, and the greenhouse effect. As described above, chemical denudation is influenced by climate. Over geological time periods, however, chemical weathering also has a significant influence on global climate. During the weathering of carbonates and silicates, atmospheric CO_2 is taken up and converted to dissolved HCO_3^-. The HCO_3^- after delivery to the oceans by rivers can be stored in the form of carbonate minerals or organic matter in sediments. Either way there is a net loss of CO_2 from the atmosphere. Given CO_2 is a greenhouse gas, any changes in its concentration affects radiative exchanges in the Earth's atmosphere (Berner *et al.* 1987). By way of example, increased rates of chemical weathering associated with the Himalayan–Tibetan uplift have been suggested as a primary cause of the late Cenozoic ice ages (Raymo and Ruddiman 1992).

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SEE ALSO: weathering; weathering and climate change

CATHERINE SOUCH

CHEMICAL WEATHERING

The biogeochemical alteration of the Earth's surface and associated processes are called WEATHERING. These processes are usually separated into chemical, physical and biologic weathering for discussion. In reality, these processes are not mutually exclusive. Chemical weathering is the process by which chemical reactions such as hydrolysis, hydration, oxidation-reduction, ion exchange, solution and organic reactions transform rocks and minerals into new chemical combinations that are stable under conditions at or near the Earth's surface. Chemical weathering begins as thermodynamically unstable minerals adjust to the surrounding environment. Rocks and minerals that are not in equilibrium with near-surface conditions of temperature, pressure and water begin to alter to new products that are chemically more stable in the near surface.

Chemical weathering processes are many. The ability to measure these processes has progressed over the years as newer technologies and interdisciplinary research have led to discoveries at all scales from the molecular to the macroscale. Although chemical weathering occurs at many different temperatures and pressures this discussion will focus on a few basic concepts common to weathering under near-surface conditions.

The resistance of rocks and minerals to chemical breakdown influences the stability of individual mineral species in the environment. This stability is related to several mineral properties including cleavage and fracture patterns, particle size and specific surface, solubility and the relative stability of the surrounding environment. Structurally, the resistance to weathering increases as the complexity of silicate linkage increases, particularly the number of shared oxygens, from independent tetrahedral structures (e.g. olivine) to single chain silicates (e.g. enstatite, a pyroxene) to sheet silicates (e.g. talc) to continuous framework silicates (e.g. quartz). A weathering sequence that illustrates this concept for common rock-forming silicate minerals is illustrated in Figure 23. Stability of minerals increases from top to bottom. Additional guides to mineral stability are discussed in Ritter (1986).

Other factors being equal, minerals formed in environments resembling those in which weathering takes place will be the most resistant. This concept is based on thermodynamic principles. For example, olivines and calcium plagioclase feldspars form at higher temperatures and pressures and weather more rapidly than muscovite and quartz-rich minerals which form at lower temperatures. These latter conditions are more similar to near-surface weathering conditions.

Chemical weathering processes

Solution occurs when a mineral dissolves to form ions or dispersed colloidal molecular units. It is one of the simplest of weathering processes. Bicarbonate ($2HCO_3^-$) is derived from the dissociation of carbonic acid (H_2CO_3) that in turn formed from the dissolution of carbonate rock and atmospheric CO_2 dissolved in water:

$$H_2O + CO_2 + CaCO_3 \leftrightarrow Ca^{++} + 2HCO_3^-$$

Bicarbonate is one of the most abundant anions in weathering systems. Its weathering effects have been studied in detail relative to limestone KARST systems. Bicarbonate ions can also form from dissolution of CO_2 in plant and microbial respiration processes:

$$H_2O + CO_2 \leftrightarrow H_2CO_3 \leftrightarrow H^+ + HCO_3$$

HYDROLYSIS is the reaction of compounds with water to produce a weak acid or weak base. Water molecules are attracted to surfaces of

Olivine (Mg, Fe)	Calcic plagioclase (Ca, Al)
\	/
Augite (Ca, Mg, Fe, Al)	Calcic-alkali plagioclase (Ca, Na, Al)
\	/
Hornblende (Ca, Na, Mg, Fe, Al)	Alkali-calcic plagioclase (Na, Ca, Al)
\	/
Biotite (K, Mg, Fe, Al)	Alkali plagioclase (Na, Al)
\	/

Potash feldspar (K, Al)

|

Muscovite (K, Al)

|

Quartz (Si)

Figure 23 Weathering of common rock-forming silicate minerals
Source: Data from Goldich, S.S. (1938) A study in rock weathering, *Journal of Geology*, 46, 17–58

minerals due to the attraction of polar water molecules to the polar surfaces of many minerals. Here, forsterite hydrolysis produces silicic acid:

$$Mg_2SiO_4\ \text{(forsterite)} + 4H_2O \leftrightarrow 2Mg^{++} + 4OH^- + H_4SiO_4\ \text{(silicic acid)}$$

Natural waters usually contain dissolved CO_2 so reactions often contain carbonic acid as well. A more complete way to write the above reaction in a natural system would be:

$$Mg_2SiO_4\ \text{(forsterite)} + 4H_2CO_3 \leftrightarrow 2Mg^{++} + 4HCO_3^- + H_4SiO_4$$

H^+ (from water) can replace other ions such as K^+, Ca^{++} and Na^+ in mineral structures. The H^+ disrupts the structural bonds. If the H^+ is smaller than the cation it replaces, physical strain occurs in the mineral which in turn accelerates weathering. For example, microcline feldspar reacts with water and loses a potassium ion:

$$KAlSi_3O_8\ \text{(microcline feldspar)} + H_2O \rightarrow HAlSi_3O_8 + K^+ + (OH)^-$$

Lowering the pH increases hydrolysis because the number of H^+ ions in solution increases. For example, organic matter decomposition adds H^+ and speeds hydrolysis as do many other biologic processes such as nutrient uptake, nitrification and sulphur oxidation. Warm temperatures have an effect similar to lowered pH. Higher temperatures increase the dissociation of water molecules and provide additional H^+, potentially increasing hydrolysis in a system. Thus the microcline feldspar in the above example should weather more quickly in a warmer rather than a cold climate and in an acid rather than a more neutral environment.

HYDRATION adds water molecules to mineral structures but the water does not dissociate as in hydrolysis. Gypsum is a hydrated form of anhydrite. The reverse reaction is dehydration:

$$CaSO_4\ \text{(anhydrite)} + 2H_2O \leftrightarrow CaSO_4 \cdot 2H_2O\ \text{(gypsum)}$$

Although quartz is a resistant mineral, under specific conditions it can dissolve by hydration:

$$SiO_2\ \text{(quartz)} + 2H_2O \leftrightarrow H_4SiO_4$$

Some minerals may expand during hydration. Commonly smectite hydrates and dehydrates when water molecules enter or leave interlayers, respectively. In an expanded condition, minerals are more porous and become more susceptible to additional weathering.

Ion-exchange reactions are important and are usually related directly to clay mineral weathering and other secondary minerals because these minerals have a high capacity for exchange within the interlayer and with surface ions. During exchange, the basic structure of the mineral is unchanged, but interlayer spacing varies with each cation absorbed into the interlayer. This mechanism has a unique outcome for clay minerals in that the alteration of one clay mineral may produce another. For example, under certain circumstances smectite may form from illite with the loss of interlayer K^+. Ion exchange is an important factor in biogeochemical reactions of rocks and sediments with organic matter and colloids. Ion exchange can also occur in the initial weathering of primary minerals such as silicates.

Ion mobility is key to primary mineral weathering. Hudson (1995) discusses an update of Polynov's 1937 ion mobility series that ranks major elements from very mobile (I) to relatively immobile (V):

$$Cl > SO_4 > Na > Ca > Mg > K > Si > Fe > Al$$

where mobility phase I is Cl and SO_4, II is Na, III is Ca, Mg and K, IV is Si and V is Fe and Al. Mobility depends on charge and charge density.

In a strongly leaching environment only phase V elements would remain. As an environment became drier, phase IV through I elements would become increasingly abundant. For example, gibbsite ($Al[OH]_3$) is a common aluminium hydroxide in sediments and soils assumed to be in the latter stages of weathering where leaching conditions and free drainage occur. This would be equivalent to phase V ion mobility. At this point, silica has been so thoroughly removed from the system that phyllosilicates can no longer form. Aluminium hydroxide-rich sediments are associated with tropical environments today and in the weathering profiles of bauxite deposits of ancient silica-depleted rock systems.

OXIDATION and REDUCTION equilibria, also known as redox reactions, take place when an atom or element gains or loses net charge; oxidation, the loss of electrons and reduction, the gaining of electrons. The availability or absence of one electron acceptor leads to the reduction of another element. Elements must have at least two viable oxidation states to be involved in redox reactions. Only about six elements, oxygen, iron, manganese, sulphur, nitrogen and carbon, are abundant enough in the natural environment to take part in common redox reactions of the near-surface environment. Oxygen plays a role in most oxidation processes. In the reaction below, ferrous iron derived from the hydrolysis of an iron-bearing silicate, is oxidized from +2 to +3 oxidation state to form haematite. Oxygen is reduced from 0 to −2:

$$2Fe^{++} + 4HCO_3^- + \tfrac{1}{2}O_2 \leftrightarrow Fe_2O_3 \text{ (haematite)} + 4CO_2 + 2H_2O$$

Haematite is stable in many environments but goethite, also a ferric iron component and primary constituent of limonite, may occur with the addition of more moisture:

$$Fe_2O_3 \text{ (haematite)} + H_2O \leftrightarrow 2HFeO_3 \text{ (goethite)}$$

Oxidation of pyrite, FeS_2 to iron hydroxides or sulphates and sulphuric acid on exposure to water and oxygen has detrimental consequences. This reaction, often occurring in materials adjacent to mine sites, is a common cause for the sterile biologic conditions in sediments drained by acid waters. The term acid mine drainage is applied to these waters in which the pH can drop below 2.

Oxidation of organic carbon is often due to micro-organisms that play a major role in expediting redox reactions. An example of an organic oxidation reaction in which carbon dioxide is formed is:

$$C_6H_{12}O_6 + 6O_2 \leftrightarrow 6CO_2 + 6H_2O$$

The carbon dioxide formed is available for solution and hydrolysis reactions.

Chelation (see CHELATION AND CHELUVIATION), a form of metal complexation, is the reaction between a metallic ion and a complexing agent, usually organic, resulting in the formation of a ring structure that encompasses the metallic ion effectively removing it from the system. Hydrogen is often released during the process and becomes available for hydrolysis reactions. Chelating agents in contact with rocks or minerals can cause significant weathering (Berthelin 1988). For example, lichens and mosses remove cations from silicate minerals and may produce dissolved or amorphous silica. Some breakdown of minerals occurs from reactions with organic acids produced at the root tips of plants or produced by bacteria acting on decaying material.

Chemical weathering products

Chemical weathering results in either congruent dissolution, in which the material goes completely into solution, or incongruent dissolution, in which at least some weathering products may form new minerals (neoformation or synthesis) or leave a residue or precipitate. If limestone dissolves completely and releases Ca^{++} and HCO_3^- ions into aqueous solution it is a congruent dissolution. However, most limestones are not pure $CaCO_3$ and leave a residue. During chemical changes, particle size decreases, surface area increases and constituents continue to dissolve into aqueous weathering solutions. Water is often the transferring agent and its activity is important.

Berner (1971) and Berner and Berner (1996) emphasize the importance of water flow as a control factor on the intensity of weathering. Berners' example suggests that at moderate flow rates albite alters to kaolinite but at higher flow rates silicic acid is removed so quickly that gibbsite rather than kaolinite may form. When flow rates were very slow, material was removed slowly and if magnesium was available, the product was montmorillonite. This suggests that climate and relief control weathering products. The mineralogy of the rock weathering and the chemical composition of weathering solutions are two additional determining factors. Chadwick *et al.* (2003) present a biogeochemical model for an arid to humid climosequence on Kohala Mt., Hawaii. They found that where mean annual precipitation is high and total sediment pore space is annually full, leaching of soluble base cations and silica is nearly complete. At lower precipitation inputs, leaching losses are progressively lower. Secondary mineral weathering was controlled by metastable non-crystalline weathering products rather than soil solution composition.

Weathering products may be grouped into four categories: (1) soluble constituents; (2) residual primary minerals unaffected by weathering reactions; (3) new stable minerals produced by weathering reactions; (4) organic compounds. Soluble constituents are those that remain in solution at near-surface conditions. Three primary groups of residual minerals remain in weathered soils: (a) phyllosilicate clay minerals; (b) very resistant end products such as sesquioxides of Fe and Al; (c) very resistant primary minerals such as quartz, zircon and rutile. Each group contributes less as weathering progresses. In highly weathered soils and sediments of the humid tropics or subtropics, Al and Fe oxides and low-activity clay minerals with low Si/Al ratios may be all that remains of the original primary minerals. Feldspars, mica, amphiboles and pyroxene minerals alter to clay minerals through hydrolysis, hydration and oxidation. For example, biotite mica weathers as Fe^{++} oxidizes, K^+ leaves the structure to maintain neutrality, the structure begins to weaken and soluble cations in solution such as Ca^{++}, Mg^{++} or Na^+ replace the remaining K^+. A new phyllosilicate such as vermiculite or montmorillonite forms.

Phyllosilicates are commonly stable mineral products of weathering. They are specific clay minerals occurring primarily in the clay-size fraction of a material (see Moore and Reynolds 1997). Phyllosilicates strongly influence the chemical as well as physical properties of sediments, in part due to their unusually small particle size and resulting high surface area but also, as described earlier, due to cation exchange characteristics uniquely related to their crystal structures (see Dixon and Weed 1989; Moore and Reynolds 1997). Linus Pauling (1929, 1930), Kelley (1948) and Grim (1962) were some of the first individuals to recognize the unique chemical properties of phyllosilicate clay minerals.

Chemical weathering and landscapes

Measurement of the total amount of chemical weathering is important to a geologist or geomorphologist because it can provide some estimate of landscape evolution. Although both physical and chemical denudation affects landscapes on a catchment or global scale this discussion centres on chemical weathering. Chemical denudation can be calculated from dissolved stream loads and corrected for atmospheric input because most ions in water come from weathering reactions (Berner and Berner 1987). Annual load is multiplied by annual discharge and divided by basin area. Berner and Berner (1987) have calculated a world average. Garrels and MacKenzie (1971) ranked chemical denudation by continent: Europe $>$ North America $=$ Asia $>$ South America $>$ Africa $\gg$ Australia. Degree of weathering is often calculated on the basis of total chemical analyses comparing fresh parent rock with saprolite or soil derived *in situ* from it. Birkeland (1999) presents a good summary of this approach.

During physical weathering in an open system, the landscape is generally lowered volumetrically because solids are removed but during chemical weathering the landscape may increase volumetrically. Ions removed from a weathering rock mass might be reflected in a bulk density change such that a geomorphic surface is unchanged or is even raised. For example, when a soil forms from rock, the bulk density will decrease, sometimes by $0.5\ g\,cm^{-3}$ or more. This results in an overall volumetric expansion (Birkeland 1999). Brimhall and others (1991) developed a method for assessing chemical change during weathering that gives values for volume change as well as for losses or gains in mass. In some cases this expansion is the catalyst for increased physical weathering. Several researchers have suggested that the formation of grus from granite follows these steps (e.g. Wahrhaftig 1965; Nettleton *et al.* 1970).

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SEE ALSO: dissolution; leaching; solubility; weathering

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CHENIER RIDGE

Chenier ridges (cheniers) are sandy or shelly elongate BEACH RIDGES, differentiated from other sand or shell beach ridges by the fact that they are perched on and separated laterally from other cheniers on a chenier plain, by fine-grained, muddy (or sometimes marshy) sediments. Other types of barrier beach plains can be mistaken for cheniers if the presence of underlying and interspersing muddy sediments is not adequately determined (normally by coring). Chenier ridges frequently bend landward at the downdrift end, and branch in a fan-like fashion. The name derives from the French word chêne, meaning oak, which grows on the Louisiana USA chenier ridges. Cheniers can be up to 6 m high, tens of kilometres in length, and hundreds of metres wide. Chenier plains can be tens of kilometres wide. Cheniers are found on generally low wave energy, low gradient, muddy shorelines, in areas where there is an abundant sediment supply. They are frequently associated with river deltas and bayhead situations. Although reported at high latitudes, most examples occur in tropical or subtropical locations. Augustinus (1989) provides an overview of examples and presumed examples of cheniers. Among the most reported examples are: the west Louisiana and Texas coast; Suriname, Guyana and French Guiana; the Gulf of California; New Zealand; northern Australia; east China.

Local variations in sediment supply (such as periods of different river discharge) have been suggested as the likely cause of alternate mudflat progradation and chenier ridge deposition (Otvos and Price 1979), although synchronous development of mudflat and chenier ridges has also been reported (Woodroffe *et al*. 1983; Woodroffe and Grime 1999). Periods of higher wave energy, however, are generally regarded as providing the means by which coarser sediments (including shells) are winnowed out for accumulation in the chenier ridge, with these sediments then moved landward by wave action and OVERWASHING. Some authors argue that ‘true cheniers’ must result from transgressive processes; however, Otvos (2000) believes the term is appropriate for both stranded (regressive) and transgressive cheniers (see TRANSGRESSION). Otvos (2000) also argues that beach ridges fronting chenier plains must become isolated from the sea and inactive by the deposition of mudflats on their seaward side before they can be considered as cheniers. Chenier plains can provide a sensitive record of changes in sediment supply, sea level and environmental processes.

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SEE ALSO: beach ridge; overwashing; raised beach; transgression

KEVIN PARNELL

CHRONOSEQUENCE

The term is used to describe a series of soils that reflect the importance of time for soil formation. *Inter alia*, young soils will differ from mature soils in the degree of weathering of the soil parent material, the development of the soil horizons and the abundance of secondary minerals.

Because the time spans involved in soil development are beyond the time frame of direct observation, usually the development of soils of different age are compared. A chronosequence is thus a sequence of related soils that differ from one another in certain properties primarily as a result of the time available for soil formation. Classical examples are the soils developed on the different members of a flight of terraces, where – except for time – all soil-forming factors (as parent material, landform, climate, etc.) should be rather similar.

Different types of chronosequences can be distinguished: (1) post-incisive, (2) pre-incisive, and (3) time-transgressive. The most frequently studied is case (1) – the example mentioned above – where soils evolve on a sequence of surfaces of different age. In (2) soils that began to develop on a particular surface at the same time, but that were subsequently buried at different times at different places, form a chronosequence. Case (3) relates to a vertical stacking of sediments and PALAEOSOLs, i.e. soils that formed on the same place, but that have been buried after differing periods of development.

Chronosequences have been used to establish quantitative descriptions of soil changes with time, called chronofunctions, and to use the degree of soil development for estimating soil age. To allow for quantitative estimates, soils on dated surfaces (see DATING METHODS) are investigated and numerical indices, such as eluvial-illuvial coefficients and soil development indices, have been developed. There are limits to the range over which chronofunctions can be applied. The rates of development of most soil properties decrease with time; once this degree of development has been achieved, further inferences regarding time cannot be made. In addition, many more complex functions with clear thresholds are involved in soil development. Establishment of chronosequences is difficult in many cases because with the passage of time other soil-forming factors usually also change – as is most clear in the case of climate (see PALAEOCLIMATE). It is often also difficult to rule out the influence of soil disturbances and soil erosion.

Chronosequences have played an important role in establishing relative soil chronologies, which in turn have been used to establish stratigraphic relationships for different geomorphic surfaces. This is especially important where due to the lack of suitable methods or materials modern chronometric dating techniques cannot be applied.

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SEE ALSO: catena; soil geomorphology; weathering

ANDREAS LANG

CIRQUE, GLACIAL

Definition and Form

Cirques, also known as corries, coves, combes or cwms, are hollows formed at glacier sources in mountains and partly enclosed by steep, arcuate slopes (headwalls) (Plate 22). Cirque formation requires deepening of the floor by glacial plucking and abrasion, plus glacial removal of plucked or fallen rock encouraging continued headwall retreat. These are aided by basal slip and rotational flow of steep glaciers.

A well-developed 'armchair cirque' has a gently sloping floor and a steep headwall (giving profile closure). At least some of the floor should be gentler than 20°. The headwall curves around the floor, giving plan closure. Ideally, the floor ends in a distinct threshold beyond which the slope steepens, but this may be absent in a trough-head cirque. The headwall should exceed the angle of talus (about 31°–35°) at least in part. We can draw the boundary between headwall and floor at an angle of some 27° (a 2 mm spacing of 10 m contours on a 1 : 10,000 map). A similar gradient can be used to define the cirque crest at the top of the headwall if there is a gentler slope above. It is useful to define a 'cirque focus' in the middle of the threshold. A line from there to the top of the headwall, dividing the cirque into two halves equal in map area, to left and to right, is the median axis: this is used to measure length and overall aspect (Evans and Cox 1995).

Plate 22 East-facing cirques in Ordovician volcanic tuffs on the ridge south of Helvellyn, English Lake District; from left to right, Cock and Ruthwaite Coves, Hard Tarn (a smaller hollow) and Nethermost Cove. The cirques hang above Grisedale trough

Plate 23 North-facing cirque in Triassic metamorphic rocks on Mount Noel (2,600 m), south of Bralorne in the Coast Mountains of British Columbia. The small glacier is a remnant of a larger Little Ice Age cirque glacier which formed the two sharp lateral moraines at bottom right (August 2000)

Cirque size varies over an order of magnitude, and provides a characteristic scale to glaciated mountains: cirques are scale-specific landforms, averaging around 700 m long and broad, and a few hundred metres deep. Overall centre-line gradients, approximating those of glaciers filling the cirques, vary from 5° to 50°, with means commonly 20° to 25° (length/depth 2.1 to 2.8).

Cirque form is simple in plateau areas, more complicated in high-relief areas of coalescing cirques, and very difficult to define where modified by overriding ice sheets. Valley-head and valley-side cirques are often distinguished, but a more important distinction is between 'armchair cirques', the type normally described, and 'high-alpine cirques' found in high massifs of the European Alps and in coastal regions of British Columbia, Washington and Alaska (Plate 23). 'High-alpine' cirques are shallow and have steep, straight, abraded apron-like floors (20°–31°) hanging above troughs along which ice was rapidly evacuated. Classic armchair cirques are deeply concave in both profile and plan (concave contours). For both types, the concave break in slope between headwall and floor prevents a good fit to simple equations.

Processes

Ideas on processes of cirque development have changed considerably over time. The importance of glacial erosion was clearly established in the 1870s by Gastaldi and Helland for cirques in the Alps and Norway. Although glacial protectionists argued for fluvial or tectonic origins into the twentieth century, the existence of deep, rounded rock basins in many cirques could not be explained except by glacial erosion. Nevertheless it was accepted that processes acting around and under snowpatches (NIVATION) widened initial hollows before glaciers could become established: cirque widening formed the basis of a 'cycle of mountain glaciation' proposed by W.H. Hobbs. After 1906, W.D. Johnson's observation that frost action was active in and above the BERGSCHRUND – the initial crevasse as the glacier accelerates away from a cirque headwall – was accepted as an essential process of cirque development. Although cirques were used as evidence of former glaciation and in the reconstruction of former snowlines, their development by essentially periglacial processes was emphasized throughout the first half of the twentieth century.

Neither nivation nor frost weathering, however, explain the deepening of cirques by erosion of their floors. After a brief flirtation in the 1940s with the hypothesis of extrusion flow – the unrealistic idea that soft basal ice would be squeezed forward by the weight of overlying ice, without carrying the overlying ice forward – geomorphologists found a better way of obtaining fairly high basal ice velocities. Observations in the Jotunheim (Norway) in the late 1940s showed that steep glaciers banked against cliffs acted rather like landslides, and moved over their beds by rotational slip (McCall, and Grove; in Lewis 1960). In this way the glacial abrasion and plucking advocated by Helland became easier to explain. This also implied that basal ice must be wet, i.e. 'warm' – at its melting point.

Rotational slip, basal abrasion and plucking, and frost weathering around the bergschrund or upper glacier margin do not, however, cover the whole story of cirque development. Observations by Battle showed that temperatures within bergschrunds varied too slowly for the daily or seasonal frost cycle to be effective. Gardner (1987) showed that the *randkluft* or *rimaye* – the upper margin of a glacier against a cliff – is a more likely site for frost action, and migrates up and down the headwall over time. Whalley has pointed out that the stress field in high cliffs undercut by glaciers causes instability. With or without the help of frost action, stress concentration and release cause headwall collapse by rockfall and rock avalanche (see STURZSTROM). Many rock avalanches, both historic and older, are from cirque headwalls (Evans 1997). Active rockfall onto cirque glaciers can be observed today, especially as glacier wastage has increased the area of exposed headwall above.

While this mechanism helps to account for headwall retreat, fuller understanding of subglacial erosion has followed study of water pressure variations in glacier boreholes. Hooke (1991) and Iverson (1991) have shown that these are frequent and of high magnitude where crevasses (including the bergschrund) permit meltwater to reach the bed. On steep lee (down-ice) slopes, water-filled cavities tend to open. When water pressure falls, there is a delay before pressure falls in cracks in the rock: this aids crack extension. When water pressure rises, ice velocity increases and it is easier for joint-bounded blocks to be carried forward (entrained) in the basal ice. These neatly interrelated mechanisms provide what may well be the most important process for glacial plucking (quarrying).

Much more is now known about the frequent and large variations of climate during the Quaternary. This makes slow development with a snowpatch or glacier of a given size unlikely: a nivation phase is soon overtaken by a glacial phase. Wet-based glaciers erode much more rapidly than snowpatches. Erosion can be effective even where the margins of a glacier are frozen to the bed; Bennett *et al.* (1999) have described very rapid erosion by thrusting of sedimentary bedrock stressed by the transition from a sliding to a frozen subglacial boundary. Such 'polythermal' glaciers are common today in Svalbard, and in Sweden (Richardson and Holmlund 1996), and may deepen cirque floors as the warm, deep central ice slides over its bed. Adjacent rock may break down by growth and thawing of ice lenses as the zero isotherm migrates over the long term.

Cirque erosion thus requires basally sliding ice to quarry and abrade the bed, steepening the margins until they collapse. Erosion is aided by concentration of snow accumulation high on the glacier (Figure 24), and by gradients between 12° and 26°, both of which encourage rotation; also by rapid variations in basal water pressure. Steepened headwalls collapse by rockfall or rock avalanche, often on deglaciation but sometimes earlier or later.

Cirque development

Mountain glaciers form in concavities, deepening, widening and simplifying these to form glacial cirques. Starting with glacial occupation of any large hollow, their floors are deepened, headwalls retreat and concavity is increased. Initial concavities include gully heads, gully junctions, landslide scars, structural benches and volcanic craters. Diversity of initial concavities provides a broad range of poorly developed cirques: continued glacial erosion: (1) increases headwall gradient, (2) reduces floor gradient, (3) increases length, width and depth (amplitude), and (4) increases plan closure by eroding more deeply into the mountain mass, so that the headwall curves around the floor.

Rotational slip, deepening the floor, is favoured in cirque glaciers, but erosion continues even if glaciers extend beyond cirques. Gordon (1977) proposed a model of cirque development whereby cirques lengthen, broaden, deepen and increase their concavity as they enlarge. Nevertheless, correlations between the measures of development (headwall gradient, inverse floor gradient, and plan closure) are very weak, implying that cirques develop along varied paths and the influence of initial site remains important. Thus cirque form is very diverse.

Larger cirques are better developed on all criteria: they are also flatter, i.e. horizontal dimensions increase more rapidly than vertical. This has been interpreted as allometric development (see ALLOMETRY), although it relates to spatial rather than temporal variation (Olyphant 1981). A peak or a high headwall on the equatorward side of a cirque helps preserve snow: a pass to windward funnels more snow in (Graf 1976). Better developed cirques are thus more effective in sheltering glaciers. This gives a positive feedback, so that a steady state is unlikely unless surrounding ridges are lowered at the same rate as all parts of the cirque.

Where initial hollows are closely spaced, the formation and retreat of cliffed headwalls eventually sharpen the intervening ridge into an ARÊTE. Gullies and structural irregularities in cliffs lead to marked rises and falls in arête crests, giving a series of GENDARMES. Since cirque glaciation is commonly asymmetric, arêtes arise from lateral intersection, i.e. two sidewalls retreating into each other. When the EQUILIBRIUM LINE OF GLACIERS falls sufficiently for glaciers to form on opposing slopes, e.g. east and west, steeper arêtes flank a deep parabolic col of 'interosculation'.

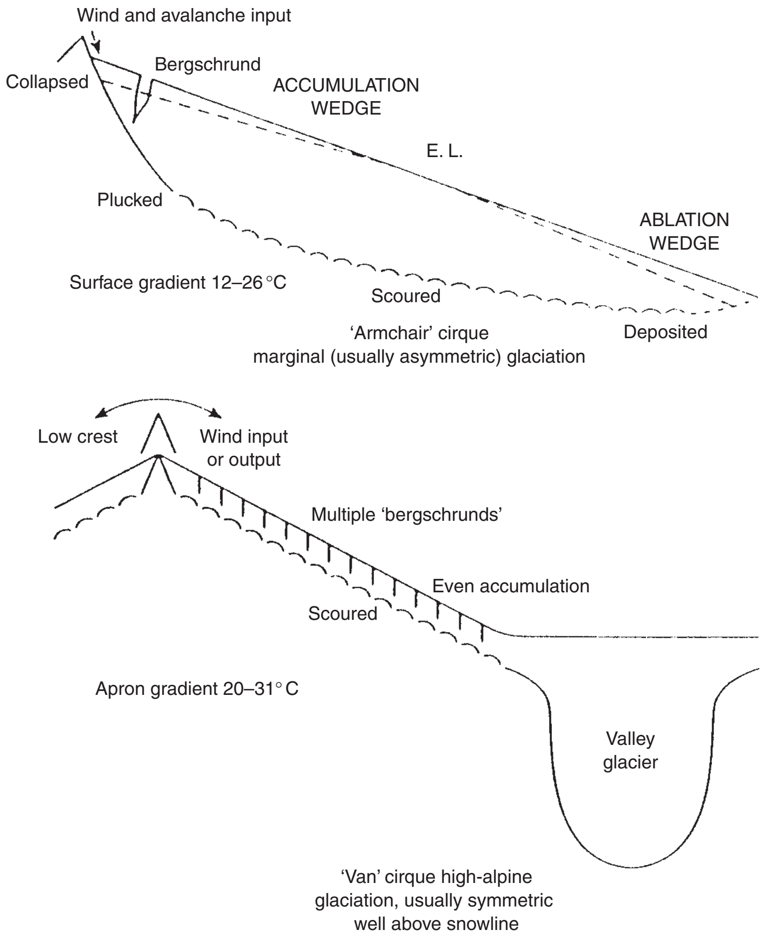

Figure 24 A model of the contrast between classical armchair cirques (above) and high-alpine cirques (from Evans 1997: 162)

Lateral enlargement continues, perhaps by coalescence, with no clear upper limit so long as a mountain mass remains. In the Antarctic, troughs and cirques have developed over a much longer period of glaciation and some cirque complexes are very large: forms of local glaciation developed in the earlier part of this period, and many are now 'fossilized' under very cold ice.

Many areas have suffered both local and ice sheet glaciation and cirques away from ice divides have been further modified (degraded) by overriding ice. On deglaciation, cirques are degraded by subaerial mass movements, talus accumulation and gullying: these reduce headwall gradient and obscure the floor, but do not affect plan closure.

Relation to geology, climate and topography

Cirques form on all rock types, but postglacial degradation of headwalls is most likely on the weakest rocks such as shales. The expected effect of rock liability to abrasion has not been demonstrated, but the importance of joint sets is often noted (Haynes 1968). Inward-dipping joints or beds favour excavation of a rock basin, and these are more common on crystalline rocks. Simple, rounded cirques are found on homogeneous or frequently alternating rocks, e.g. flat-lying volcanic and sedimentary rocks. Greatest distortions in form come from single major contrasts, for example, juxtaposition of limestone and shale or quartzite and gneiss, giving a floor or a steeper side on the less-jointed rock. Cirque headwall heights and gradients should relate to ROCK MASS STRENGTH: results to date, however, show considerable scatter.

Well-developed cirque floors relate to the former snowline (Equilibrium Line), for example in being much lower on windward sides of major mountain ranges (Derbyshire and Evans 1976). This means they relate to snowfall rather than to the freezing level. More locally, snow is blown to leeward slopes and preserved longer on shady slopes, and cirques are more frequent on corresponding aspects.

In some regions, cirques are separated from each other by plateau areas or rolling topography: elsewhere they intersect and form more arêtes and horns. In the early twentieth century, influenced by the Davisian model, this was regarded as a developmental sequence. It is more likely that these contrasts relate to regional topography (Gordon 2001), including relief and drainage density, due ultimately to tectonic setting and climatic history.

Further work

Most morphometric studies have been confined to single regions, or based on selected cirques: consistent results from complete populations of cirques are needed, to establish variations between different regions and to start accounting for variations in relation to climate, geology and topography.

We have little information on periods of time required for cirque development, though currently observed rates of glacial erosion are adequate to erode cirques in a few hundred thousand years – only a proportion of the Quaternary. Headwall retreat per glaciation would be some 10 m, and floor deepening a few metres. New techniques of exposure dating of surfaces only tell us when ice last disappeared; techniques such as fission-track dating give some idea of rock uplift history, but with very broad error margins. More precise dating approaches are needed to provide specific chronologies of cirque development. In small regions cirques may have developed simultaneously (Evans 1999). But cirques outside areas of ice sheet glaciation could develop at glacial maxima, whereas those within probably developed largely as ice built up.

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SEE ALSO: aspect and geomorphology; freeze–thaw cycle; glacial erosion; glacial protectionism; glacier

IAN S. EVANS

CLAY-WITH-FLINT

The chalklands of southern Britain (and northern France) are mantled over extensive areas by a group of deposits called clay-with-flint (*argile à silex*) (Laignel *et al*. 2002).

They are highly variable in composition, ranging 'from heavy reddish brown clays with large unworn flint nodules to almost stoneless yellow or white sands, yellowish to reddish brown silt loams, brightly mottled (red, lilac, green and white) stoneless clays, and beds of rounded flint pebbles' (Catt 1986: 151). Early English geologists tended to regard them as an insoluble residue, left after a long period of dissolution and weathering of the chalk. However, although some of the constituent material of clay-with-flint may have been derived from this source, it is not an adequate explanation of the variability of the material nor of the presence of miscellaneous types of clay, sand and flint shape. Much of it is probably derived and reworked from Palaeogene beds and other Cenozoic deposits, as Jukes-Browne (1906) so astutely recognized.

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A.S. GOUDIE

CLIFF, COASTAL

A cliff is a steep slope (usually >40°, often vertical and sometimes overhanging), exposing rock formations (Plate 24). Most coastal cliffs have been produced by wave ABRASION at the cliff base, but some have been formed by faulting or earlier fluvial or glacial erosion.

Cliffs cut in unconsolidated formations are known as Earth cliffs (May 1972), and those at

Plate 24 Chalk cliffs at Seven Sisters, Sussex, England, retreat as the result of wave abrasion, but are also influenced by solution, bioerosion and rock falls due to freeze–thaw effects and groundwater discharge

the seaward ends of glaciers ice cliffs. Hard rock cliffs, which change very slowly, have been relatively neglected in coastal research. In humid regions soil and vegetation may cover coastal slopes, except on actively receding cliff faces. Vegetated bluffs are not necessarily stable: on the Oregon coast they are cut back as cliffs during occasional severe storms or tsunamis, and then revegetate.

Cliffs rising 100–500 m above sea level are termed high cliffs, and those >500 m (as in Peru and western Ireland) megacliffs (Guilcher 1966). Cliffs less than a metre high are termed microcliffs.

Coastal cliffs recede as the result of basal marine erosion accompanied by subaerial erosion of the cliff face. A sharp angle or notch (see NOTCH, COASTAL) at the cliff base generally indicates active marine erosion. Some cliff profiles are of uniform gradient, others concave or convex, or a combination of these. Concave profiles occur where subaerial erosion exceeds marine erosion and convex profiles where marine erosion has been dominant (Emery and Kuhn 1982), but cliff profiles are also related to the position and inclination of resistant strata. A resistant caprock forms bold cliffs, hard outcrops in the cliff face produce ledges, and a resistant formation at the cliff base slows marine erosion (Figure 25A). A seaward dip facilitates landslides, horizontal strata may form stepped profiles and a landward dip produces an escarpment cliff (Figure 25B). Slope-over-wall profiles may be related to weak above resistant formations or an undercut seaward dipslope (Figure 25C: 1, 2). Joints, bedding planes, faults and intrusions influence cliff morphology, and lateral changes in lithology result in changes in cliff profiles, as on Triassic sandstones and clays in south-east Devon, England. On limestone coasts marine erosion exposes caves and cauldrons produced by earlier karstic dissection.

Cliff outlines in plan are also related to geological structure, with headlands where resistant formations outcrop at the cliff base and bays where weaker formations are excavated by marine erosion; headlands often coincide with ridges and bays with valleys. The Dorset coast, east of Weymouth in southern England illustrates these relationships (Bird 1995).

Cliff-base erosion is achieved by wave quarrying, which dislodges and removes rock material, and abrasion where waves throw sand or gravel against the cliff base. Cliff outcrops may disintegrate as the result of WETTING AND DRYING WEATHERING of surfaces subject to spray, splash and rainwash, or SALT WEATHERING where salt crystallizes from sea splash, notably on arid coasts. Solution by runoff, seepage, spray and sea water contributes to cliff-base erosion, particularly on limestone coasts where distinctive flat-floored solution notches may form, in contrast with sloping ramps where wave abrasion is dominant. Bioerosion (by plants and animals that live on the cliff and shore) also contributes.

Cliff faces may be indurated by calcareous or ferruginous compounds precipitated from groundwater seepage, forming crusts that eventually crack and exfoliate, exposing uncemented rock. Cliff faces are also indurated by carbonates precipitated from sea splash, particularly on headlands. Downwashed sediment may adhere to a cliff face as stalactitic structures, notably on limestone and AEOLIANITE (Hills 1971). By contrast, fine-grained sediment winnowed from a cliff face by onshore winds has been deposited as a cliff-top levee on the Port Campbell coast in south-eastern Australia (Baker 1943).

As a cliff is undercut it may collapse, producing a debris fan below a fresh rock scar. Sediment yield from cliffs depends on the rate of recession and the effects of weathering and erosion. Accumulation of sediment at the cliff base slows recession, but usually the debris fan is dispersed by erosion, and when it has been removed basal undercutting resumes.

MASS MOVEMENTs occur on cliffs where the groundwater load becomes excessive, where stresses develop as the result of freeze and thaw, where a massive caprock exerts pressure on underlying weaker formations, or where there is expansion or base exchange, weakening clay minerals. Breakaways develop at the cliff crest where masses of rock topple down the cliff, and slumping produces irregular topography as rock outcrops disintegrate and material slides, flows or creeps down the slope towards a basal receding cliff. Such cascading systems, with instability transmitted upward to the cliff crest, occur on the Dorset coast (Brunsden and Jones 1980). In Oregon coastal landslides in weathered rock are commoner in winter, when stronger wave action attacks formations saturated by heavy rain, but may also be triggered by tectonic movements or tsunamis generated along the nearby plate edge.

Some cliffs descend to SHORE PLATFORMS cut by marine erosion and weathering processes as the

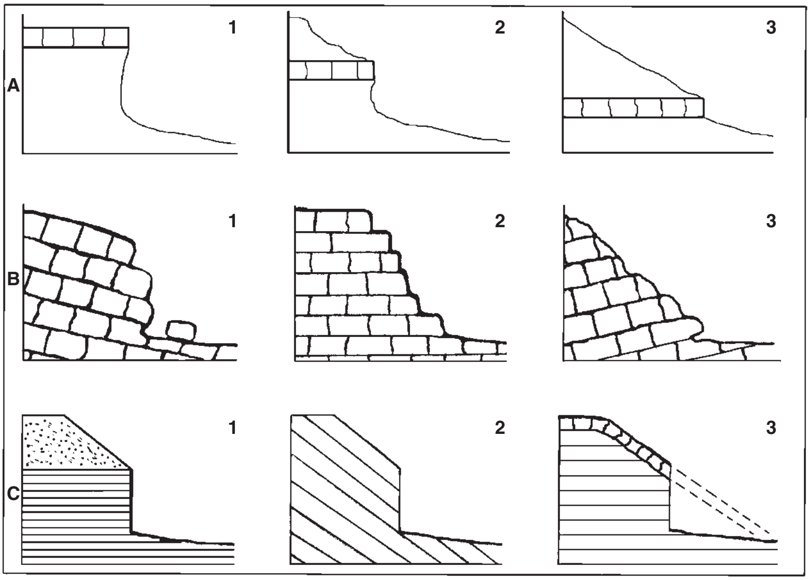

Figure 25 A, the effects of a resistant formation on cliff profiles; B, variations related to the dip of strata; C, slope-over wall cliffs: 1, related to lithology; 2, related to structure; 3, retaining a slope formed by periglacial solifluction

cliff recedes; others are fronted by irregular rocky shores, particularly where the geological formations are of intricate structure with resistant elements; others (plunging cliffs) continue below sea level, either because of partial marine submergence (where they descend to submerged coastlines) or because they formed by faulting, glaciation or vulcanicity.

Some cliffs are actively receding; others are inactive behind persisting basal talus or a prograding beach, or because of lowering of sea level. Inactive cliffs may decline into subaerially shaped slopes which become vegetated. Cliffs stranded by land uplift or sea-level lowering become bluffs behind emerged beaches and shore platforms.

Rates of cliff recession vary with cliff height, rock resistance, structure, weathering and exposure to wave attack. They are usually reported as annual averages, but are generally episodic, related to occasional storms or mass movements. Rapid cliff recession ($>1\,\mathrm{m\,yr^{-1}}$) occurs on soft rock formations, and rates of $>100\,\mathrm{m\,yr^{-1}}$ have been reported on cliffs in volcanic ash and arctic tundra deposits (humates with melting ice), but some hard rock cliffs have shown little or no recession in the period (up to 6,000 years) that the sea has stood at its present level.

Where cliff recession has been slow, features inherited from earlier environments may persist. Examples of this are the slope-over-wall profiles on the Atlantic coasts of Britain, where the slope (which may be convex, a straight bevel or concave) is mantled by earthy gravel (termed Head) formed by periglacial SOLIFLUCTION in cold phases of the Pleistocene, and the wall is a receding undercliff (Figure 25C: 3): the proportion of slope to wall diminishes as exposure to wave attack increases. Active periglaciation forms steep slopes of angular debris on arctic coasts, as on Baffin Island in Canada. In northern Britain and Scandinavia the periglacial slope gives place to slopes formed by glacial erosion or deposition, also undercut by Holocene marine erosion. In the humid tropics slope-over-wall profiles occur where a coastal slope on deeply weathered rock has been undercut by marine erosion, and in arid regions the undercut coastal slope may have been a pediment.

Cliff recession is likely to accelerate (and coastal landslides become more frequent) during a rising relative sea level, and when storminess increases in coastal waters: protective beaches diminish, and wave attack on the cliff base becomes stronger and more sustained.

Human impacts on cliffs include stabilization by the building of basal sea walls or boulder ramparts to halt coastline retreat, the grading, vegetating or concreting of cliff faces, and the introduction of drains to hasten groundwater discharge. By contrast, cliffs become more unstable as the result of the reduction of beaches (when beach sand or gravel are extracted), an increase in groundwater load and levels (when previously dry cliff-top terrain is irrigated) and cliff-top loading by buildings and other structures.

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SEE ALSO: slope, evolution

ERIC C.F. BIRD

CLIMATIC GEOMORPHOLOGY

The part of the discipline that seeks to explain the form and distribution of landforms in terms of climate. It developed during the period of European colonial expansion and exploration at the end of the nineteenth century, when unusual and often spectacular landforms were encountered in deserts, polar regions and the humid tropics. In addition, it was a time when regionalization and classification were major endeavours in geography and cognate subjects. Attempts at climatic, soil and vegetation classifications were being made by scientists like Köppen, Dokuchayev and Schimper. They sought to understand the regional patterns of the phenomena they were classifying, and climate was seen as a major control at their scale of investigation.

In the USA, W.M. Davis recognized 'accidents', whereby non-temperate and non-humid climatic regions were seen as deviants from his normal cycle of erosion and he introduced, for example, his arid cycle (Davis 1905). Some (see Derbyshire 1973) regard Davis as one of the founders of climatic geomorphology, although the leading French climatic geomorphologists Tricart and Cailleux (1972) criticized Davis for his neglect of the climatic factor in landform development. Much important work was undertaken on dividing the world into climatic zones (morphoclimatic regions) with distinctive landform assemblages, in France (e.g. Birot 1968), in Germany (e.g. Büdel 1982) and in New Zealand (Cotton 1942). This version of geomorphology was seen as essentially geographical (Holzner and Weaver 1965).

In the later years of the twentieth century the popularity of climatic geomorphology became less as certain limitations became apparent (see Stoddart 1969).

(1) Much climatic geomorphology was based on inadequate knowledge of rates of processes and on inadequate measurement of process and form. Assumptions were made that, for example, rates of chemical weathering were high in the humid tropics and low in cold regions, whereas subsequent empirical studies have shown that this is far from inevitable.
(2) Some of the climatic parameters used for morphoclimatic regionalization were meaningless or crude from a process viewpoint (e.g. mean annual air temperature).
(3) Macroscale regionalization was seen as having little inherent merit and ceased to be a major goal of geographers, who eschewed 'placing lines that do not exist around areas that do not matter'.
(4) Conversely, and paradoxically, climatic geomorphology had a tendency to concentrate

Table 8 Büdel's morphogenetic zones of the world

Zone	Present climate	Past climate	Active processes (fossil ones in brackets)	Landforms
(1) Of glaciers	Glacial	Glacial	Glaciation	Glacial
(2) Of pronounced valley formation	Polar, tundra	Glacial, polar, tundra	Frost, mechanical weathering, stream erosion (glaciation)	Box valleys, patterned ground, etc.
(3) Of extra-tropical valley formation	Continental, cool temperate	Polar, tundra continental	Stream erosion (frost processes, glaciation)	Valley
(4) Of subtropical pediment and valleys formation	Subtropical (warm; wet or dry)	Continental, subtropical	Pediment formation (stream erosion)	Planation surfaces and valleys
(5) Of tropical plantation surface formation	Tropical (hot; wet or wet–dry)	Subtropical, tropical	Planation, chemical weathering	Planation surfaces and laterites

on bizarre forms found in some 'extreme' environments rather than on the overall features of such areas.

(5) Many landforms that were supposedly diagnostic of climate (e.g. pediments in arid regions or inselbergs in the tropics) are either very ancient relict features that are the product of a range of past climates or they have a form that gives an ambiguous guide to origin.

(6) The impact of the large, frequent and rapid climatic changes of the Quaternary and of the very different climates of the Tertiary has disguised any simple climate–landform relationship. For this reason, Büdel (1982) attempted to explain landforms in terms of fossil as well as present-day climatic influences (Table 8). He recognized that landscape was composed of various 'relief generations' and saw the task of what he termed 'climato-genetic geomorphology' as being to recognize, order and distinguish these relief generations, so as to analyse today's highly complex relief.

Although these tendencies have tended to reduce the relative importance of traditional climatic geomorphology, notable studies still appear that look at the nature of landforms and processes in different climatic settings (e.g. M. Thomas 1994 on the humid tropics; D. Thomas 1998 on arid lands; and French 1999 on periglacial regions). In addition, a concern with GLOBAL WARMING and its geomorphological impact leads to a renewed concern with climate-landform links.

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A.S. GOUDIE

CLIMATO-GENETIC GEOMORPHOLOGY

Climato-genetic geomorphology is the systematic field investigation of landforms in a certain area according to their evolution. Many different methods may be applied, but the basis for it is the observation of an assemblage of rested relief elements. There are two roots to climato-genetic geomorphology. First, the more palaeoforms were acknowledged, it became clear that their systematic investigation was necessary, not only to explain the relief but also to estimate their influence on recent processes. Second, the system of 'klimatische Geomorphologie' was developed. It is the basis for relief forming processes or process fabric, which is applied to the different RELIEF GENERATIONS, the constituents of climato-genetic geomorphology.

The terminology is not very clear for originally the term 'klimatische Geomorphologie' was introduced to differentiate it from tectonic or structural geomorphology. However, it was misleading. Dynamic geomorphology would have been a much better term, as it is the study of processes mainly at the medium scale. 'Climatic geomorphology' is the literal translation, but this has the very different aim of relating landforms to climatic or hydrological data. 'Klimatische Geomorphologie' investigates the relief forming processes in a certain MORPHOGENETIC REGION. Climatic geomorphology looks more for single forms or processes. More or less similar to 'klimatische Geomorphologie' are 'research in a morphoclimatic zone', 'climato-geomorphology', 'forms of morphoclimates' or 'DYNAMIC GEOMORPHOLOGY'. For the evolution of landforms there are the words 'klimatische Morphogenese' (literally climatic morphogenesis) and 'klimagenetische Geomorphologie' (literally climato-genetic morphology). Thus the position of the words geomorphology, climate and genetic might change without any generally agreed special connotations. There is a distinction though between processes and evolution in the terms. It seems that there is a difference in the English and German use of the word 'genetic' in geomorphology, that is, development and historical outline of natural phenomena.

Processes were deduced rather early on in the search for an explanation of landforms, and their relation to exogene (i.e. climate controlled force) was acknowledged. In Europe the work of glaciers was studied on recent examples and similar landforms and deposits were classified accordingly (ACTUALISM). In the west of the USA early research detected the specific processes of the arid zone. Palaeoforms have been increasingly acknowledged since the early twentieth century. The systematic approach to climatic geomorphology dates from 1948. After the Second World War the overseas research of palaeoclimatology (see PALAEOCLIMATE), deduced from morphological features like moraines and solifluction forms, increased. All these research efforts were the basis for the concept of the development of RELIEF GENERATIONS. There are several possibilities for applying this concept besides the explanation of relief evolution. It may serve to control erosion rates, especially their extrapolation and the distinction of human accelerated rates. On the other hand, the extension and preservation of the different relief generations shows the intensity and specific location of recent land forming activity. This is a good basis for applied questions like soil erosion. In connection with ecological studies, relief generations are a basis for the spatial extent of investigated features, e.g. the distribution of soil types.

The recent process fabric is either observed directly or deduced from fresh landform scars after catastrophic events. Similar forms of different size, and sequences, are extrapolated to get an idea about intensity, recurrence and the forming power of special processes and their interrelation. The relation between denudation and linear erosion is investigated as well as between erosion and deposition. This is counterchecked by the known facts of climatic change and tectonic movements, which give an estimate of the change of the processes. As the process fabric is a systematic combination of single geomorphological activities one can ask for completeness of processes as well as forms. A simple example may illustrate this: in the northern foreland of the Alps the rivers now carry sand and show a rather low activity. The slopes are more or less undisturbed. Thus the younger process fabric is not strong and not widespread. There is a large amount which remains unexplained, which from the analysis of the forms is easily classified as moraines. If a soil is developed on them, they are inactive. Moraines are known from the surroundings of recent glaciers in their form and sedimentary structure. Connected landforms are outwash plains in front of them and overdeepening to the rear of the moraines. With this form assemblage the older process fabric can be extended beyond the moraines. Gravel terraces in front of them are of fluvioglacial origin, while lakes to the rear are a sign of glacial scour. There is a feedback in the analysis of forms and processes. Many more details and several stages of the advancing and retreating glaciers have been classified and mapped, e.g. for the Inn Chiemsee glacier by Carl Troll.

With the advancement of knowledge about relief generations it became clear that older forms are widely distributed. There are a few landscapes, like young volcanoes or badlands, which consist only of one relief generation.

Therefore fieldwork should start with the oldest forms and look for the nested younger form assemblage. By interpolation, the younger process fabric is derived. There are the same feedback mechanisms as those named above. This method has two advantages: the existence of remaining unexplained phenomena can be avoided, which one is always inclined to keep as low as possible. Second 'Mehrzeitformen' (i.e. forms shaped in different climates) are more easily detected. It is self-evident that there is a slight change to all the forms of the older relief generation (e.g. the removal of the soil cover of an old plain), but there are a few forms which were noticeably changed by younger processes, like blockfields in the mid-latitudes (cf. RELIEF GENERATIONS).

Climato-genetic geomorphology works not only by analysing landforms, but also uses a wide range of other methods. As a genetic science the connection to the well-developed soil science is especially close. For the tropical zone soil analysis in the field and in the laboratory can solve problems of allochthonous or autochthonous weathering, of relative age, and especially of palaeofeatures. In the humid mid-latitudes, relics of tropical weathering, the periglacial cover of solifluction and loess are counterchecks for the distinction of relief generations. All direct and indirect dating methods are helpful.

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HANNA BREMER

COASTAL CLASSIFICATION

Coastal classification is the grouping of similar coastal features in categories that distinguish them from dissimilar features. The aim is to elucidate the relationships between coastal landforms and processes and to understand coastal evolution. Simple classifications are implicit in the topics identified in chapter headings in coastal textbooks, and when coastal features are categorized and shown on maps of coastal morphology.

Some attempts to classify coastal landforms (including shores and shoreline features) have been genetic, based on the origin of the landforms, rather than descriptive (e.g. cliffed coasts, delta coasts, mangrove coasts). The difficulty is that genetic classifications can only be applied when the mode of origin of coastal landforms is known, and as only a small proportion of the world's coastline has been investigated in sufficient detail to determine evolution such classifications remain somewhat speculative. Certainly the assumption that particular types or associations of landforms can be used as indicators of particular modes of origin can be misleading, for some coastal landforms (e.g. barrier islands, beach ridges, cuspate forelands, shore platforms) may evolve in more than one way: a phenomenon termed multicausality (Schwartz 1971). Various kinds of coastal classification are now described, with references.

Atlantic and Pacific type coasts

Suess (1906) distinguished Atlantic coasts, which run across the general trend of geological structures, from Pacific coasts, which run parallel to structural trends. The former are characteristic of the Atlantic shores of Britain and Europe; the latter of the Pacific coasts of North and South America.

Cliffed coastlines that transgress geological structures are termed discordant, whereas those that follow the strike of a particular geological formation are termed concordant.

Classification and plate tectonics

Inman and Nordstrom (1971) devised a geophysical classification based on PLATE TECTONICS, recognizing that the Earth's crust is a pattern of plates separated by zones of spreading and zones of convergence, with plate margins moving at rates of up to $15\,\mathrm{cm\,yr^{-1}}$. They contrasted subduction coasts, where one plate is passing beneath another, with trailing-edge coasts on a diverging plate margin and marginal sea coasts on the lee side of island arcs, and described features characteristic of each of these. It was a broad-scale classification, dealing with first-order (continental) features (*c*.1,000 km long $\times$ 100 km wide $\times$ 10 km high).

Coasts of submergence and emergence

Gulliver (1899) distinguished coasts formed by submergence from coasts formed by emergence. This was developed into a genetic classification by Johnson (1919), who described coastlines (he used the American term shorelines) of submergence, coastlines of emergence, neutral coastlines (with forms due neither to submergence nor emergence, but to deposition, e.g. delta coastlines, alluvial plain coastlines, glacial outwash coastlines and volcanic coastlines) and compound coastlines (with an origin combining two or more of the preceding categories). Most coasts fall into the compound category, because they show evidence of both emergence, following high sea levels in interglacial phases of the Pleistocene, and submergence, due to the Late Quaternary (Flandrian) marine transgression.

Classification based on climate

Aufrère (1936) proposed a coastal classification based on climate, which distinguished coasts with a permanent ice cover (no marine processes), coasts with a seasonal ice cover (seasonal marine processes and abundant sediment from glacial sources), temperate humid coasts (as in Europe), tropical humid coasts (with abundant fluvial sediment in deltas and coastal plains), arid coasts (without rivers; marine sediments dominant) and semi-arid coasts (some river features; SABKHAs). The global distribution of coastal climates shows sector variations related to latitude and wind regime with coastwise transitions that are generally gradual, although there are rapid transitions from humid tropical to arid within comparatively short distances in Ecuador and Colombia, in west Africa and northern Madagascar.

Classification based on coastal processes

Variations in coastal processes effective around the world's coastline were discussed by Davies (1980), who defined and mapped swell and storm wave environments, coasts subject to trade winds, monsoons and tropical cyclones, the distribution of high, moderate and low wave energy coasts, tidal types (semi-diurnal, mixed and diurnal) and mean maximum tide ranges divided into microtidal (<2 m), mesotidal (2–4 m) and macrotidal (>4 m), to which may be added megatidal (>6 m).

Initial and subsequent coasts

A distinction can be made between initial forms, which existed when the present relative levels of land and sea were established and marine processes began work (on most coasts about 6,000 years ago) and sequential forms, those that have since developed as the result of marine action. Shepard (1976) devised a classification on this basis, making a distinction between primary coasts shaped largely by non-marine agencies and secondary coasts that owe their present form to marine action. It was essentially a genetic classification, with descriptive detail inserted to clarify the subdivisions, and it recognized that, because of the worldwide Late Quaternary marine transgression, the sea has not long been at its present level relative to the land, so that many coasts have been little modified by marine processes.

Shepard's aim was to devise a classification that would prove useful in diagnosing the origin and history of coastlines from a study of charts and air photographs, but it is dangerous to assume that the origin and history of a coast can be deduced from such evidence without field investigation. A straight coast may be produced by deposition, faulting, emergence of a featureless seafloor or submergence of a coastal plain; an indented coast by submergence of an undulating or dissected land margin, emergence of an irregular seafloor, differential marine erosion of hard and soft outcrops along the coast or transverse tectonic deformation (folding and faulting) of the land margin. It is doubtful whether configuration can be taken as a reliable indicator of coastal evolution.

Leontyev *et al.* (1975) also considered initial and sequential forms (using the cycle of youth, maturity and old age) in a classification based on coasts not changed by the sea, coasts formed by abrasion or accumulation, and a combination of the two.

Stable and mobile coasts

Cotton (1952) made a distinction between coasts of stable and mobile regions, stable regions being those that escaped the Quaternary tectonic movements that have affected mobile regions, especially around the Pacific rim, where they still continue. On the coasts of stable regions he separated those dominated by features produced by Late Quaternary marine submergence from those dominated by inherited (mainly Pleistocene) features preserved by earlier emergence. On the coasts of mobile regions he separated those where the effects of Late Quaternary marine submergence have not been counteracted by recent uplift of the land from those where recent uplift of the land has caused emergence.

Morphological classification

De Martonne (1909) used a morphological distinction between steep and flat coasts as a basis for classification, suggesting a number of subtypes, some descriptive (estuary coasts, skerry coasts), others genetic (fault coasts, glacially sculptured coasts). Ottmann (1965) followed a similar approach, recognizing three categories of cliffed coast (cliffs plunging to oceanic depths, cliffs with shore platforms and cliffs plunging to submerged platforms), partially submerged uncliffed coasts, and low depositional coasts behind gently shelving seafloors.

Zenkovich (1967) classified depositional coastal features into five categories: attached forms (including beaches and cuspate forelands), free forms (including spits), barriers, looped forms (including tombolos) and detached forms (including barrier islands).

Geology in coastal classification

Russell (1967) advocated classification of rocky coasts on the basis of geology and structure, noting the striking similarity of features developed on crystalline rocks, irrespective of their climatic and ecological environments: granites that outcrop on parts of the coasts of Scandinavia, southwest Australia, South Africa and Brazil all show similar domed surfaces related to large-scale spalling and conspicuous joint-control. Limestones (including chalk and coral), basalts and sandstones also show distinctive kinds of coastal landforms. Bedrock coasts are commoner in cold, arid and temperate regions than in the humid tropics, where there has been deep weathering and depositional aprons are extensive.

Advancing and receding coasts

A coastline may advance because of coastal emergence and/or progradation by deposition, or retreat because of coastal submergence and/or retrogradation by erosion. Valentin (1952) used this analysis as the basis for a system of coastal classification that could be shown on a world map. Coasts that had advanced were divided into those produced by emergence, by organic deposition (mangroves, coral) and by inorganic deposition (marine and fluvial), while coasts that had retreated were divided into those produced by submergence of glaciated landforms and fluvially eroded landforms and those shaped by marine erosion. Bloom (1965) elaborated Valentin's scheme by considering historical evolution where the response to emergence, submergence, erosion and deposition has varied through time. Thus on the Connecticut coast, where radiocarbon dates from buried peat horizons have yielded a chronology of relative changes of land and sea level in Holocene times, there is evidence that at some stages the sea gained on the land during submergence, even though deposition continued, while at other stages deposition was sufficiently rapid to prograde the land during continuing submergence: at present there is widespread erosion on the seaward margins of saltmarshes, possibly because of resumed submergence.

The advantage of such non-cyclic classifications is that they pose problems and stimulate further research instead of trying to fit observed features into presupposed evolutionary sequences.

Composite classifications

McGill (1958) produced a map of the world's coastline which showed the major landforms of the coastal fringe, 8–16 km wide. This was a composite classification in which major coastal landforms were classified in terms of lowland or upland hinterlands, with additional information on selected features (constructional or destructional) in the backshore, foreshore and offshore zones, categorized by the agent responsible: sea, wind, coral or vegetation.

Artificial coastlines

Little attention has been given in coastal classifications to the fact that long sectors of coastline have become artificial during recent decades, partly as the result of engineering works designed to combat erosion and partly as a consequence of embanking or infilling to extend coastal land. On developed coasts the proliferation and extension of anti-erosion works, notably sea walls and boulder ramparts, has resulted in large proportions of artificial coastline: 85 per cent in Belgium, 51 per cent in Japan, 38 per cent in England. Coastal land has been artificially extended on a large scale in Singapore, Hong Kong, Tokyo Bay in Japan, western Malaysia and the Netherlands. The category of artificial coastlines is increasing rapidly, and much more of the world's coastline will become artificial as attempts are made to halt submergence and erosion.

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SEE ALSO: coastal geomorphology; global geomorphology

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COASTAL GEOMORPHOLOGY

The industrial, recreational, agricultural and transportational activities of growing human populations are exerting enormous pressures on coastal resources. To manage these activities in the least detrimental way, we need to have a better understanding of the dynamic nature of coastal landforms and the operation and interaction of marine and terrestrial processes. Differences in climate, changes in relative SEA LEVEL, wave environments, tides, winds, the morphology, structure and lithology of the hinterland, terrestrial and marine sediment sources, human activity and numerous other factors provide almost infinite variety to coastal scenery around the world. Coastal regions consist of a mosaic of diverse elements, some of which are contemporary, whereas others are ancient vestiges of periods when climate and sea level may have been similar or different from today's. Small-scale elements of depositional coasts, which can experience rapid changes in morphology, may attain a rough state of balance with their environmental conditions, but other features – particularly on hard rock coasts – require long periods to adjust to changing conditions. Furthermore, even if environmental conditions remain constant, individual coastal landforms still have to adjust to slow changes in the morphology of the coast itself. For example, whereas the profiles of sandy BEACHes respond fairly quickly to changing wave conditions, they may also have to adjust slowly to long-term changes in coastal configuration, sediment budgets, offshore gradients, climate, sea level and increasingly the effects of human interference.

Coastal classification

There have been many attempts to classify coasts, although none are entirely satisfactory. Most COASTAL CLASSIFICATIONS use at least two of three basic variables: the shape of the coast; changes in relative sea level; and the effect of marine processes. Some classifications are genetic, others are descriptive and others combine the two approaches. Genetic classifications are hindered by a lack of relevant data, however, and descriptive classifications, which have to accommodate an enormous variety of coastal types, tend to be cumbersome. Two classifications, which consider the nature of coastal environments and the effect of PLATE TECTONICS on coastal development, are particularly useful.

Davies (1972) proposed that coastal processes are strongly influenced by morphogenic factors that vary in a fairly systematic way around the world. Davies's morphogenic classification was based upon four major wave climates, although differences in coastal characteristics also reflect variations in tidal range, climate and many other factors. The highest WAVES are usually generated in the storm belts of temperate latitudes. Beaches in storm wave environments tend to have dissipative or gently sloping and barred profiles, and the major constructional features are often composed of coarse clastic material. Constructional features are oriented more by local fetch than by the variable direction of the deep water waves, and mechanical wave erosion is important in the formation of cliffs (see CLIFF, COASTAL) and SHORE PLATFORMS. Long, low constructional waves dominate swell environments between the northern and southern storm wave belts. The beaches have berms, and they tend to be towards the steeper, reflective, non-barred end of the spectrum. The direction of longshore currents is more constant than in storm wave environments, and large, sandy constructional features are oriented toward the approaching swell. Mechanical wave erosion of cliffs and platforms is probably slower than in storm wave environments, and this, combined with warmer climates, makes CHEMICAL WEATHERING and biological WEATHERING more important in swell wave environments. Sheltered, enclosed seas and ice-infested waters are low energy environments. Waves are flat and constructional, and beaches have prominent berms. The orientation of sandy constructional features, which are common in partially enclosed seas, is largely determined by local fetch.

Plate tectonics provide a partial explanation for the distribution of a variety of coastal elements, although the degree of explanation decreases with the decreasing size of the feature. Inman and Nordstrom (1971) proposed that the morphology of the largest, or first-order, coastal elements can be attributed to their position on moving tectonic plates. Three main geotectonic classes were identified: continental and ISLAND ARC collision coasts form along the edges of converging plates; plate-imbedded or trailing edge coasts face spreading centres; and marginal sea coasts develop where island arcs separate and protect continental coasts from the open ocean. The structural grain of collision coasts is parallel to the shore and they are therefore fairly straight and regular. Tectonically mobile collision coasts have narrow continental shelves and high, steep hinterlands, often with flights of raised terraces. The high relief provides an abundant supply of sediment to the coast. Plate-imbedded or trailing edge coasts usually have hilly, plateau, or low hinterlands, and wide continental shelves. The structural grain may be at high angles to the coast, which can therefore be very indented. Marginal seacoasts range from low-lying to hilly, with wide to narrow shelves, and they are often modified by large rivers and RIVER DELTAS.

Coastal modelling

Models provide one of the best ways of investigating the poorly understood components of a coastal system. They provide insights into the interrelationships between and among variables, and they are indispensable in enhancing our efforts to monitor, manage, control and develop the coastal system and its associated resources.

Physical models are simplified and scaled representations of the real world. They can be used to control and isolate variables, to provide insights into phenomena not yet described or understood, to provide measurements to test theoretical results and to measure complicated phenomena that cannot be theoretically analysed. Coastal engineers have constructed a wide variety of fixed-bed hydraulic scale models to study the action of waves, tides and currents, and to assist in the design of coastal structures. Geologists and geomorphologists have used movable bed models to examine sediment transport and the dynamics and formation of bars (see BAR, COASTAL), barriers (see BARRIER AND BARRIER ISLAND) and beaches. Unlike natural oceanic waves, however, the shallow water waves generated in most wave tanks have no orbital kinetic energy and are nearly pure solitons. Physical models therefore have not been able to describe accurately the hydrodynamics and sedimentary processes operating in coastal systems, and the results obtained from them always have to be verified or corroborated with other evidence.

Because of their generality, versatility and flexibility, mathematical models are the most common type used by coastal workers. Unfortunately, however, our lack of knowledge of coastal processes and the frequent reliance on laboratory data to determine the value of coefficients, casts doubt on the applicability of many mathematical models to the real world. There are several types of mathematical model. Deterministic models, which are based on the principles of fluid mechanics, seem to work best in conjunction with laboratory experiments that allow parameters to be held constant while one is varied at a time. Simulation models involve the manipulation of process–response equations on computers, compressing years of coastal development in the prototype into minutes. This allows the behaviour of a system to be determined under a variety of situations and conditions, and to test the sensitivity of the system to changing input parameters. Statistical models can be used to study the relationships between a set of variables, and to verify possible relationships identified by theoretical models. To use equations derived from one area for predictive purposes in another, however, often requires the determination of a different set of coefficients.

Coastal inheritance

There is growing evidence that because interglacial sea levels were similar to today, contemporary coastal features often formed close to, or were superimposed on top of, their ancient counterparts. Although evidence of past sea levels and climates is generally easily obliterated in unconsolidated coastal deposits, many sandy coasts retain sedimentary and morphological elements of former environmental conditions. Coastal deposits from the last interglacial stage are being cannibalized in some areas to provide sediment for the construction and maintenance of modern coastal features, and barrier systems have sometimes developed on top of older Pleistocene barriers, or are located somewhat seaward of them. Most barrier islands on the German North Sea coast and in places on the Atlantic coast of the USA, for example, consist of a core of Pleistocene deposits, mantled by Holocene sediments. In south-eastern Australia, a distinct inner barrier of the last interglacial age is separated from an outer Holocene barrier by a lagoon and swamp tract. Pleistocene dunefields are adjacent, and probably under Holocene coastal dunes (see DUNE, COASTAL) in some places, especially in Australia and the Mediterranean, although they are generally absent in northern Europe, where most dunes were built at different stages during the Holocene. The presence of near-surface discontinuities shows that Holocene limestones, ranging from a few metres up to about 30 m in thickness, also form veneers over foundations of older reef-rock. The concept of INHERITANCE is particularly important on resistant rock coasts which have probably evolved very slowly during successive periods of high interglacial sea level. It has been demonstrated that some cliffs, sea caves, ramps (see RAMP, COASTAL) and shore platforms are at least last interglacial in age, and modelling suggests that many platforms have developed during interglacial stages during the middle and late Pleistocene.

Coastal management

Despite the problems associated with flooding, erosion, pollution and other hazards, and the increasing aesthetic and practical impetus for sustainable coastal management, rising populations and growing economic pressures are accelerating the pace of human interference and degradation on the world's coasts (Plate 25). We lack reliable models, however, that can be usefully employed by managers, planners and decision-makers for INTEGRATED COASTAL MANAGEMENT and to predict the effects of sea-level changes, human activities and other factors on the coast. The available field data on coastal changes are often of questionable reliability and usually too short-term to analyse the interaction of a large number of variables. Coastal changes are also frequently complex and non-linear (see NON-LINEAR DYNAMICS), and may reflect the interaction and exchange of sediment between the coast and the CONTINENTAL SHELF, and between the coast and the land, a relationship that is increasingly influenced by anthropological activities.

Long stretches of coastlines are now essentially artificial, with GROYNEs, breakwalls and other engineering structures (Plate 26). These structures

Plate 25 Crowded beach on the Costa del Sol, southern Spain

Plate 26 Groynes on Gold Beach, Normandy, France

are aesthetically unpleasant and they interfere with sediment transport and other natural processes, although this can be partly mitigated by artificial BEACH NOURISHMENT. Human removal of beach material continues in some areas today, although legislation has been enacted to discourage it in many areas. The importance of dunes as a natural coastal defence for low-lying land is reflected in laws relating to dune stabilization dating back to the thirteenth century. Humans affect coastal dunes in many direct and indirect ways, including sand extraction, forestation and deforestation, trampling and off-road vehicles, introduction of exotic species and grazing and burrowing animals, and changes in the water table resulting from forestation or residential and industrial development. Dune stabilization and construction has been undertaken in many countries, although it can reduce morphological variety and species diversity. The protection of dunes also impairs their ability to replenish beaches during storms. It has been suggested that construction of a high protective barrier dune on the northern barrier islands of North Carolina threatens their existence, because it prevents OVERWASHING, the opening of inlets and natural barrier recession. Others, however, consider that the artificial dune reduces erosion by nourishing the beach during storms. In many areas, as in dunefields, SALTMARSHes and MANGROVE SWAMPs, one must understand the workings of coastal ecological as well as geomorphological systems to solve coastal problems (Viles and Spencer 1994). Large saltmarsh areas have been reclaimed for agriculture, housing, industry and airports, although there is increasing interest in their preservation with the recognition that they are important and productive ecosystems. Human activities, including deforestation for rice paddies, fuel, construction materials and industrial uses, are continuing to cause irreversible damage to coastal mangroves in tropical regions, however, where there is often little appreciation of their value to native populations. Estuarine dynamics and siltation patterns are being affected by deforestation, mining and quarrying, urbanization, DAM construction, sewage discharge, dredging, dock and marina construction, the reclamation of TIDAL DELTAs, flats and marshes and the diversion of water from one watershed into another. Although much human activity is deleterious to deltas, deforestation, agricultural intensification and extensive soil erosion have sometimes been responsible for their formation or growth. Many deltas are receiving less water and sediment as rivers are dammed for irrigation, flood control and power generation. Much of the loss of wetlands in the Mississippi Delta has natural causes, but it is being exacerbated by deforestation of the drainage basin and dam construction, the building of LEVEES and other attempts to confine and control the Mississippi River for navigation and flood control. Human activity has been modifying the Nile Delta since predynastic time, but, with construction of the Aswan Dams, almost no fluvial sediment now reaches the delta, and this has resulted in accelerated coastal erosion and marine encroachment. Coral communities and reefs are also threatened by a variety of human activities, including dredging, mining, land clearance, effluents from desalination, sewage discharge, the use of chlorine bleach and explosives for fishing, nuclear weapon testing, oil, chemical and sewerage pollution, thermal pollution from electrical generating stations, careless anchoring, boat grounding and the collection of precious corals and other marine organisms. The Great Barrier Reef Marine Park in Australia was created to manage reefs comprehensively, but economic pressures are more severe in developing areas, and conservation policies more difficult to enforce.

Global warming

One of the greatest challenges facing coastal populations will be to plan for, and manage, the effects of rising sea level resulting from global warming. There is continuing debate over the rate and magnitude of the changes that are to be expected, however, although there has been a trend towards progressively more conservative predictions of sea-level rise in this century. The 2001 third assessment report of a working group for the intergovernmental panel on climatic change (IPCC) has concluded that sea level will rise by between 0.09 and 0.88 m between 1990 and 2100.

Global warming and rising sea level will cause tidal flooding and the intrusion of salt water into rivers, estuaries (see ESTUARY) and groundwater, and it will affect tidal range, oceanic currents, upwelling patterns, salinity levels, biological processes, runoff and landmass erosion patterns. Increasing rates of erosion will make cliffs more susceptible to falls, landslides and other MASS MOVEMENTs, exacerbating problems where loose or weak materials are already experiencing rapid recession. Nevertheless, the effect of rising sea level will vary around the world according to the characteristics of the coast, including its slope, wave climate, tidal regime and susceptibility to erosion.

It has been estimated that about half the world's population lives in vulnerable coastal lowlands, subsiding RIVER DELTAs and river floodplains. The effects of climatic change will be particularly acute in these densely populated regions. It is often the rate of sea-level change rather than the absolute amount that determines whether natural systems, such as coastal marshes and CORAL REEFs, can successfully adapt to changing conditions. Human and natural systems can adjust to slowly changing mean climatic conditions, but it is more difficult to accommodate changes in the occurrence of extreme events. It is not yet known, however, whether higher sea temperatures will increase the frequency and intensity of tropical storms and spread their influence further polewards, or whether higher temperature gradients between land and sea will increase the intensity of monsoons and affect their timing.

Human responses to the rise in sea level will depend upon available resources and the value of the land being threatened. High waterfront values will justify economic expenditure to combat rising sea level in cities, but less attention is likely to be paid to the deleterious effects on saltmarshes, mangroves, coral reefs, lagoons and ice-infested Arctic coasts. The decision-making process associated with coastal erosion and flooding is complex, because of constraints imposed by financial considerations and a myriad of physical, social, economic, legal, political and aesthetic factors. There is public and political pressure on coastal planners and managers to be seen to be doing something about the problem, and this can result in engineering projects that provide only short-term benefits, or which may even exacerbate the original problem. Several managerial options are available, however, ranging from the 'do nothing' approach, to the construction of a completely artificial coast.

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ALAN TRENHAILE

COHESION

The force by which particles are able to stick together. Cohesion is important in soil mechanics, as it is one of two parameters (alongside the angle of internal friction) that characterize a soil's resistance to an applied stress (though the two parameters are not always independent of each other). Soils with high levels of cohesion (termed cohesive soils) commonly contain a significant amount of clay, which are able to cement the soil internally (yet these typically have low frictional strength). Conversely, dry sand is termed non-cohesive (as particles are easily moved in isolation), with the only resistance to shear coming from the internal friction of sand particles. When sand is moist (though unsaturated) the surface tension of the water menisci between the grains provides an apparent cohesiveness to the sand. This is removed when the sand either dries or becomes saturated. Rocks are commonly high in both parameters. Cohesion becomes proportionately stronger as grain size decreases, allowing fine grain sediments (muds and silts, etc.) to remain stable on high-angle slopes.

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SEE ALSO: adhesion

STEVE WARD

COLLUVIUM

Sedimentary material that has been transported across and deposited on slopes as a result of mass movement processes and soil wash. It is frequently derived from the erosion of weathered bedrock (eluvium) and its deposition on low-angle surfaces, and can be differentiated from material which is deposited primarily by fluvial agency (alluvium). Colluvium can be many metres thick and can infill bedrock depressions (Crozier *et al*. 1990). It often contains palaeosols, which represent halts in deposition, crude bedding downslope, and a large range of grain sizes and fabrics (Bertram *et al*. 1997). Cut-and-fill structures may represent phases when stream incision has been more important than colluvial deposition (Price-Williams *et al*. 1982).

Colluvium may provide a rich record of long-term climatic change (see, for example, Nemec and Kazanci 1999), preserve archaeological materials, indicate phases of accelerated anthropogenic soil erosion during the Holocene and act as a medium into which gullies may be incised (see DONGA).

Colluvial deposits are known from almost all climatic zones from former glacial (Blikra and Nemec 1998) and periglacial environments (Mason and Knox 1997) through to the tropics (Thomas 1994).

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A.S. GOUDIE

COMMINUTION

Refers to the reduction of rock debris to fine powder or to small pieces. In nature, comminution is usually as a result of ABRASION and attrition, and is often linked with problems of coastal erosion due to reduction of shingle.

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STEVE WARD

COMPACTION OF SOIL

The term compaction refers to a progressive decrease in the volume of a soil element over time, resulting in an increase in density. Recently deposited sediments tend to exhibit a progressive increase in density over time, as consolidation occurs due to self weight and loads imposed by overlying sediment. A commonly used measure of the relative degree of compaction of a soil within engineering soil mechanics is the overconsolidation ratio: $OCR = \sigma'_{max} / \sigma'_{pres}$. Here σ'_{max} refers to the maximum normal EFFECTIVE STRESS which the soil material has experienced over geologic time, while σ'_{pres} is the present-day normal effective stress. Effective stress is defined as total stress minus ambient PORE-WATER PRESSURE (Barnes 2000). Normally consolidated (NC) soils have $\sigma'_{pres} \approx \sigma'_{max}$, and include most postglacial fluvial and colluvial sediments. Overconsolidated (OC) soils have $\sigma'_{max} \gg \sigma'_{pres}$, and include basal tills and geological strata such as clays and shales which have experienced normal stress reduction caused by erosion of superjacent materials. There are large ranges of OCR from approximately 1.0 to several 100, depending on the history of load changes that the soil has experienced. A transient condition, known as underconsolidation, refers to effective stress below the NC value. This is possible where part or all of the total overburden pressure is borne by the pore fluid, and thus positive excess pore pressures prevail shortly after deposition. It is common where fine-grained, saturated materials are deposited rapidly as QUICKCLAY earthflows or muddy DEBRIS FLOWS. Underconsolidation may also occur where formerly submerged muds become abruptly subaerial, due to either rapid tectonic uplift or lake drainage.

Although the rate of consolidation is controlled strongly by the normal stresses imposed by external loads, soil compaction also varies according to the compressibility of the soil particles themselves, the water content, and the hydraulic conductivity (Barnes 2000). In unsaturated soils, having a high air content, rate of consolidation is controlled primarily by the compressibility of the soil matrix, which is a function of particle shape, sorting, and mineralogy. In saturated soils, rate of consolidation is regulated by soil hydraulic conductivity, since expulsion of virtually incompressible pore fluid is a prerequisite for consolidation. Conductivity varies by several orders of magnitude, depending on particle size and *in situ* density.

Within the normally consolidated class of soils, which comprise many soils worldwide, significant variations in ambient *in situ* density occur as a result of both geomorphic and sedimentological factors. Mixed, poorly sorted materials, such as LANDSLIDE deposits, often possess a relatively high *in situ* density since a wide range of particle sizes ensures that voids between large clasts are filled with finer material (Bement and Selby 1997). It is possible that natural, vibration-induced compaction of rapidly emplaced landslide materials further enhances densification. By contrast, very well-sorted aeolian materials, such as LOESS and DUNE sand, exhibit a much lower *in situ* density, especially if fairly equant grains are dominant in the deposit. Such soils are inherently very compressible.

In the near-surface zone, the effects of geological consolidation are periodically offset by MECHANICAL WEATHERING processes, which lead to a volume increase, and hence a density decrease, relative to that of the unweathered material below. By contrast, the amount of net volume increase brought about by CHEMICAL WEATHERING appears to be slight (Birkeland 1984). In cold regions, FREEZE–THAW CYCLE processes cause seasonal and shorter term cycles of heave and settlement. Thaw and consolidation of the ACTIVE LAYER during spring and summer may produce transient excess pore pressures if the water generated by ice lens melting is slow to escape. This may be due to either a low material conductivity or the existence of an impermeable PERMAFROST table (Williams and Smith 1989). Thaw-consolidation has been credited with the development of very low effective stresses within a thawing active layer, allowing SOLIFLUCTION lobes to move on slope angles as low as $\frac{1}{4}\,\phi'_r$, where ϕ'_r is the residual angle of shearing resistance. Cycles of HYDRATION and dehydration also produce appreciable cyclical volume changes, especially in soils containing montmorillonite clays. However, the magnitudes of the resultant cyclical volume changes are generally far lower than the values attained within seasonally ice-rich sediments.

Rainfall impact, together with infiltration seepage, is also a well-documented soil compacting process, especially in semi-arid environments where it leads to the development of a surface crust of reduced infiltrability. The widespread conversion of grassland and forest soils to arable use has caused significant rainfall compaction of soil, causing reduced infiltrability, and hence accelerated runoff (see RUNOFF GENERATION) and EROSION (Morgan *et al.* 1998). In arable areas, such compaction may be rectified by ploughing and harrowing. In time, uncultivated near-surface soil becomes naturally loosened again by the combined effects of freeze–thaw cycles, bioturbation from soil micro- and macrofauna, in addition to root growth and decay, and downward mixing of low density organic material.

Several problem soils have been identified within engineering soil mechanics based on their poor performance under surcharge stresses or cyclical shear loads. Normally consolidated clays are prone to significant consolidation under structural loads, and may require the placement of fill materials to effect soil consolidation prior to construction (Barnes 2000). NC soils are also more prone to landsliding than are OC materials, since the lesser degree of compaction in the former is generally associated with lower shear strength. A common problem in loess soils is HYDROCOMPACTION (Derbyshire 2001), which involves a localized collapse of soil structure in response to vertical seepage forces. It is a widespread problem where loess is subjected to flood irrigation.

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MICHAEL J. BOVIS

COMPLEX RESPONSE

Landforms respond to the controlling variables of tectonics, sea level, climate and biotic activity over time. They also respond to the changes, rhythms and thresholds of Earth. Available data suggest that over timescales of 10^2 years increases of geomorphological rates of activity change with a frequency of *c.*2,000 years. Over 10^3 years rates and process balances change with a frequency of 30,000–50,000 years and over 10^{4-5} years full system control changes occur every 100,000–150,000 years. Flux in sediment yield and landform adjustment should be regarded as the norm. Regularity of landform may then be the product of polygenetic landform origins. A central proposition of geomorphology, therefore, is that landform change (response) takes place as states of equilibrium, stability or tranquillity are upset by complex episodic changes to the environmental controls. This may be called 'complex cause' (see LANDSCAPE SENSITIVITY).

The *response* to the hierarchy of controls and events also varies on all timescales and are variably distributed in space. Complex response (Schumm 1973, 1975, 1977, 1979, 1981; Schumm and Parker 1973) describes the way in which the internal structure of a system controls the reaction and relaxation of the system after an impulse of change. There are many aspects to be considered: the effect of internal thresholds (see THRESHOLD, GEOMORPHIC) that control sudden change; the fluctuation between cut-and-fill as the capacity of the system dictates temporary storage of eroded sediment; the effect of area as an impulse moves from a point application (e.g. a river mouth base level change), along a sensitive linear pathway (e.g. a channel, a joint) to diffuse over a catchment as a wave of erosional aggression moving inland (e.g. from a sea cliff or an incising river). Such changes occur after every effective event and the direction of change follows every structural instability.

Landform 'evolution' is a never-ending set of adjustments to impulses of change on all temporal and spatial scales. It is complex.

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DENYS BRUNSDEN

COMPLEXITY IN GEOMORPHOLOGY

Complexity is a way of describing complicated, irregular patterns that appear random. It is something tangible that is observable in geomorphic systems, such as in turbulent flow in streams. Much chaotic complexity in geomorphology underlies a larger scale geomorphic order, and overlies smaller scale, more orderly and understandable components. Chaotic turbulent flow is part of a larger scale order seen in the predictable rate and direction of mean streamflow; it is also the result of a huge number of well-understood individual particle trajectories describable by the basic laws of physics. Complexity in geomorphic systems is thus often part of a hierarchy of interrelated structures and processes. Similarly, simple geomorphic patterns, such as beach cusps, commonly arise from complex underlying dynamics; at the same time, they are but a part of broader scale complex patterns. Beach cusps result from complex non-linear interactions between beaches and waves or the complicated formation of edge waves (waves trapped at the shoreline by refraction); at the same time, they are a part of irregular coastline geometry.

One line of explanation for complexity rests in non-linear dynamical systems theory, which has revolutionized many branches of science (see Stewart 1997). To understand the general reasoning involved, it may help to define a few terms first. An unstable system is susceptible of small perturbations and is potentially chaotic. A chaotic system behaves in a complex and pseudo-random manner purely because of the way the system components are interrelated, and not because of forcing by external disturbances, or at least independently of those external factors. The equations describing the system generate the chaos, which is deterministic; chance-like (stochastic) events do not. Systems displaying chaotic behaviour through time usually display spatial chaos, too. Therefore, a landscape that starts with a few small perturbations here and there, if subject to chaotic evolution, displays increasing spatial variability as the perturbations grow. This happens when rivers dissect a landscape and relief increases. Self-organization is the tendency of, for example, flat or irregular beds of sand on streambeds or in deserts to organize themselves into regular spaced forms – ripples and dunes – that are rather similar in size and shape. Self-organization also occurs in patterned ground, beach cusps and river channel networks. Self-destruction (non-self-organization) is the tendency of some systems to consume themselves, as when relief is reduced to a plain. An attractor is a system state that controls system changes and into which other system states are drawn.

Many geomorphic systems are complex, but not all are. Some non-linear geomorphic systems are unstable, chaotic and self-organizing, but some are not. Nevertheless, plentiful evidence suggests that complexity is common in geomorphic systems and begs an explanation. The truly puzzling fact is that most geomorphic systems display order and complexity concurrently. Are the complexities (irregularities) merely deviations from an orderly norm, or are they informative in their own right? A growing body of evidence from field studies, laboratory studies, and real-world datasets suggests that in some geomorphic systems complexity is significant in its own right. Signs of complex behaviour in systems include deterministic chaos, instability, increasing variability over time, self-organization, divergence from similar initial conditions and sensitivity to initial conditions (Phillips 1999: 39–57). Evidence exists for all these indicators of complexity.

Several hydrological records, tree rings series and topographic images reveal chaotic patterns. In other cases, field investigations have confirmed chaotic behaviour predicted in models, as in the genesis of Ultisols in eastern North Carolina.

Field examples of dynamical systems' instability and sensitivity to small perturbations abound, including river meander initiation generated by the unstable growth of small flow perturbations.

Some studies demonstrate patterns of spatial variability that become increasingly complex (less uniform) over time: there is a spatial differentiation of the landscape. Desertification appears to involve an increasingly more complex pattern of vegetation and soil-nutrient resources through time.

In some geomorphic systems, orderly self-organizing patterns seem to emerge from complex non-linear dynamics. Field and laboratory work confirms theoretical work showing that sorted nets in non-periglacial environments may develop spontaneously on any piece of unobstructed land with little or no slope, proving it carries a loose and discontinuous cover of pebbles, each of which may move in small steps with equal probability in all directions (Ahnert 1994).

Much field evidence strongly suggests that some geomorphic systems evolve by diverging from the same, or very similar, initial conditions. In the Norfolk marshes, England, vegetated marsh traps more sediment than bare marsh, so reducing the chances of inundation and lowering (or stabilizing) salinity. The bare marsh becomes lower land that traps more water and the salinity rises, inhibiting vegetation colonization and growth.

Several studies indicate that small variations in initial conditions amplify as a geomorphic system evolves. In podzolized soils in Canada, microtopographic variations produce favoured sites for infiltration and 'funnel' effects that eventually create large variations in the thickness of A and B soil horizons (Price 1994).

Related to complexity are the ideas of fractals and self-organized criticality. Fractal landscapes display self-similar patterns repeated across a range of scales. A small section of coastline may be self-similar to a much larger piece of coastline, of which it is part. Drainage networks, sedimentary layers and joint systems in rocks possess fractal patterns. Self-organized criticality is a theory that systems composed of myriad elements will evolve to a critical state, and that once in this state, tiny perturbations may lead to chain reactions that may affect the entire system. The classic example is a pile of sand. Adding grains one by one to a sandbox causes a pile to start growing, the sides of which become increasingly steep. In time, the slope angle becomes critical: one more grain added to the pile triggers an avalanche that fills up empty areas in the sandbox. After adding sufficient grains, the sandbox overflows. When, on average, the number of sand grains entering the pile equals the number of grains leaving the pile, the sand pile has self-organized into a critical state. Landslides, drainage networks, and the magnitude and frequency relations of earthquakes display self-organized criticality.

Phillips (1999) identifies eleven 'principles of Earth surface systems' that follow from theoretical and empirical work on order and complexity in geomorphic systems. Some of these principles appear to conflict, but that is the nature of complexity. In summary, and applied specifically to geomorphic systems, the principles are (see Huggett 2002: 339–41):

1 Geomorphic systems are inherently unstable, chaotic and self-organizing. Many, but definitely not all, geomorphic systems display a tendency to diverge or to become more differentiated through time in some places and at some times, as when an initially uniform mass of weathered rock or sediment develops distinct horizons.
2 Geomorphic systems are inherently orderly. Deterministic chaos in a geomorphic system is governed by an >attractor=that constrains the possible states of the system. Such a geomorphic system displays dynamic instability but does not behave randomly. The dynamic instability has bounds. Beyond these bounds, orderly patterns emerge that include the chaotic patterns inside them. Thus, even a chaotic system must exhibit order at certain scales or under certain circumstances. For example, at local scales, soil formation is sometimes chaotic, with giant spatial variations in soil properties; as the scale is increased, regular soil–landscape relationships emerge.
3 Order and complexity are emergent properties of geomorphic systems. This principle means that, as the spatial or temporal scale is altered, orderly, regular, stable, and non-chaotic patterns and behaviours and irregular, unstable, and chaotic patterns and behaviours appear and disappear. In debris flows, deterministic chaos governs collisions between particles where the flow is highly sheared and the collisions are sensitive to initial conditions and unpredictable. However, the bulk behaviour of granular flows is orderly and predictable from a relationship between kinetic energy (drop height) and travel length. Therefore, the behaviour of a couple of particles is perfectly predictable from basic physical principles; a collection of particles interacting with each other is chaotic; and the aggregate behaviour of the flow at a still broader scale is again predictable.
4 Geomorphic systems have both self-organizing and non-self-organizing modes. This principle

follows from the first three principles. Some geomorphic systems may operate in self-organizing and non-self-organizing modes at the same time. The evolution of topography, for example, may be self-organizing where relief increases, and self-destructing where relief decreases. Mass wasting denudation is a self-destructing process that homogenizes landscapes by decreasing relief and causing elevations to converge. Dissection is a self-organizing process that increases relief and causes elevations to diverge.

5 Both unstable–chaotic and stable–non-chaotic features may coexist in the same landscape at the same time. Because a geomorphic system may operate in either mode, different locations in the system may display different modes simultaneously. This is the idea of >complex response =, in which different parts of a system respond differently at a given time to the same stimulus. An example is channel incision in headwater tributaries occurring concurrently with valley aggradation in trunk streams.

6 Simultaneous order and disorder, observed in real landscapes, may be explained by a view of Earth surface systems as complex non-linear dynamical systems. They may also arise from stochastic forcings and environmental processes.

7 The tendency of small perturbations to persist and grow over finite times and spaces is an inevitable outcome of geomorphic system dynamics. In other words, small changes are sometimes self-reinforcing and lead to big changes. Examples are the growth of nivation hollows and dolines. An understanding of non-linear dynamics helps to determine the circumstances under which some small changes grow and others do not.

8 Geomorphic systems do not necessarily evolve towards increasing complexity. This principle arises from the previous principles and particularly from Principle 4. Geomorphic systems may become more complex or simpler at any given scale, and may do either at a given time.

9 Neither stable, self-destructing nor unstable, self-organizing evolutionary pathways can continue indefinitely in geomorphic systems. No geomorphic system changes ad infinitum. Stable development implies convergence that eventually leads to a lack of differentiation in space or time, as when different elevations in a landscape converge to form a plain. Disturbances disrupt such stable states by reconfiguring the system and resetting the geomorphic clock. Divergent evolution is also self-limiting. For example, base levels ultimately limit landscape dissection.

10 Environmental processes and controls operating at distinctly different spatial and temporal scales are independent. For example, processes of wind transport are effectively independent of tectonic processes, although there are surely remote links between them.

11 Scale independence is a function of the relative rates, frequencies and durations of geomorphic phenomena.

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RICHARD HUGGETT

COMPUTATIONAL FLUID DYNAMICS (CFD)

Fluid motions play a central role in sculpting a great variety of landforms, both terrestrial and submarine. Examples range from river channels to aeolian dunes to barrier islands. Naturally, investigations into landform origins often involve applications of fluid dynamics. Fluid dynamics is that branch of mechanics that concerns the physics of fluid motion. The motion of a non-turbulent, Newtonian fluid is described by the Navier–Stokes equations. These equations express continuity of mass and momentum in three dimensions in a continuum fluid subject to gravitational, inertial, viscous and pressure forces. The equations take on different forms depending on whether the fluid is compressible (e.g. air) or incompressible (e.g. water to a close approximation). Normally the Navier–Stokes equations are combined with models of turbulence (for application to turbulent flows) and with models of boundary friction (for any flows involving contact with a surface, such as a channel bed). Except in special cases, the Navier–Stokes equations cannot be analytically solved. Their solution can, however, be approximated using numerical methods (see, e.g. Cheney and Kincaid 1999; Press *et al.* 1993) combined with a set of specified initial and boundary conditions. Such methods involve dividing up space and time into discrete elements, within which the variables of interest – such as velocity and pressure – are either interpolated or held constant. The development of numerical solution methods for different types of equations, including fluid flow equations, is a major area of research in the fields of mathematics and computing science. Numerical solutions of equations for fluid motion can be quite computationally intensive, and the computer models that implement these solutions are referred to as Computational Fluid Dynamics (CFD) models. Depending on the methods used and the degree of approximation involved, the computer codes can be quite complex, and there are many commercially available packages as well as research codes developed within universities. Applications of CFD are increasingly widespread in geomorphology. CFD models have been of great benefit, for example in understanding interactions between fluid flow, bed morphology and sediment transport in river channels (e.g. Hankin *et al.* 2002; Lane *et al.* 2002; Ma *et al.* 2002). CFD models of airflow dynamics have been used to understand the interactions between airflow and dune morphology (e.g. Walmsley-John and Howard 1985). Other applications have been wide-ranging; examples include water flow in karst conduits (e.g. Hauns *et al.* 2001), circulation and sediment movement in ancient epeiric seas (e.g. Slingerland *et al.* 1996), coastal morphology (e.g. Deigaard and Fredsoe 2001) and paleoflood hydrology (e.g. House and Baker 2001).

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GREG TUCKER

CONCHOIDAL FRACTURE

A smoothly curved fracture, marked by concentric rings and resembling a bi-valve shell in shape. Conchoidal fractures are the most common type of fracture, and are also known as clamshell fractures. They occur when bonds between atoms are approximately the same in all directions within a mineral, and result in breakage along smooth, curved surfaces. Conchoidal fractures occur particularly in amorphous materials (i.e. those showing no definite crystalline structure) such as obsidian, and are also common in quartz, chert and glass.

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STEVE WARD

CONFLUENCE, CHANNEL AND RIVER JUNCTION

River channel confluences, the sites at which two open channels combine, are ubiquitous features of all river networks and channel patterns. These sites mark nodes of significant change in hydraulic geometry (Richards 1980), flow and sediment discharge, and are characterized by a complex three-dimensional flow field and variable bed geometry (Mosley 1976; Best 1988; Bradbrook *et al*. 2000; Rhoads and Sukhodolov 2001). River channel junctions are often points of significant bed scour (e.g. Best and Ashworth 1997), and are critical in considerations of sediment/pollutant dispersal and mixing in channel networks (Plate 27). Study of these complex fluvial sites has progressed through field, physical and numerical modelling and has identified five principal controls on flow, sediment transport and bed morphology at channel confluences: (1) the angle of convergence between the confluent channels; (2) the ratio of discharge, or flow momentum, between the incoming channels; (3) the planform shapes of the junction (for instance, ‘Y’ or ‘⊥’ shaped) and upstream channels (i.e. curved, straight; single, multiple); (4) the presence of any depth differential between the incoming channels; and (5) the relative roughness of the confluence (ratio of flow depth to grain size), with the hydrodynamic influence of the particles beginning to dominate at larger grain sizes (e.g. Roy *et al*. 1988).

Plate 27 Junction of the Paraná and Paraguay Rivers, Argentina. The Paraguay River enters from the right and is picked out by its higher suspended sediment concentration. The shear layer between the two rivers displays a series of large vortices and the mixing layer remains distinct for many tens of kilometres downstream. Width of Paraguay River inflow at confluence ~1 km

Fluid dynamics

Channel confluences are zones of complex, three-dimensional, turbulent flow where significant local flow acceleration and deceleration may occur due to both the increased combined fluid discharge and the specific fluid dynamics of the confluence region. Experimental, field and numerical studies have shown confluences to be dominated by seven fluid dynamic zones (Figure 26).

1 A zone of flow stagnation near the upstream junction corner; this fluid deceleration is caused by turning and hence centrifugal forcing of the flows as they approach the junction, together with the influence of a pressure gradient within the junction that is generated by water surface superelevation in the junction centre.

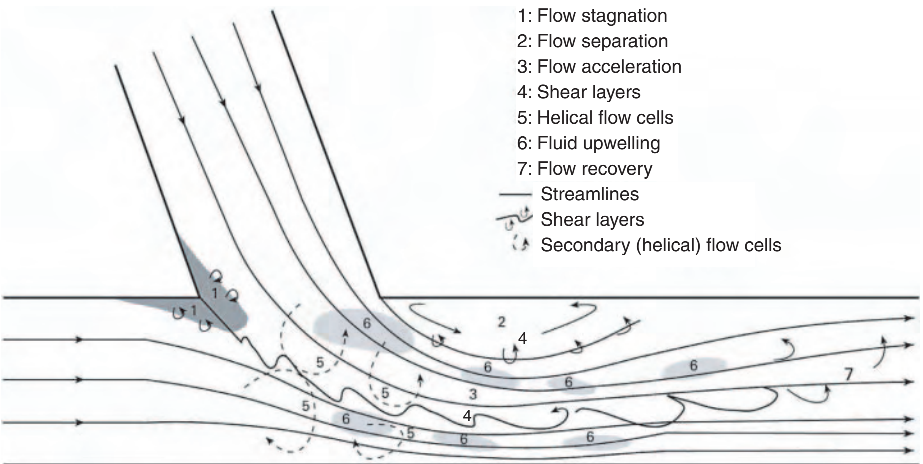

Figure 26 Schematic diagram of the seven principal fluid dynamic zones that may be present at channel confluences

2 A region of flow separation may occur downstream from the downstream junction corner(s); flow cannot remain attached to the boundary at sudden changes in geometry, and an adverse hydrostatic pressure gradient here causes the flow to separate from the wall and form a region of slow, recirculating flow. In symmetrical confluences, the downstream separation zone may form on both sides of the junction. In an asymmetric confluence, the downstream separation zone may only form on the angled (i.e. tributary) side of the junction. The size of the downstream separation zone(s) increase(s) with junction angle and tributary discharge (Best 1988; Bradbrook *et al.* 2000), but may be modified/absent at natural junctions where the angle of bank divergence at the downstream junction corner(s) may be modified by formation of a point bar through sediment deposition (e.g. Rhoads and Sukhodolov 2001) and/or bank erosion.

3 A region of flow acceleration forms at the centre of the confluence that is generated by both the increased fluid discharge passing through the junction (see streamline convergence, Figure 26) and also the constricting influence of any region of flow separation.

4 Distinct shear layers are generated along regions where velocity gradients are severe. Thus, shear layers can be present on either side of the flow stagnation region, along the mixing interface between the two joining flows, bounding any regions of flow separation and also arising from any steep changes in bed topography (such as the avalanche faces that may dip into the central scour – see below). Large, turbulent and 3D flow structures arising along these shear layers, termed Kelvin–Helmholtz instabilities, may give rise to high turbulent shear stresses that are influential in fluid mixing and sediment transport.

5 Helical flow may develop within the junction due to the presence of streamline curvature (Figure 26; streamlines are lines drawn in the fluid of which the tangent at any point is the direction of velocity at that point) and water surface superelevation within the confluence. In ideal cases, where the tributaries are near symmetrical and have equal flow momentum, these secondary flows may be expressed as surface convergent, bed divergent flows much as in placing two meanders back to back, although the duration of streamline curvature through the bend means that it is unlikely that an entire helix is ever completed. The presence of both flow separation at the junction corner, changing pressure gradients or flow separation associated with bed

topography ('topographic forcing' of the flow) or a depth differential between the two incoming tributaries, may lessen the effects or destroy such large-scale secondary flows. Additionally, the time-averaged picture of a series of individual turbulent events, such as fluid upwelling in the confluence, may be manifested as apparent secondary circulation (Lane *et al*. 2000).

6 Regions of distinct fluid upwelling may be generated by both distortion of the shear layer and flow associated with bed topography, such as where the beds of the tributaries are discordant in their height at the junction. This may encourage upwelling of one stream into the other, thus greatly increasing the rate of mixing at the junction (Gaudet and Roy 1995).

7 Finally, a region of flow recovery downstream of the confluence has been observed. This is where the effects of the junction lessen and flow returns to a more uniform cross-stream distribution. However, the flows may remain unmixed for many channel widths downstream if the velocity differential across the shear layer is minimal and the local turbulence at the junction does not mix the flows (see Plate 27).

Bed morphology

The topography of river channel confluences is often characterized by four distinct elements. First, a central scour hole is often present whose orientation approximately bisects the junction angle. The depth of scour increases at both higher junction angles and momentum ratios, and some of the largest alluvial scours are found at these sites. Scours at junctions may reach between two and ten times the depth of the upstream confluent channels: for instance, scour depth at the confluence between the Ganges and Jamuna (Brahmaputra) Rivers in Bangladesh has been recorded as up to 30 m below the upstream bed level in the confluent channels (Best and Ashworth 1997). The position and cause of the confluence scour have been related to (a) flow acceleration in the centre of the confluence (e.g. Roy *et al*. 1988); (b) the influence of turbulence along the shear layer between the flows; (c) downwelling created by secondary flows that may cause higher momentum fluid to be transferred towards the bed at the centre of the confluence; and (d) the differential routing of sediment around the scour.

Second, tributary mouth bars have been observed that terminate at the junction. These bars may possess a steep slipface that dips into the central scour, although the angle of this surface can range from only a few degrees to angle-of-repose for the sediment (~20°–35°). The position of these faces is controlled by the momentum ratio between the confluent channels, with tributary mouth bars migrating further into the junction as the discharge from that channel becomes a greater fraction of the combined confluence flow.

Third, bars may form within regions of flow separation downstream of the junction corners. Flow separation provides a low velocity region into which sediment can accumulate and these bars may show appreciable fining of sediment since only the finer grained sediment can be entrained into this area. Accumulation of sediment in this region will alter the velocity and pressure gradients within this zone and may lead to a lessening of the extent and influence of flow separation.

Finally, mid-channel bars may form in the region of flow deceleration downstream of the junction scour, especially in 'Y'-shaped junctions, or where sediment delivery is high, and they may mark regions of deposition of sediment eroded at the junction scour. Ferguson (1993) has identified the confluence–diffluence unit as a fundamental braided river building block, in which confluence scour creates the sediment that, as the channel widens to cope with the increased discharge, encourages mid-channel bar development and diffluence formation. However, little study has been conducted on sediment transport through confluences, although experimental work suggests that the bed scour may be a zone of reduced transport rates and that sediment may be routed around and not through the scour. This also reflects the streamline pattern within the junction (Figure 26) and the influence of both shear layers and secondary flows within the confluence.

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JIM BEST AND STUART LANE

CONTINENTAL SHELF

The continental shelf generally is defined as the zone adjacent to a continent or around an island that is between the shoreline and a noticeable break in slope, the shelf break, to the steeper continental slope or, where there is no break in slope, to a depth of about 200 m. Along with the coastal plain, continental slope and continental rise, the shelf is considered part of the continental margin (Gary *et al.* 1972: 153) and usually is synonymous with the term continental platform (Baker *et al.* 1966: 38). The division of the shelf into the inner, mid (occasionally) and outer continental shelf is arbitrary and may be based on logical or scientific criteria, such as the depth to which waves agitate the seafloor, or by legal criteria, such as the geographic limit of jurisdiction by a government. In many regions, the continental shelf is physically continuous with the coastal plain; the separation between the two being the location of the shoreline. The dynamic nature of the continental shelf is symbolized by the active character of the shoreline which moves laterally and vertically in spatial and temporal scales that vary by orders of magnitude.

The shelf is important for several reasons. It is a zone of many physical and biological transitions from oceanic to terrestrial conditions and processes. As everything that moves from land to the ocean must pass across the continental shelf, the suite of processes acting on the shelf is vitally important. The shelves are regions of abundant biological activity as there are substantial supplies of nutrients from both upwelling and upland runoff and there generally is good light penetration. Finally, the continental shelves are sites of major economic interest ranging from the commercial and recreational fisheries of the shelf waters to the sands, gravels and other hard minerals of the surficial sediments to oil and gas that have formed from included biotic sediments and accumulated in any of several types of traps within the body of the shelf.

At the smallest scales, the shoreline and, hence, the boundary between the coastal plain and continental shelf shifts within seconds and hours in response to waves and tides. While, probably more importantly, the multi-millennial, glacial-eustatic SEA-LEVEL changes during the Quaternary have moved the shoreline several tens of kilometres laterally and a hundred or so metres vertically. Furthermore, consequences of local or regional tectonic activity, which usually is spasmodic, are additive to eustatic trends. The presence, or absence, and cause of the tectonics contribute to the overall form of the shelf. The proximity of the continental shelf to the edge of a crustal plate (see PLATE TECTONICS) and the type of inter-plate dynamic play crucial roles in the form and function of the shelf.

Perhaps the least geologically mature shelves are those along convergent plate boundaries and other ACTIVE MARGINS as commonly occur around much of the Pacific Ocean and along the northern shore of the Mediterranean Sea. Although this situation presents a potentially complex and geologically interesting continental margin, the rate of tectonic activity tends to limit the length of time during which marine processes are able to act on a specific body of sediment or location on the continental shelf. However, the same processes that restrict the geographic domain of the shelf result in a rapid, gravity driven flux of material between the often steep and high, near-coastal continental areas and the deep ocean. Milliman and Syvitski (1992) indicate that the small drainage, high relief river systems of active margins contribute a vast quantity of sediment to the ocean basins. Residence time of sediment on the shelf is short and the movement of the sediment across the narrow continental shelf mostly is controlled by oceanographic processes that respond to shelf morphology among other factors. As an example, the zone in which WAVES shoal and resuspend bottom sediments is relatively narrow. This narrowness, in turn, results in sharp gradients in the intensity of wave transformation and related processes.

The contrasting situation is a continental shelf on a PASSIVE MARGIN well removed from a spreading centre, as is the situation along much of the Atlantic coasts of North and South America, Europe and Africa. Such broad, gently sloping continental margins can be significant sites of sediment accumulation over an extensive time. Studies along the east coast of North America indicate a kilometres-thick depositional sequence that began with the filling of early Mesozoic rift valleys (see RIFT VALLEY AND RIFTING) or basins and continues through the present.

Large-scale – many tens of metres – changes in sea level play a major role in the development of the continental shelf. Wright (1995), studying the mid-Atlantic shelf of North America, considers the cumulative time during which any portion of the seafloor potentially is subject to wave energy of sufficient magnitude to agitate the bottom sediments. This zone extends from the shoreline/surf zone offshore to a depth determined wave dynamics and assumptions about the likelihood of specific waves occurring within the area. The width of this zone of bottom agitation primarily is a function of the slope of the shelf surface. The rate of movement of the zone across the shelf is a function of both the rate of sea-level change and the bottom slope. In regions such as that studied by Wright (1995), where sea-level history mainly is governed by eustasy, the determination of the duration of potential bottom activity is comparatively straightforward whereas in areas with a complex history, with major tectonic or glacio-isostatic sea-level component (Kelley *et al.* 1992), the process history is more complex.

In addition to growth by upward or outward sedimentation, other factors can influence the trapping of sediments and the subsequent form of the shelf. Lengthy barrier reefs, shore parallel lines of DIAPIRs, fault blocks or folds can form dams to cross-shelf sediment transport with consequent ponding of sediments. In the situation where the offshore shelf dam has substantial relief and catches a significant quantity of sediment, the mass of the accumulated sediments can trigger isostatic subsidence which results in a deepening of the depositional basin and further trapping of sediment. This seems to have been the case with the growth of the up to 15-km thick Baltimore Canyon Trough which appears to have formed both as the fill in Mesozoic grabens or rift basins and, in places, behind a Jurassic/Cretaceous barrier reef (Schlee 1980).

Several factors including shelf width and slope, rate of change of relative sea level, the availability of sediment and the characteristics of that sediment, and the intensity of physical oceanographic processes determine whether a continental shelf builds laterally or vertically or does not accrete while serving as a conduit for sediment moving from the continent to the deep sea. Similarly, the interaction of the rate and locus of sediment deposition on the shelf with the rate of sea-level rise influences whether an area experiences marine transgression or regression. An understanding of the occurrence and forms of RIVER DELTAS may serve as a surrogate for a similar knowledge of continental shelf growth especially as most deltas grow on or across the shelf. These same factors in combination with others, such as climate, determine the character of sediments that are resident on and within the shelf. Hayes (1967) observed that mud is a major constituent of inner continental shelf sediments offshore of areas with high temperature and high rainfall (strong CHEMICAL WEATHERING), coral is most common in areas with high temperatures, gravel is most common offshore of areas with low temperatures (where MECHANICAL WEATHERING dominates and there is substantial ice transport of large particles), and that rock is abundant in cold areas (perhaps due to scouring of sediments by ice) but is strongly correlated with the slope of the inner shelf.

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CONTRIBUTING AREA

In hydrological terms a contributing area is the part of a DRAINAGE BASIN that provides stormwater RUNOFF GENERATION. The link between precipitation input and catchment outflow is largely determined by variability in soil moisture storage and the spatial distribution of contributing areas for surface runoff. Almost all stormwater runoff is generated by surface or near-surface flow processes. Therefore runoff-contributing areas within drainage basins are mainly dominated by subsurface stormflow and OVERLAND FLOW. Two processes can generate overland flow. Infiltration-excess overland flow, occurs when precipitation intensity exceeds the rate of water infiltration into the soil. This process tends to occur in catchments in semiarid regions where natural vegetation is sparse or where there has been disturbance of the land (e.g. extensive agriculture). The second process is saturation-excess overland flow, which occurs when precipitation falls on a saturated soil surface. During a storm, when antecedent soil-moisture conditions in a catchment are high, the water table may temporarily intersect with the ground surface producing saturation-excess overland flow.

The spatial extent and pattern of runoff-contributing areas is affected by climate, soil and topography. Contributing areas of infiltration-excess overland flow are determined by the interaction of rainfall intensity and soil permeability. The least permeable soils in a basin are the most likely to contribute infiltration-excess overland flow. As rainfall intensity increases, areas with more moderate permeability also may contribute overland flow.

However, at the start of rainfall soil moisture is not evenly distributed but is concentrated in the areas adjacent to perennial water courses and in topographic hollows. Overland flow may be generated by return flow when seepage is concentrated and surface soils become fully saturated. Under these circumstances the water table is high and ground water is in close proximity to the surface. These areas preferentially generate storm runoff so the storm hydrograph peak is generated from a relatively small part of the catchment – the partial contributing area (Betson 1964). This runoff-producing area will expand during the course of a storm.

Figure 27 shows the extent of saturation in a small drainage basin at three stages: pre-storm, mid-storm and late storm. Prior to a storm the area of saturation is preferentially concentrated in hollows and in soils adjacent to stream channels. As the storm progresses the saturated area expands into the hillslope hollows at the channel heads creating saturated overland flow from return flow. This coalesces into stream flow resulting in extension of the channel network. By late storm, channel heads are fully saturated and small perennial streams are flowing. The question as to where channels begin has been addressed in a model by Montgomery and Dietrich (1988) who predict the contributing area required for channel initiation in channel heads generated in landslide hollows. It follows that the areas contributing to runoff in a drainage basin are fairly restricted, occurring mainly at the base of slopes or channel heads where subsurface runoff is at its maximum and groundwater tables are very shallow; where subsurface flow converges in the soil in hillslope hollows; and areas of reduced soil moisture storage.

The importance of the contributing area idea is underpinned by several important hydrological concepts. Betson (1964) developed the concept of partial area storm runoff. This was based on a series of simple mathematical models that used Hortonian infiltration theory to predict the areas contributing to runoff during a storm. The

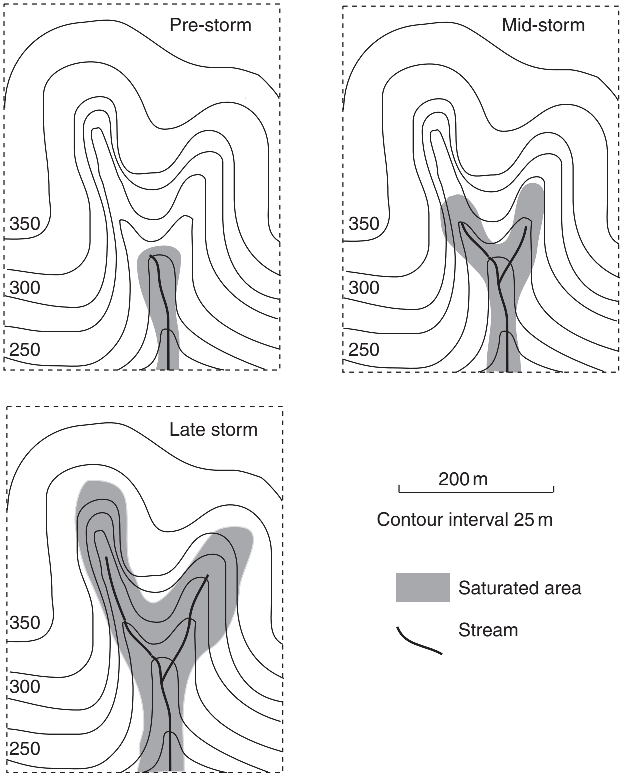

Figure 27 Sequence of expansion of the saturated area of a first-order stream catchment in response to a storm event

equations developed, which can be thought of as functions of apparent watershed infiltration capacity, demonstrate that runoff originates from a small, but relatively consistent, part of the catchment. Using the basic hydrological variables of storm precipitation, storm duration and runoff volume, Betson defined the contributing area as peak stream runoff divided by peak rainfall intensity. This ratio defines the effective runoff-producing area of a drainage basin, which can be expressed as a percentage or proportion of the total catchment area. This is calculated over a storm as total storm runoff divided by total storm rainfall. Typical values are less than 10 per cent for small well-vegetated catchments. The observation that storm runoff frequently occurs from only a small part of the catchment and the size of the runoff-contributing area does not vary greatly within a drainage basin is the basis of the partial area storm runoff concept. However, because this idea is based on Hortonian infiltration-excess runoff theory this is not generally applicable to all catchments.

Working in the humid forests, workers began to recognize the importance of subsurface stormflow generation by throughflow and saturated overland flow at rainfall intensities far less than infiltration excess overland flow (Troendle 1985). This led to the concept of the variable source area, whereby storm runoff was generated in only certain parts of the catchment (Hewlett 1961). The extent of the runoff-generating areas varied from storm to storm and from season to season. On lower slope where groundwater levels are nearest the surface and soil water seepage results in elevated soil moisture storage during a storm, subsurface flow may resurge towards the base as saturated overland flow. Hewlett (1961) and Hewlett and Hibbert (1967), working in forested catchments of North Carolina, demonstrated the importance of this runoff mechanism as opposed to infiltration-excess overland flow so widely popularized by Horton. Other work, particularly by Dunne and Black (1970) working in Vermont, USA clearly established saturation overland flow could be the dominant source of stormwater runoff in a stream.

These ideas are manifest today in the partial contributing area concept which is implicit in the dynamic contributing area concept in recognition of the fact that the area contributing runoff is not fixed but expands during a storm as the saturated areas at the foot of slopes and channel heads extend. When precipitation stops and slopes begin to drain the contributing areas contract. Given that contributing areas are defined by the spatial pattern of surface storm runoff, including overland flow, topography is fundamental in determining the extent. For example, hillslope hollows and swales tend to concentrate saturated overland flow.

Contributing areas of saturation-excess overland flow are determined by the interaction of topography and soil-moisture conditions (Anderson and Burt 1978). The degree of concentration is determined by the area drained per unit contour length (a) and the local slope gradient (s). The a/s index (Kirkby 1978) defines areas of flow convergence and divergence that dictate local drainage conditions for both saturated overland flow and seepage. This topographic control on saturation-excess overland flow can be quantified for the drainage basin as a whole using the topographic wetness index (TWI) (Wolock and McCabe 1995). The TWI is calculated as $\ln(a/s)$ for all points in a catchment. The areas of a catchment with the highest TWI values are the most likely to contribute saturation-excess overland flow. During dry periods when soil-moisture storage is low, only areas with the very highest TWI values are likely to be saturated and contribute overland flow runoff. Under saturated conditions areas with lower TWI values will contribute to runoff.

Land use strongly affects the nature of runoff within a catchment, both in terms of physical processes and solute dynamics. Factors such as surface vegetation, soil permeability and land management practices determine the relative importance of runoff from different types of land use. Furthermore, the pathway of flow through the soil is likely to alter the solute balance of stormwater runoff (e.g. the take-up of nitrates from agricultural fertilizer). In this respect the land use not only influences runoff pathways but will also be important in controlling the sources, types and amounts of contaminants that enter runoff. Furthermore, the dominance of surface or near-surface flow processes in generating storm runoff is an important consideration in the stability of slopes. Many soil mechanics problems can only be addressed by having a good knowledge of hillslope hydrology.

The concept of contributing areas within drainage basins has provided much better understanding of stormwater runoff mechanisms. This has led to better hydrological predictions and the development of distributed runoff models. These models can now be coupled with sediment transport and erosion models to provide realistic simulations of drainage basin development (Willgoose *et al.* 1991).

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SEE ALSO: drainage basin; models; overland flow; runoff generation

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CORAL REEF

Coral reefs are natural structures of calcium carbonate made largely from the skeletons of hard corals and coralline algae. Some modern reefs have been forming for millions of years and can stretch for hundreds of kilometres off tropical coasts.

Distribution in time and space

Coral reefs are found mainly between 25°N and 25°S. The reef-building (hermatypic) corals and associated organisms live best in sea-surface temperatures between 25 °C and 29 °C. Hermatypic corals mostly live only within the upper few metres of water, the 'photic' zone into which sufficient light can penetrate for their symbiotic algae (zooxanthellae) to be able to photosynthesize.

In a general sense, the distribution of fossil coral reefs suggest that sea-surface temperatures have constrained their spread since their appearance in the early Triassic (Birkeland 1997). At a sub-regional scale, other factors were important. For example, the presence of terranes in the central tropical Pacific aided the dispersal of corals across the wider-than-present Pacific during the Palaeozoic and much of the Mesozoic (Grigg and Hey 1992). West–east ocean currents helped the development of coral reefs in the easternmost Pacific during the Cretaceous, when the gap between the Americas was open but species exchange gradually became less during the Tertiary as the Panama Isthmus rose. In the Hawaii group, coral reefs became established only during the Oligocene following the intensifying of the North Pacific ocean-surface gyral circulation (Grigg 1988). Subsequent changes in species composition may be an effect of episodes of extinction and recolonization associated with Quaternary climate changes.

As temperatures and sea levels oscillated during the Quaternary, coral reefs were alternately exposed and drowned. During glacial periods, when sea levels were low, the distribution of coral reefs was much less and in marginal areas of the modern coral seas (like the Hawaiian Islands; Grigg 1988) reefs died out entirely. Owing to cooler temperatures, coral reefs grew at slower rates, and many were comparatively ephemeral. As temperatures increased and sea levels rose at the end of the glacial periods, reefs gradually became reestablished across wider areas of the coral seas. Depending on oceanographic factors, upward-growing coral reefs were either able to 'keep-up' with rising postglacial sea level, or they later were able to 'catch-up', or they had to 'give-up' and thereby forming a drowned reef (Neumann and MacIntyre 1985).

Drowned reefs occur in many parts of the Pacific and Indian Oceans in particular. Most are thought to have failed to keep up with rising sea level during a period of sea-level rise, for reasons associated with climate and sea-level history, palaeolatitude, seawater temperature, and light (Flood 2001). Other 'drowned' coral reefs, particularly those on the flanks of the Hawaiian Ridge, have slipped hundreds of metres downslope.

In many parts of the world, but especially near convergent plate boundaries, coral reefs are found raised above their modern counterparts and, as such, often provide important insights into reef structure and history (Plate 28). Emerged reef staircases on islands like Sumba in

Plate 28 The Talava Arches on Niue Island, central South Pacific. Niue is a fine example of an uplifted atoll, with a well-preserved atoll reef (now 70 m above the modern reef) and lagoon floor. Around the fringes of the emerged atoll reef are a series of emerged fringing reefs. The emerged reef shown here dates from the Last Interglacial period. The modern reef here is rising and is consequently narrow except in embayments as shown here (Photo by Patrick D. Nunn)

Indonesia have been studied in detail (Pirazzoli *et al*. 1991).

In the Pacific and Indian Oceans during the Holocene, keep-up coral reefs grew above their present levels around 4,000 cal. yr BP and have since been exposed as sea level fell. These fossil reefs form the cores of many reef islands (see CAY) in the central Pacific and have been critical factors in their habitability and persistence.

On the basis of their form, coral reefs can generally be either FRINGING REEFs, barrier REEFs or ATOLL reefs. Fringing reefs are juvenile, sometimes ephemeral, and grow outwards from a coast. Barrier and atoll reefs are older, often being composed of reefs of many different ages; reef upgrowth during postglacial periods has been followed by subaerial exposure and erosion during the following glacial period, followed by renewed upgrowth. A good example is that of Midway Island in Hawaii where reef dating back to the mid-Tertiary has been cored (Lincoln and Schlanger 1987). Barrier reefs are separated from a nearby coast by a lagoon while atoll (or ring) reefs enclose a lagoon. The three types were first linked by Darwin (1842) in his Theory of Atoll Development. In this he envisaged that a young volcanic island would develop a fringing reef. As the island subsided so the fringing reef would become transformed into a barrier reef and finally, as the last vestiges of the island were submerged, an atoll reef. Deep drilling of atolls demonstrated the essential correctness of Darwin's model.

Coral reefs in geomorphological research

Since corals are temperature-sensitive organisms, we can learn a lot about palaeoclimates from studying their fossil distributions (see above). We can also use coral reefs as palaeosea-level indicators, both for the Last Interglacial when it is of interest to know whether or not sea level reached 6 m above its present level, as suggested by studies of the emerged reef series on the Huon Peninsula of Papua New Guinea (Chappell 1983). In the Pacific, studies of Holocene emerged reefs have given us much information about the sea-level maximum about 4,500 cal. yr BP (see Nunn and Peltier 2001, for example) and another about 650 cal. yr BP which marked the start of the 'AD 1300 Event' (Nunn 2000). There have also been successful studies of stable isotopes in long-living corals to generate climate data prior to the start of the instrumental record in key regions such as the South Pacific (Quinn *et al*. 1993).

Much research has focused on modelling the relationship between coral reefs and the shorelines which they commonly adjoin, particularly in terms of sediment production, lagoonal dynamics, beach nourishment and shoreline erosion; good studies are those of Munoz-Perez *et al*. (1999) and Hearn and Atkinson (2001). It is clear, for example, that along many tropical coasts, coral reefs are the main producers of the fine-grained sediments which supply nearby beaches and that, should those reefs become degraded, then these beaches can become starved of sediment and destabilized.

Human impacts on coral reefs

It has been realized only comparatively recently how fragile coral-reef ecosystems are, and how much they have been affected by and/or are vulnerable to a variety of human impacts, direct and indirect (Bryant *et al*. 1998). Recently evidence has been presented suggesting that the first human colonizers of some remote Pacific Island groups ~3,000 years ago inadvertently brought with them alien organisms which occupied coral reefs causing reef-surface growth to cease for several hundred years (Nunn 2001). Modern human impacts are more familiar and better understood. These include direct impacts, ranging from the overexploitation of reef organisms (including corals) for sustenance or sale to the dynamiting of reef waters to maximize fish catches, which commonly cause structural reef damage. Indirect impacts are from pollution, including excessive sediment inputs from logging into nearshore areas and chemical pollutants from mineral processing or domestic waste disposal, for example.

Many coral reefs have become degraded as a result of such impacts, manifested as a loss of corals and associated reef organisms, and a reduction in species diversity. Certain more hardy organisms such as sea grasses and various algae (especially *Halimeda*) often cover such degraded reefs. Sometimes reef degradation allows reef predators like the crown-of-thorns starfish (*Acanthaster*) sufficient access to result in an infestation which then exacerbates the process of degradation.

Tourism along tropical coasts often focuses on coral reefs; around 80 per cent of the visitors to the Maldives in 2001 wanted to dive on their reefs. While reef-associated tourism can be sustainable, in many cases it is not because the effects of constructing tourist infrastructure and the effluents which must be disposed of when a large hotel or resort exists in a particular place all reduce the health of the reef ecosystem.

Coral-reef conservation initiatives, including the establishment of marine-protected areas, are often well intentioned but ineffective. Good examples are found in parts of the Caribbean and tropical Pacific Islands where the idea of marine reserves is anathema to people who have been accustomed to free access to reef areas for susbistence purposes (Birkeland 1997).

The future of coral reefs and the implications for geomorphology

Many coral-reef ecosystems have become significantly degraded as a result of human impacts (see above). Many reefs are now being pushed to the brink of extinction because of the additional stress associated with rising sea-surface temperatures (Hoegh-Guldberg 1999). High levels of stress often cause corals to become bleached, the loss of colouration being associated with the ejection of the symbiotic algae that live within coral polyps. Whole reefs can die as a result of bleaching episodes, and there are no instances where a formerly bleached reef has been able to recover its former state. As sea-surface warming continues over the next few decades, so the instances of bleaching resulting from prolonged periods of high temperatures (often associated with El Niño) are likely to increase. The Great Barrier Reef is likely to be experiencing annual bleaching events by 2030.

The implications of regular bleaching for the world's coral reefs are extremely serious, and will have huge implications for many subsistence coastal dwellers in the tropics, who depend daily on reefs for sustenance, and for those countries which depend heavily on revenue generated from reef-associated tourism. The effects for coastal landscapes will involve drastic reductions in the amounts of fine calcareous sediment being generated in reef-lagoon areas, perhaps with many beaches disappearing as a result. This may in turn increase the vulnerability of sandy shorelines to erosion, also an effect of larger waves crossing reefs which are unable to grow upwards in response to projected sea-level rise (Birkeland 1997).

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SEE ALSO: atoll; fringing reef; reef

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CORNICHE

Corniches are narrow organic ledges, 0.5 to 2 m in width, growing on steep rock surfaces at about mean sea level. The best examples are on limestones where notches (see NOTCH, COASTAL) develop in the spray zone. Corniches in the northwestern Mediterranean consist of algae, particularly the calcareous alga *Tenarea tortuosa*, although Serpulid (see SERPULID REEF) worms or Vermetid (see VERMETID REEF AND BOILER) gastropod tubes can play a similar role. Although the interiors are generally quite hard, corniches cannot resist very strong waves and they are best developed in inlets on exposed coasts.

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ALAN TRENHAILE

CORROSION

Corrosion is synonymous with solutional erosion, the erosion of material by chemical activity. The majority of studies of corrosion have been undertaken on carbonates and these are the primary focus of this entry. However, similar considerations apply to evaporites and the estimation of gypsum corrosion rates is particularly problematic because of the more rapid solution and the consequently greater spatial and temporal variability of dissolution.

Corrosion rates are commonly expressed as mm/1000a, implying that all corrosion contributes to surface lowering and that environmental conditions have remained broadly the same for millennia. The former is incorrect, particularly in karst, while the latter is also highly questionable. The preferred unit is $m^3\ km^{-2}\ a^{-1}$ and $1\ mm\ ka^{-1}$ is equivalent to $1\ m^3\ km^{-2}\ a^{-1}$. Where surface lowering is measured directly then units of $\mu m\ a^{-1}$ are appropriate.

Limestone corrosion rates may be estimated from knowledge of dissolution kinetics, runoff, carbon dioxide and temperature, but there remains a need for field measurements to provide actual values of regional denudation; to compare rates in contrasting environments and by different processes; to understand landform evolution; and to understand how processes operate in a complex natural environment as opposed to the laboratory. When evaluating results from past studies it is important to understand what was actually measured and how the corrosion rates were calculated. Most field measurements of corrosion in carbonate karst were based on spot samples, with denudation being estimated from the Corbel formula. This suffers several problems, the three most important being: the carbonate concentration is frequently the average of a few spot measurements, with the implicit assumption of a linear relationship between carbonate hardness and discharge; carbonates present in solution are assumed to only come from karst denudation; and measurements are usually made at only one point, commonly the output of a drainage basin, with the implicit assumption that this is representative of conditions upstream.

Where water samples have been collected over a range of flow conditions it is apparent that the relationship between dissolved load and discharge is usually non-linear and particularly in small drainage basins may be complicated by hysteresis effects (usually higher concentrations per unit discharge on the rising limb). It is virtually impossible to correct for hysteresis, but by collecting samples over a range of discharges it is usually possible to construct a reliable discharge-concentration or discharge-load rating curve. This can be applied to the discharge curve and the results summed to obtain the total annual solute load. Greater accuracy may be obtained using a logging conductivity meter, developing a conductivity-concentration rating curve, and using this to predict the concentration at each measured discharge.

Having computed the total solute load (TSL) at a point it is important to realize that this is made up of corrosion of karst rocks by both autogenic (CKAu) and allogenic waters (CKAl), less any deposition of previously dissolved material (D), together with corrosion of non-karst rocks by allogenic waters (CNK), solute accessions in rainfall and snowfall (AC), and any anthropogenic inputs such as fertilizers (AN). The gross karst solution is then (CKAu + CKAl) whereas the net karst solution is (CKAu + CKAl − D). Where precipitation of previously dissolved carbonates is minimal then gross and net solution will be similar, but elsewhere failure to account for deposition may result in a significant underestimate of gross denudation, which is the real measure of relief transformation. In contrast, failure to take into account the solution of non-karst rocks and solute accessions in precipitation will result in an overestimate of karst corrosion. Error in estimating corrosion rates can arise from many sources and even in a careful study using hydrochemical budgeting and taking into account non-denudational components potential errors of around 25 per cent are possible.

Corrosion rates for whole drainage basins derived by sampling of water at the basin outlet are unlikely to be representative of any specific location within the basin. This information may best be obtained by an extension of the hydrochemical budgeting method discussed above. Water samples are collected from the full range of sites – bare limestone surfaces, the soil zone, the subcutaneous zone, the main body of bedrock, and cave streams in both vadose and phreatic zones. These, together with estimates of the proportion of water following the various pathways through the system, permit the breaking down of the overall corrosion budget. Those few studies that have been made show that a high proportion of corrosion (50–85 per cent) occurs within several metres of the surface in the soil (if present) and subcutaneous zone (uppermost bedrock). Caves account for very little of the erosion when averaged over the whole basin.

The principal drawback of the hydrochemical approach is that it requires frequent, ideally continuous, measurement of discharge and sufficient samples to establish the pattern and extent of variations in solute concentrations. As this is not always possible alternative methods that integrate erosion over a longer time period have been derived. The two most commonly used are the micro-erosion meter and rock tablets. In contrast to the hydrochemical method these are highly site-specific and may only be used to assess corrosion rates on bare limestone surfaces, in the soil zone, at the soil–bedrock interface, and in cave streams. Tablets have been found to give estimates two orders of magnitude less than those calculated using hydrochemical data. The most likely explanation is that natural rock surfaces come into contact with larger volumes of water than do isolated rock tablets, simply because of their greater lateral flow component. Thus, the two methods measure fundamentally different phenomena and the hydrochemical method provides the only reliable means of estimating corrosion rates on limestone surfaces. Different problems arise if tablets are placed in cave streams as they will project above the natural surface and as a consequence are likely to erode more rapidly. They are also likely to suffer from abrasion as well as corrosion, although this can be exploited by placing the tablets in nylon cages with differing mesh sizes and comparing the erosional losses suffered.

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SEE ALSO: dissolution

JOHN GUNN

COSMOGENIC DATING

Cosmogenic dating is a group of related techniques for estimating landform ages and erosion rates. It is based upon the generation of rare isotopes within minerals by cosmic rays. Primary cosmic rays composed largely of highly energetic protons interact with gases in the Earth's atmosphere to produce showers of secondary subatomic particles, mostly neutrons and muons. These secondary cosmic rays induce nuclear reactions within the Earth's terrestrial surface, producing cosmogenic nuclides. The length of surface exposure, or alternatively, the rate of surface erosion, is computed from the concentration of cosmogenic nuclides in a landform.

Six cosmogenic nuclides have found widespread application in geomorphology (Table 9). These are stable isotopes of the noble gases helium and neon (^{3}He and ^{21}Ne), and radioactive isotopes of beryllium, carbon, aluminum, and chlorine (^{10}Be, ^{14}C, ^{26}Al , and ^{36}Cl). The nuclides ^{10}Be, ^{14}C and ^{36}Cl are also produced within the atmosphere by cosmic rays. The best known example of atmospheric production is ^{14}C which forms the basis for radiocarbon dating (see DATING METHODS). To avoid confusion, nuclides generated within mineral lattices in the Earth's solid surface are termed *in situ*-produced terrestrial cosmogenic nuclides (TCN).

Most TCN production is from neutron spallation (Lal 1991). TCN spallation occurs when a

Table 9 Properties of *in situ*-produced terrestrial cosmogenic nuclides (TCN)

Nuclide	Mean lifetime (yrs)	Host mineral
^{3}He	Stable	Olivine, clinopyroxene
^{21}Ne	Stable	Olivine, clinopyroxene, quartz
^{10}Be	2.2 Myr	Quartz
^{14}C	0.82 kyr	Quartz, calcite
^{26}Al	1.0 Myr	Quartz
^{36}Cl	430 kyr	Calcite, dolomite, whole rocks

secondary neutron with energy > 10 MeV collides with a target nucleus in a mineral lattice, breaking protons, neutrons or clusters of these particles from the nucleus. Spallation products always consist of an isotope of lower atomic number than the target. As neutrons do not penetrate deeply in rocks, most neutron spallation occurs within about one metre of the surface. Thermal neutrons (energy ~0.025 eV) are absorbed by some target nuclei, causing radioactive decay and production of a cosmogenic isotope. Thermal neutron production is important for cosmogenic ^{36}Cl. Muons also create cosmogenic nuclides but at rates much lower than neutron spallation. Muons penetrate far more deeply than neutrons, creating measurable quantities of cosmogenic nuclides at depths of over 20 metres (Granger and Muzikar 2001).

The production rate of cosmogenic nuclides by all reaction mechanisms is low, ranging (at sea level and latitudes $> 60°$) from about 5 to 6 atoms g^{-1} a^{-1} for ^{10}Be to about 120 atoms g^{-1} a^{-1} for ^{3}He. Cosmic rays are attenuated by the atmosphere and the geomagnetic field; consequently production rates vary significantly with altitude and latitude. For this reason, TCN production rates are always quoted for sea level and high latitude, and scaled to the altitude and latitude of study sites using empirical functions (Lal 1991; Stone 2000; Dunai 2000). Production rates must be precisely known for reliable TCN results. This is a difficult task because both atmospheric shielding and geomagnetic field intensity vary with time. Calibration sites are used to determine production rates. At a calibration site, TCN concentrations are measured in independently dated geomorphic surfaces with near-zero erosion rates such as lava flows, glacially eroded bedrock or large landslides.

Applications of TCN fall into two main categories: surface exposure dating and measurement of erosion rates. Both applications yield model results with accuracy highly dependent on the validity of simplifying assumptions. In exposure dating, the first requirement is that the geomorphic surface must have formed over a short time period. Examples of such surfaces include fault scarps (see FAULT AND FAULT SCARP), LAVA LANDFORMS, LANDSLIDES and ERRATIC boulders. Surfaces forming incrementally over long periods of time have cosmogenic nuclide concentrations best interpreted in terms of erosion rates. The second requirement is that the geomorphic surface be free of TCN at the time of surface formation. Remnant TCN from past periods of surface exposure is termed nuclide inheritance. Lava flows and large glacial erratics generally have little or no nuclide inheritance. The final requirement for accurate exposure dating is that the primary geomorphic surface form must be preserved over the period of exposure. Erosion rates must either be known, or be assumed to be zero. Surface exposure dating therefore requires careful analysis of landscapes and sampling of surfaces experiencing very low rates of erosion. The requirement of near-zero erosion limits the age range of TCN exposure dating. In most geomorphic environments, reliable exposure ages generally range from about 5,000 years to less than 100,000 years. Younger ages are limited by detection limits for measuring TCN while older surfaces are generally destroyed by erosion or buried by sediments. The polar deserts of east Antarctica are a major exception, with exposure ages of over 5 million years. The precision of exposure ages and erosion rates, as estimated by analytical errors, varies with isotope and application but generally ranges between ± 3 per cent to 15 per cent.

In TCN erosion rate studies, an assumption of equilibrium between TCN production and loss by erosion and radioactive decay is made (Bierman and Steig 1996; Granger *et al.* 1996). Under these circumstances, exposure time is not important and TCN concentrations vary inversely with erosion rates. For example, steep hillslopes with high erosion rates have low TCN concentrations because of the short residence time of target minerals within the zone of production. Averaging time is the time necessary to achieve equilibrium conditions. The lower the erosion rate, the longer the averaging time. TCN averaging times for erosion rates typical of temperate climates range from about 100,000 years to about 5,000 years. Averaging erosion over such timescales makes the TCN method very useful for investigating links between climate and tectonics, and for establishing baseline erosion rates unrelated to human activities. Two types of sampling are applied in TCN erosion rate studies. Bedrock samples give information about minimum rates of landscape lowering and the influence of lithology on erosion rates. Alluvial samples average erosion rates for the contributing catchment and therefore are easiest to compare with traditional methods of measuring short-term erosion rates such as suspended sediment studies.

TCN vary greatly in terms of ease and cost of measurement, sample preparation and host minerals. ^{3}He and ^{21}Ne are measured in olivine and clinopyroxene using noble gas mass spectrometry techniques similar to those employed for ^{40}Ar/^{39}Ar studies. They are primarily used for dating mafic volcanic rocks, and for studies of long-term landscape evolution in Antarctica where extremely low rates of erosion require the use of a stable nuclide (Summerfield *et al*. 1999). The most used TCN are ^{10}Be and ^{26}Al. These nuclides are popular because the host mineral (quartz) is present in the majority of geologic settings, production reactions are relatively simple and well understood, and both nuclides can be measured in the same sample. Since the mean lives of ^{10}Be and ^{26}Al differ significantly (Table 9), measurement of the nuclides in the same sample can constrain both erosion rate and exposure time as well as indicate periods of burial (Lal 1991; Bierman *et al*. 1999). It is also possible to use this nuclide pair for dating the burial of sediments (Granger and Muzikar 2001). Measurement of ^{10}Be and ^{26}Al is by accelerator mass spectrometry (AMS). Sample preparation requires preparation of high purity quartz separates and removal of atmospheric ^{10}Be with hydrofluoric acid etching. ^{36}Cl is also widely applied to geomorphic problems, particularly in carbonate and volcanic landscapes where ^{10}Be cannot be used. Production rates for ^{36}Cl are less well established than for other TCN because of more complex production reactions. ^{36}Cl is produced by neutron spallation on K and Ca, and by thermal neutron capture on ^{35}Cl. Production rates vary with rock composition, and major and trace element data are needed to compute rates. AMS is also used to measure ^{36}Cl concentrations. *In situ*-produced ^{14}C has not been widely applied in geomorphology because of problems separating atmospheric contamination. However, applications with this nuclide are likely to increase. The mean life of ^{14}C is much shorter than any other TCN, therefore, when measured in conjunction with ^{10}Be and ^{26}Al, it can be used to establish production rates, estimate erosion corrections, and detect periods of burial by sediment or ice.

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COULEE

In western North America, coulee (French *couler*: to flow) is a common term used to describe a dry valley, canyon, gulch or wash. Most coulees were formed rapidly in late glacial times by large discharges of melt water, particularly with the emptying of proglacial lakes (Bretz 1969). Selby (1985: 458) adopts this as a specific origin. Coulees may have ponded water bodies, intermittent or underfit streams. Parallel sets of coulees in southern Alberta, Canada, may have been aligned by regional joint patterns (Babcock 1974), or possibly formed in postglacial time through some imperfectly understood process controlled by prevailing wind direction (Beaty 1975). Less commonly, the term coulee is used to describe a short lobe of viscous lava on the flanks of a volcano and a lobe of debris moved by gelifluction.

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COVERSAND

Originally a Dutch term applied to aeolian SANDSHEET deposits overlying older sediments. Its generic nature has led it to be applied to a range of deposits. However, the common denominator has been its application to sandsheet deposits of cold-climate (see PERIGLACIAL GEOMORPHOLOGY) aeolian origin. The latter is proven by the occurrence in coversands of frost cracks, involutions and ice wedge casts (see ICE WEDGE AND RELATED STRUCTURES), as well as from pollen and beetle evidence obtained from intercalated organic deposits. The aeolian origin of coversands has been determined on the basis of their concordance with dune forms (see DUNE, AEOLIAN), occurrence with VENTIFACTs and/or on the basis of particle characteristics (mineralogy, sorting, rounding, surface matting and textures). Whilst predominantly aeolian derived, coversands can incorporate components of sand derived from other processes, e.g. niveo- and/or fluvio-aeolian (see NIVEO-AEOLIAN ACTIVITY).

Coversands in northern Europe (Schwan 1988) and mid-continental north America although relict, are widespread, extending over 10,000s of km^2 as nearly spatially continuous sheets with flat to undulating relief (less than 5 m) and a notable paucity of dunes (Koster 1988). This differentiates coversand from more recent sand deposits which tend to have been formed into dunes, e.g. Drift sands in The Netherlands (Koster *et al.* 1993). The coversand is typically of a uniform thickness of up to several metres; only in valleys, depressions or against topographical barriers is it thicker. The coversands also tend to be (sub)horizontally stratified, composed of thin beds, setting them apart from the high angle bedding of coastal dune (see DUNE, COASTAL) sands and the cross bedding, troughs and ripples of riverine sands.

Detailed examination of coversand stratification in Europe has led to classification of coversands into two types which are in turn subdivided into two: Older coversands I and II and Younger coversands I and II. The Older coversand is characterized by an alternation of well-sorted parallel-laminated beds of greyish loam/fine sand and yellowish fine/medium sand. The Older coversand I has evidence of more cryogenic deformation and frost wedge casts, especially in its upper layers, than Older coversand II and the two facies are commonly separated by a disconformity, e.g. the Beuningen pebble. The Younger coversand is typically a unimodal, well-sorted, parallel-laminated medium sand with a large sand component derived from local sources. The sand is rarely buried or cross-bedded, has a low relief and has no evidence of ice wedge formation in it. The primary differentiation between the two Younger coversand facies is on sedimentary structures which indicate that the Younger coversand II was deposited under drier conditions.

Fragmentary evidence indicates coversands have been deposited during several Pleistocene glacials and are not unique just to the last glacial cycle. The northerly limit of the relict but extensive European coversands found in Britain, The Netherlands, Germany, Denmark, Poland and the Baltic states is broadly coincident with the maximal position of the Late-Weichselian (Devensian) ice sheet. In general, the last era of north-west European coversand activity started after the last interglacial, increasing in intensity throughout the Weichselian period. Two main phases of coversand deposition have been reported: one around 18,000–15,000 years ago (Older Coversand II) and another more intense period between 14,000–11,000 years ago (Younger Coversand) (Koster 1988; Bateman 1998; Singhvi *et al.* 2001). Older coversand I appears to have been dominantly deposited separate to, but contemporary with, the widespread LOESS deposits of north-western and eastern Europe which were mostly deposited just prior to the last glacial maximum and appear to have stabilized by approximately 13,000 years ago (Singhvi *et al.* 2001). However, evidence also suggests localized environmental conditions blurred these discrete aeolian phases with Older coversand type facies still being deposited in places during the so-called Younger coversand phase (Kolstrup *et al.* 1990; Kasse 1997).

Formation and preservation of the Late-Weichselian coversands is thought to have been aided by enhanced sand sources as a result of glaciation, sparse vegetation, low relief and low sand supply due to periodically wet, frozen or cemented depositional surfaces (Kasse 1997). Use of the orientation of dune morphology, bedding inclination and unit thickness has enabled the reconstruction of palaeo-wind directions. The Older coversands type was deposited by predominantly north-westerly to westerly winds and the Younger coversands deposited by more westerly to south-westerly winds. Such information has been used to inform palaeoclimatic models for north-western and central Europe (e.g. Isarin *et al.* 1998).

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MARK D. BATEMAN

CRATER

Craters are bowl-shaped, approximately circular depressions that typically form by high-energy impact or explosive activity. There is a fundamental geomorphological problem in distinguishing crater origins by volcanic versus impact processes. The latter involve collision with a planetary surface by meteors, comets and asteroids. It is also possible that a crater can form by the explosion, just above the ground, of a meteor or comet, known in this context as a *bolide*. Volcanic craters generally form at the summits of volcanic cones and result from explosive eruptions or the accumulation of *pyroclastic* material in a rim around a volcanic vent. Of course, human activity can produce explosion craters, perhaps the most spectacular of which resulted from nuclear testing. Interestingly, it was the well-funded study of physics for the latter that ultimately led to considerable advancement in understanding the natural process of impact cratering (Roddy *et al.* 1977).

One of the great controversies in planetary geomorphology concerned the origin of craters on the moon. G.K. Gilbert (1893) used geomorphological reasoning to argue that the moon's craters had an impact origin. Until the 1930s, however, most astronomers thought that the circularity of the moon's craters required an origin by volcanic processes. Objects striking the moon, it was thought, would include many oblique impacts, and these would not be circular. Only later in the twentieth century did the physics of the cratering process come to be well understood enough to show that most oblique impacts produced circular, rather than elliptical, craters. Nevertheless, some astronomers continued to argue for the volcanic origin until the Apollo missions of the 1970s returned incontrovertible proof of the impact origin for nearly all lunar craters.

Volcanic craters

Craters can be a variety of depressions associated with volcanic or pseudovolcanic activity, including mud volcanoes, mound springs, hot springs, and even pingos. The geomorphology of truly volcanic craters was reviewed by Fairbridge (1968), who considered the large complex collapse and explosion structures known as *calderas* separately from other volcanic craters. Magmas rich in silica tend to produce highly explosive activity in which the volcanic materials become fragmented into *pyroclastic* rock. Domes of silica-rich volcanic rock, including obsidian, commonly fills preexisting pyroclastic craters. Explosive activity for basaltic magmas produces spatter cones with craters over rift zones, and a variety of pit craters. One of the most famous of these is Halemaumau, a pit crater on the floor of Kilauea Caldera, on Earth's most continuously active volcano at the southern end of the island of Hawaii. There are also many volcanic craters on other planets, including spectacular calderas on the volcanoes of Venus, Mars and Io (the highly volcanically active satellite of Jupiter).

A special type of crater, known as a *maar*, derives its name from the Rhineland dialect of German. The term was originally applied to volcanic explosion craters near Eifel, Germany. Maar craters may be associated with *diatremes*, which are breccia-filled volcanic pipes that form by gas explosions. They also occur within fields of monogenetic volcanic cones, which develop during single eruptive phases. Maar craters usually have a ring of erupted pyroclastic material, and lakes often occur on their floors. They generally form by the interaction of rising lavas with near-surface ground water.

Another interesting crater form is known as a pseudocrater, or rootless cone. These were first recognized in Iceland, where basaltic lava flows advanced over substrates that were rich in water or ice. The interaction of the lava and water produced pyroclastic explosions that formed the craters. The advancing flow may then separate the crater or cone from its source zone. Such features range from a few tens of metres to hundreds of metres in diameter.

Impact crater morphology

It is one of the major discoveries of recent planetary exploration that the surfaces of rocky objects in the solar system are almost all marked by numerous impact craters. These occur over an immense range of size scales. The smallest are microcraters or pits, which form from the impact of micrometeorites or high-velocity cosmic dust grains on exposed rocks. These only form on bodies that lack atmospheres, which would induce the very small projectiles to burn up before impact. Simple craters are larger, bowl-shaped depressions that form on land surfaces. They range up to several kilometres across, and typically have diameters across their rims that are about five times their depths from rim top to crater floor. Simple craters are familiar to many geomorphologists because they were much in evidence during the Apollo landings of humans on Earth's moon. On Earth one of the most famous simple craters is the 1-km diameter Barringer Crater in northern Arizona, also known as Meteor Crater. It is interesting that a major controversy occurred in regard to its origin, with Gilbert (1896) eventually concluding that it had a volcanic origin, despite making a strong argument for impact as well.

Most of the larger craters visible on planetary and satellite surfaces are complex craters. These have rims marked by terraces along their inner margins. Their floors are broad and flat, and there is often a central peak. Such craters are generally from a few tens to a few hundred kilometres in diameter, and they are well known from observations of the moon (Figure 28). Because of their flat floors and very high ratios of width to depth, these features are usually described as impact basins, rather than craters. Much larger impact structures are also known, and many of these are multi-ring basins. They have multiple concentric rings, each consisting of rugged hilly terrain. The floors of these exceptionally large craters are commonly flooded by lava. They can have diameters of up to two thousand kilometres or more.

Recent work has shown that many of the projectiles generating impact craters arrive in groups, rather than as single projectiles. One of the most spectacular examples of this phenomenon was the comet Shoemaker-Levy 9, which broke into fragments as it collided with Jupiter in July 1994. Asteroids also may break up when they interact with a planet's atmosphere. Among the 150 or so

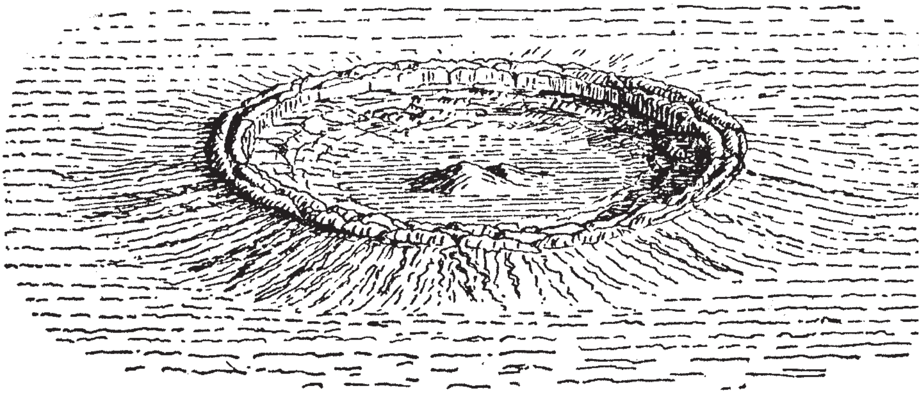

Figure 28 Sketch of a complex lunar crater made by Grove Karl Gilbert (1893: 243)

Plate 29 Henbury impact craters in central Australia. These structures formed when a group of meteors struck a pediment surface less than about 5,000 years ago (Milton 1968). The largest of the craters, part of a tight group of four, is about 150 m in diameter and about 10 to 15 m deep. Note the capture of drainage by the craters

Earth impact sites are many that include multiple craters (Plate 29). The Kaali impacts, which struck Estonia about 2,400 to 2,800 years ago, consist of nine craters, the largest of which is 110 m wide and 20 m deep.

Impact crater processes

Meteors and comets arrive at velocities of many metres per second, causing an immense transfer of energy in an exceedingly short period of time as they strike the surface of a planet. The actual cratering process is surprisingly orderly, and very well known from both theoretical and experimental work (Melosh 1989). The initial phase is contact and compression, which lasts only a few times longer than the time it takes for the projectile to traverse its own diameter. This produces prominent very high-velocity jets of highly shocked material that shoot upward from the margins of the deforming projectile. A zone of phenomenally high pressure is produced at the front of the projectile, as it is deformed by contact with the target material. In the inner solar system the target is usually rock, but in the outer solar system the satellites of Jupiter, Saturn, Uranus and Neptune are commonly icy. The ices are so cold that they generally behave like rock.

Contact and compression is followed by an ejection or excavation phase. The projectile is melted or evaporated by a shock wave propagating into it, while another shock wave propagates into the target. The shock wave is followed by rarefaction waves that decompress the material and generate excavation flows that open up a transient crater. This excavation process may last several minutes, depending on the energy level of the original impact. Material ejected from the crater will then comprise an outwardly expanding ejecta curtain, which has the form of an inverted cone, centred on the impact site. Material deposited from this curtain will comprise an ejecta blanket that covers the terrain out to about two crater radii from the rim. Additional large ejecta blocks may create additional impacts, or secondary craters. These have distinctive morphologies because of the slower projectile velocities, highly oblique paths and radial structure in relation to their source craters.

At the end of the excavation stage the transient crater will often experience collapse and modification. For the larger complex craters this produces terraces and central peaks. The terraces develop by slumping of the crater rim after all material has been excavated. The central peaks represent uplift of the floor material beneath the transient crater cavity. A peak ring may form as the central peak grows and collapses.

Cratered landscapes

Cratered landscapes dominate on the surfaces of rocky objects in the solar system. This is mainly because most of those surfaces are very old. In general, the density of impact craters on a surface corresponds approximately to the age of that surface. However, this relationship holds on very long timescales. Moreover, it is not linear. During the early part of solar system history, the impacting rate was extremely high. From the final accretion of planets and many satellites, about 4.5 billion years ago, until about 3.9 billion years ago for the moon, and perhaps a few hundred million years later for Mars, there was a period of intense heavy bombardment. This produced overlapping craters with sizes up to the scale of the multi-ring basins. The scaling is very regular with many more craters of smaller sizes than of larger. After the heavy bombardment, which was caused by many objects left over from solar system formation, the impacting rates declined by more than an order of magnitude. On the moon these timescales of cratering have been confirmed by radiometric dates on rocks returned to Earth by the Apollo missions. The lunar highlands correspond to the heavy bombardment phase, and much lower crater densities mark the younger volcanic plains of the lunar mare, which occur on the floors of very large impact basins. On Mars there is a similar dichotomy between old, heavily cratered highlands and younger, lightly cratered plains. Unlike the moon, however, many Martian craters are highly degraded by erosion, including the action of fluvial, periglacial and glacial processes.

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SEE ALSO: astrobleme; caldera; extraterrestrial geomorphology; volcano

VICTOR R. BAKER

CRATON

The central core of extensive, stable continental crust in present-day continents that has achieved tectonic stability. All cratons are older than 570 million years, dating from the Precambrian period. Cratons have essentially rigid foundations, composed of predominantly granite and metamorphosed rocks that have been deposited on pre-existing older basement rocks. They are generally low-lying with little relief, as a result of erosion. Cratons have only been affected by EPEIROGENY and are devoid of orogenic features and recent volcanic activity.

The term craton is derived from the Greek word 'kraton' meaning shield and should only be applied to continents and not to oceans, according to the theory of plate tectonics. Cratons are added to by the process of cratonization, an important mechanism for continental growth. Sediments accumulate within thick linear troughs on the cratonic margins. Here, the material is eventually deformed and partially melted onto the existing craton. Early Achaean cratons were smaller and greater in number, yet through the process of cratonization throughout Phanerozoic time cratons became larger and fewer in number as they were fused together.

The area of a craton that becomes exposed is termed a SHIELD. Shields are composed of ancient crystalline basement rocks, and represent the core of the craton. The Canadian Shield is an example;

it is composed of granite and metamorphic rocks (e.g. gneisses), alongside heavily deformed metamorphosed sedimentary (e.g. quartzites) and other volcanic rocks. The term shield is also sometimes employed as a synonym for craton.

The shield is unconformably overlapped at its margins by thin sedimentary units, termed platforms. Platforms are typically *c.*1 km thick and derived from the Palaeozoic and Mesozoic periods, predominantly composed of shallow marine sandstones, limestones and shales.

Since cratons are tectonically stable, sediments tend to spread out widely into any areas of relatively low-lying ground, such as the intra-cratonic basins. These are typically shallow (though can range up to 3,000 m), bowl-shaped, and are characterized by very slow subsidence. Basin sediments thicken regularly towards the centre, yet their fill is discontinuous. As such, the stratigraphy reflects major transgressions across the entire craton, punctuated by periods of stability. Many of them develop as shallow 'sag' lakes, such as Lake Chad in North Africa. The cause of intracratonic basins remains contentious.

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STEVE WARD

CROSS PROFILE, VALLEY

In most introductory physical geography and geology textbooks a distinction is made between V-shaped valley cross profiles, described as characteristic of a system dominated by active fluvial erosion, and U-shaped valley cross profiles, described as characteristic of a system dominated by GLACIAL EROSION. This process-oriented distinction gained wide popularity as a component of classic Davisian landscape classification, particularly in the first half of the twentieth century, and continues to be used in more modern landscape interpretation and analysis. Morphometric analyses have been used to show that glaciated valley cross-section profiles can be approximated by the mathematical equivalent of the letter U, a parabolic equation, whereas fluvial valley side slopes are more nearly linear. In addition, the amount of rock removal required to convert a V-shaped cross-profile geometry to a U-shape has been used as a measure of the glacial erosion component of valley development, and the extent of valley development towards a particular form has been used as a measure of the degree of valley modification by fluvial or glacial processes. However, valley cross profiles include a much wider variety of forms than the two-fold division into V- and U-shaped suggests, and cross-profile forms are not only controlled by glacial and fluvial erosion, but also by patterns of hillslope erosion and deposition, and by patterns of rock resistance to erosion.

Typical explanations for the development of V-shaped valleys in areas with active fluvial erosion include several components: (1) that river erosion is dominantly vertical; (2) that the river is capable of transporting all of the material supplied to it by hillslope processes; and (3) that valley side slopes are steepened to a critical angle for hillslope transport or failure. This ideal set of conditions results in uniform valley side slope angles either side of a central river that is eroding vertically into the landscape with little or no floodplain, i.e. a V-shaped cross profile. However, if the river is not capable of transporting all the material supplied to it by hillslope processes, if there are significant lithological variations along the slope profile, or if different hillslope processes dominate in separate parts of the slope profile, then more complex hillslopes and valley bottoms will develop than the simple linear form required for a V-shaped cross profile.

Typical explanations for the development of U-shaped valleys as a result of glacial erosion rely on the argument either that glacial 'valleys' are actually glacial channels, and that steep side walls and a relatively flat bottom is a characteristic form for fluid flow in channels, or that the cross-sectional pattern of erosion under a glacier includes a wide central maximum leading to steep side walls and a low gradient profile section in the channel centre. Numerical modelling linking ice dynamics, subglacial erosion patterns and cross-profile form development has demonstrated that U-shaped cross sections can result solely from glacial erosion in homogeneous bedrock. However, when spatial variations in rock resistance to erosion are introduced to the model, a wide variety of cross-section shapes can develop, including V-shaped forms. In addition, many glaciated valleys used to illustrate U-shaped valleys or included in morphometric analyses of valley form include substantial depositional components; the U-shaped form arises from the combination of a low gradient valley floor (fluvioglacial deposition), and a concave talus slope (postglacial and ice marginal slope processes) below steep bedrock walls (glacially modified).

Although the distinction between idealized U-shaped and V-shaped valleys for areas dominated by glacial and fluvial erosion is useful, there is in fact a wide variety of valley cross-profile forms. Other and more complex cross profiles result from temporal and spatial variations in processes across the profile, including both erosion and deposition, and from patterns of surface material resistance to erosion.

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SEE ALSO: hillslope, form; hillslope, process; valley

JON HARBOR

CRUSTAL DEFORMATION

Motions of the lithosphere disrupt and modify rocks and the topographic surface. As a manifestation of PLATE TECTONICS, these deformations maintain continental forms that protrude above sea level. Crustal deformations, such as fault offsets and folds, produce diverse constructional landforms dependent on local material properties and surface processes. ACTIVE MARGINs are shaped by competition between deformation and erosion. Deformation occurs at the timescale of plate motions (centimetres per year) but can be slower along individual structures. Recent technologies have revolutionized crustal deformation studies, such as space-based geodesy (e.g. GPS) and seismology that constrain short-term behaviour, dating techniques (e.g. COSMOGENIC DATING) that constrain chronologies of offset geologic markers, and DIGITAL ELEVATION MODELS that permit topographic assessment of large areas.

Crustal deformation leads to OROGENESIS and basin formation over the long term, producing wholesale surface uplift, DENUDATION, and SUBSIDENCE (see TECTONIC GEOMORPHOLOGY). Fluvial systems respond to perturbations in BASE LEVEL where the crust has risen or fallen (Burbank and Anderson 2001). Long-profiles (see LONG PROFILE, RIVER) of stream channels adjust to uplift via KNICKPOINT migration and incision, often leaving behind suites of terraces. Drainage networks may also be modified, as streams can be deflected by zones of uplift or forced to migrate by tilting (see ASYMMETRIC VALLEYs). Sediment loading and gradient changes further influence fluvial form, such as the occurrence of meandering versus BRAIDED RIVER channels. Adjustment to base level in turn affects hillslope processes, leading to increased RELIEF, hillslope length and sediment production. Glacial and coastal erosion similarly respond to uplift and subsidence. Displaced geomorphic features, such as river terraces (see TERRACE, RIVER), shorelines, ALLUVIAL FANs, strata, MORAINEs and PLANATION SURFACEs, serve as markers that are valuable constraints on relative uplift.

Crustal deformation is most commonly associated with faults (see FAULT AND FAULT SCARP). Dislocations occur along lengths of faults during rupture events, producing earthquakes as a side effect. Ruptures that break the surface are

Plate 30 Crustal deformation in alluvium produced locally along the Emerson fault during the 28 June 1992 Landers earthquake in California (M = 7.3). The scarp faces to the south-west and is approximately 1 m high. Its height is locally accentuated by lateral offset of the hilly topography. Dextral offset of ~5 m is evident in the displaced stream course. This photograph was taken several days after the earthquake by Kerry Sieh (California Institute of Technology, USA)

typically tens of kilometres long and involve metres of slip. They are quantified in terms of seismic moment: $M_o = \mu AD$, where μ is rigidity, A is rupture area, and D is average displacement. Earthquake size is thus partly dependent on rupture length, which is controlled by fault zone geometry and segmentation (Plate 30). Coseismic displacement also scales with rupture length, such as the tendency for slip to be ~$10^{-4} - 10^{-5}$ of the length of strike-slip fault ruptures. This scaling is related to the elastic strain the crust adjacent to faults sustains during interseismic periods. The release of accumulated strain provides for moderately regular rupture recurrence. Short-lived faulting events are thus the building blocks by which plate motion translates into long-term deformation (Yeats *et al.* 1997). Over the long term, fault displacements tend to scale as several per cent of the total fault length.

Each of the three main types of plate boundaries consists of faults characterized by certain landforms. Strike-slip faults produced by simple shear involve mainly horizontal displacement and create a minimal degree of topographic disruption. Linear troughs are common along such faults, where weakened fault rocks (see CATACLASIS) are easily eroded by deflected stream courses (e.g. the San Andreas fault). Landforms produced by transpression and transtension at restraining and releasing fault bends include pressure ridges, pull-apart basins (see PULL-APART AND PIGGY-BACK BASIN), and variably faced scarps (scissoring). Strike-slip faults also disrupt geomorphic features horizontally, creating shutter ridges (topographic steps) and deflected or BEHEADED VALLEYS and streams (Sieh and Jahns 1984).

Dip-slip faults involve primarily vertical motion. Normal faults are produced by horizontal extension, where maximum compressive stress is oriented vertically. Resulting fault planes typically dip steeply (~60°). Normal faults juxtapose tilted basement blocks and alluvial valleys in the characteristic basin and range terrain. Vertical separation tends to be asymmetric, with valley subsidence exceeding uplift of basement blocks. Edges of uplifted blocks may preserve FLAT IRONS (triangular facets) related to the fault surface. Mountain fronts typically consist of linear segments interrupted by complex transfer zones, such as the Wasatch front (Machette *et al.* 1992). Parallel normal faults produce down-dropped rift valleys (grabens) (see RIFT VALLEY AND RIFTING) and upthrown blocks (HORSTS).

Reverse or thrust faults are produced by horizontal compression, where the least principal stress is oriented vertically. Thrust fault planes dip shallowly (~30°) and produce irregular mountain fronts that involve wide belts of deformation (Philip and Meghraoui 1983). The degree to which such piedmonts are dominated by erosion, deposition and deformation is represented by numerous geomorphic characteristics, including sinuosity, fan entrenchment and valley geometry. Thrust belts typically involve overlapping arcuate fault segments in parallel series that are connected by secondary structure. These may also involve folding, as typical of foreland fold and thrust belts such as along the Nepal Himalaya (Schelling and Arita 1991). Megathrusts of subduction zones create unique cycles of elastic uplift and subsidence in both hanging wall and footwall, leading to rhythmic perturbation of coastal geomorphology.

Deformation along faults during rupture events can be complex. Fault traces tend to be irregular, such as the characteristic en echelon, anastomosing arrangement of faults within wide (~50 m) ruptures of strike-slip faults (Yeats *et al.* 1997). These shear zones can involve pervasive shearing, although slip tends to concentrate along principal displacement zones. A variety of microgeomorphic features are produced during surface ruptures (see SEISMOTECTONIC GEOMORPHOLOGY). Fault scarps record the vertical separation along faults and portray characteristics linked with fault orientation. Scarp degradation through time occurs predictably by incision and diffusive hillslope creep, such that scarp form is related to scarp age (Avouac and Peltzer 1993). These distinctive landforms record deformation history that can be unravelled using palaeoseismology.

Tectonic strain is also accommodated by FOLDING of rock and sediment, particularly in deep basins. Folding of near surface involves permanent brittle deformation in the form of penetrative intergranular shear or flexural slip between strata. Folds are often associated with blind thrust faults and evolve as faults propagate towards the surface. Fold geometry is closely linked with fault bend and tip geometry. Ongoing deposition around folds can result in piggy-back basins and growth strata that itself becomes folded. Erosion and deposition can also mask the topographic expression of folding in unconsolidated sediment. Processes of diagenetic and pedogenic lithification are thus important for fold preservation. Because strata vary in composition and resistance to erosion, ancient folds can be exhumed by erosion, such as palaeo-folds of the Appalachian Valley and Ridge.

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JAMES A. SPOTILA

CRUSTING OF SOIL

Crusts are thin layers, different in character from the soil beneath, that develop at the interface between the soil and the atmosphere.

One class of crust is often termed inorganic or rain-beat crusts. Large amounts of energy and high transient forces are imparted to the surface of the soil by the impact of raindrops. These break down soil aggregates, compress the surface and dislodge particles. This physical disruption, which may be especially effective where vegetation cover is limited, is aided by certain chemical processes, which include dispersion, which cause further breakdown of soil aggregates. The resulting dense surface layer forms a surface seal, and when this seal dries, a crust forms. Such a crust can have profound effects on runoff and on erosion by wind and water (Poesen and Nearing 1993).

Recently it has been recognized that organic (also called microphytic, microbiotic, cryptogamic or biological) crusts in and on the surfaces of soils play important hydrological and geomorphological roles (Eldridge and Rosentreter 1999). Organic compounds, including plant waxes, can produce hydrophobic (water repellent) substances, as can a range of fungi and soil micro-organisms. Although water repellent soils occur in more humid environments, there are many examples of them that have been reported from semi-arid areas (Doerr *et al.* 2000). These hydrophobic surfaces tend to be zones of reduced soil infiltration capacity and thus of increased overland flow. Following from this is the likelihood that enhanced soil erosion also occurs. Removal of the crusts has been shown to have a dramatic effect on infiltration rates (Eldridge *et al.* 2000).

Likewise biological soil crusts have an influence on aeolian processes. A cover of cyanobacteria, green algae, lichens and mosses is important in stabilizing soils in drylands and thus protects them from wind erosion. They play a role in dune stabilization (Kidron *et al.* 2000). Unlike vascular plants, the cover of organic crusts is not reduced in drought years and they are present the whole year round. However, they are very susceptible to anthropogenic disturbance (Belnap and Gillette 1997). Filamentous cyanobacteria mats are especially effective against wind attack (McKenna-Neuman *et al.* 1996). The filaments and extracellular secretions of cyanobacteria also form water stable aggregates that help soils to resist water erosion and raindrop impact effects (Issa *et al.* 2001). It also needs to be appreciated that not all organic crusts are hydrophobic and that by eliminating the effect of raindrop impact, they prevent the rapid development of a sealed rain-beat crust conducive to runoff generation (Kidron and Yair 1997).

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A.S. GOUDIE

CRYOPLANATION

Cryoplanation (Bryan 1946) is a morphogenetic term introduced to explain and describe low-angled slope surfaces occurring on higher valley-side and summit positions (cryoplanation terraces, or benches), or in valley-side foot positions (cryopediments) in periglacial regions. Cryoplanation and altiplanation are synonymous, and both are forms of equiplanation. Cryoplanation terrace has subsumed several other periglacial terrace terms, including: goletz, altiplanation, NIVATION and equiplanation.

Alleged cryoplanation terraces have flats or treads of 1° to 12° with a sharp inflection in slope at the upslope limit (sometimes called the knickpoint) where risers are often 25° to 35°. Terrace width is often only a few metres, but claims in excess of one kilometre exist (Demek 1969). Sets of cryoplanation terraces produce a staircase effect on a hillside, and their convergence on a summit from two or more sides may produce a summit flat. Both terrace size and frequency appear to increase with time since deglaciation, but terraces may also occur in unglaciated regions. Terrace relationship to permafrost and rock structure is extremely uncertain, but adjustment to rock type is reported. Transport of debris across entire sets of cryoplanation terraces seems essential in some circumstances. This appears to be problematic as lower terraces would have to export all debris from upslope terraces unless it was shed laterally which seems unlikely.

Cryopediments are subject to the same uncertainties associated with tropical pedimentation. A cryopediment is viewed as expanding by headward incision by freeze–thaw weathering, or nivation more broadly. The flat is viewed as a bedrock surface veneered by debris experiencing common periglacial mass wasting processes, e.g. SOLIFLUCTION.

As a process cryoplanation has no unique elements but appears to be synonymous with nivation (Thorn and Hall, 2003) which itself merits more precise articulation. While emphasis has been placed on nivation during the early stage of cryoplanation specifically (Demek 1969), no other specific process (while implied) has ever been invoked for the mature stages. If large perennial snowpatches are protective rather than erosive, as well may be the case, largeness and/or increasingly cold climate may not favour headward expansion. The presence of patterned ground on the tread or transport surface is often, but not always, invoked as an indicator of inactivity.

While the landforms designated cryoplanation terraces or cryopediments are clearly found in periglacial environments, the general absence of sound process research (but see Hall 1997) renders their origins unknown. This problem is exacerbated by the apparent present inactivity of many such features. Nelson (1989) has suggested that cryoplanation terraces may be a periglacial analogue of cirque glaciation reflecting a precipitation/temperature regime unable to sustain full glaciation. Hall (1998) and Thorn and Hall (2003) have suggested that the distinction between cryoplanation and nivation forms and processes needs careful re-examination as it is presently far from clear.

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SEE ALSO: nivation

COLIN E. THORN

CRYOSTATIC PRESSURE

The elevated water potential in saturated, coarse-grained sediments caused by freezing in a closed system. Where the volumetric expansion of water by 9 per cent on becoming ice cannot be accommodated in freezing sediment, pore water is expelled into proximal unfrozen ground, raising the water pressure. Cryostatic pressure is responsible for the uplift of closed-system pingos, beneath which pressures of up to 0.4 MPa have been measured (Mackay 1977). In fine-grained soils, cryostatic pressure may develop at the beginning of laboratory freezing tests, but it has not been measured under field conditions.

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C.R. BURN

CRYPTOKARST

Cryptokarst is a form of karstification limited to the EPIKARSTIC zone. It is always developed under a cover of superficial formations resulting from deposition (loess, etc.) or weathering (alterite). The quality and the thickness of the superficial formation have a direct influence on cryptokarst activity (Nicod 1994).

The main source of acid in ground water is the surface, with the percolation of humic acid (biological activity) and the gaseous exchanges between atmosphere and rainwater. Thus the epikarstic zone is submitted to intense dissolution (Klimchouk 1995) due to the proximity of the surface. The formation of the cryptokarst is enabled by the layer of superficial formation that distributes the water in a diffuse manner, avoiding the concentration of water with high dissolution potential. The chemical equilibration between water and terranes (Stumm and Morgan 1981) implies that the dissolution capacity will decrease proportionally to the residence time of water in the epikarst zone. To stay in the cryptokarst phase, the karstification has to be aborted before reaching the underlying rocks. It means that there will not be any transmission of aggressive water below the epikarst zone, i.e. no water at all either because there are no fast paths (diaclases) or water is non-aggressive because it has already reached equilibrium with the rocks. Concentration of clayey particles that originate from the weathering of the superficial formations can also lead to the clogging of the bedrock interface. In some circumstances, the superficial formations may be drawn down with the vertical progression of the cryptokarstic front and this may induce surface depressions like dolines. On the other hand, the layer of superficial material protects the cryptokarst from surface mechanical erosion.

The geological and topographical conditions for cryptokarst are found in the chalky Cretaceous formations of the Paris Basin (Rodet 1992), England and Denmark. The chalky basement is slightly tectonized (with a resultant low density of fracturation) and the relief is composed of plateaus separated by DRY VALLEYS. The carbonate components of the chalk are easily dissolved but the argillaceous part remains in place. The argillaceous particles are hardly removed by the horizontal water movement. This causes a reduction of the permeability and of the capacity of the basement to be eroded (Lacroix *et al*. 2002).

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SEE ALSO: chemical weathering; epikarst; ground water; karst; palaeokarst and relict karst

MICHEL LACROIX

CRYPTOVOLCANO

A roughly circular area of greatly disturbed rocks and sediments that is morphologically suggestive of being the result of volcanic activity but does not contain any true volcanic materials. Very often the origin of the features has been a matter of controversy. For example, the Pretoria Salt Pan crater in South Africa and the great Vredefort Dome have in the past sometimes been interpreted as volcanic features, but there is now an accumulation of evidence that they both result from meteorite impact (Reinold *et al.* 1992; Reinold and Coney 2001). Conversely, Upheaval Dome in the Canyonlands National Park, Utah, USA, has variously been attributed to meteoric impact, fluid escape, cryptovolcanic explosion and salt doming, with the last explanation now being favoured (Jackson *et al.* 1998). Some features, including a group of eight structures running in a 200-km straight line across the USA, may be the result of comet or asteroid impact (Rampino and Volk 1996) (see ASTROBLEMES, CRATERS). The cryptovolcanic features discussed above show a great range in size. The Pretoria Salt Pan crater is 1.13 km in diameter, whereas the Vredefort structure is 250–300 km across. The aligned structures in the USA are *c.*3–17 km in diameter.

The problem of establishing the origin of closed depressions and circular structures is made evident when one considers the range of hypotheses that have been put forward to explain the Carolina Bays in the eastern USA (Ross 1987):

1. Spring basins
2. Sand bar dams or drowned valleys
3. Depressions dammed by giant sand ripples
4. Craters of meteor swarm
5. Submarine scour by eddies, currents or undertow
6. Segmentation of lagoons and formation of crescentic keys; original hollows at the foot of marine terraces and between dunes
7. Lakes in sand elongated in direction of maximum wind velocity
8. Solution depression, with wind-drift sand forming the rims
9. Solution depressions, with magnetic highs near bays due to redeposition of iron compounds leached from the basins
10. Basins scoured out by confined gyroscopic eddies
11. Solution basins of artesian springs with lee dunes
12. Fish nests made by giant schools of fish waving their fins in unison over submarine artesian springs
13. Aeolian blowouts
14. Bays are sinks over limestone solution areas streamlined by ground water
15. Oriented lakes of stabilized grassland inter-ridge swales of former beach plains and longitudinal dunefields with some formed from basins in Pleistocene lagoons
16. Black hole striking in Canada (Houston Bay) throwing ice onto coastal plain
17. Cometary fragments exploding above surface, their shock waves creating depressions
18. Drought with subsequent fire in peat bogs followed by aeolian activity.

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A.S. GOUDIE

CUESTA

An asymmetric ridge built of dipping sedimentary rocks of alternating resistance against weathering and erosion, elongated along the strike of strata, is called a cuesta. The steep front slope is opposite to the dip, whereas the gently sloping backslope is more or less parallel to the dip. The top part of the cuesta face and the backslope are built of more resistant strata; less resistant ones are exposed in the lower part of the front scarp.

Because of contrasting slope and lithology, each side of a cuesta is shaped by different sets of processes. Rapid mass movement and gully erosion predominate on the steeper slope, and fluvial incision and slow mass movement operate on the backslope. Hence in the long term a cuesta both retreats and is worn down. There is a number of theories how cuesta ridges form but most emphasize differential fluvial erosion within a monocline, which leaves outcrops of more resistant strata as divides and initial cuestas, which then begin to retreat. Bevelled ridge tops indicate that cuesta have developed from a former plain through river incision.

Cuesta is an example of a structure-controlled and climate-independent landform. Classic cuesta landscapes include the Colorado Plateau in North America, the Paris Basin in France and Southern German Uplands.

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SEE ALSO: caprock; escarpment; mesa; sandstone geomorphology; structural landform

PIOTR MIGOŃ

CURRENT

The hydrodynamics responsible for sediment entrainment and transport, erosion and accretion, and morphological change within the coastal zone, consist of *oscillatory motions* associated with WAVES of various frequencies and forms and *quasi-steady, unidirectional currents*. The currents are forced by: (1) a secondary effect of the waves themselves, i.e. *wave drift* or *wave streaming*; (2) *tides*; (3) *wind* stress; (4) *pressure* and *density* gradients; and (5) a variety of motions resulting from the dissipation of wave energy at, and landward of, depth-controlled breaking (*surf zone*). Here the kinetic energy of the waves is transformed into: (a) increased macro and micro turbulence; (b) *drift currents* associated with secondary progressive or standing waves, generally of lower frequency than the incident waves (e.g. edge waves, leaky waves, etc.); (c) *longshore currents*; (d) *rip currents*; (e) *undertow*; (f) *swash*.

Wave streaming (wave drift)

Stokes in 1847 first recognized that WAVE orbital motions were not closed in the case of small amplitude waves in a perfect non-viscous fluid, even in deep water. The fluid particles have a second-order, wave-averaged, mean *Lagrangian* velocity and thus there is a finite *mass flux* of water. Since horizontal velocities increase slightly with distance above the bed, so that particle motion under the crest is slightly larger than under the trough, conservation of mass causes a stratification of flow (Figure 29a). In shallow water, with greater bed friction, the wave orbits become elliptical and drift velocities increase (~$0.1\,\mathrm{m\,s^{-1}}$). A *Eulerian* measure of mass flux can also be obtained by integrating the horizontal velocities beneath the crest and trough over space and time; the same mass flux is obtained although the vertical distribution is different. For real viscous fluids, and waves in finite depth, Longuet-Higgins (1953) showed that there is a time-averaged, net downward transfer of momentum into the boundary layer at the bed, producing a *Eulerian* streaming in addition to the Stokes drift. Again by conserving mass, a stratified profile of the mean current is obtained (Figure 29b); flow is in the direction of propagation at the bed and a reversal occurs at mid-depth; Klopman (1994) has confirmed this pattern through laboratory experiments. In strongly asymmetric flows over steep slopes, shear stresses within the boundary layer may cause a reversal (upwave) mean current at the bed.

Surf zone currents

Currents in the surf zone interact with the instantaneous wave orbital motions (over the full range of short and long period waves) producing a rather complex time-dependent three-dimensional pattern (Svendsen and Lorenz 1989; Figure 30). This is usually disaggregated into a number of distinct components:

Longshore currents are generated when waves break at an oblique angle to the shoreline, and the alongshore component of the onshore directed

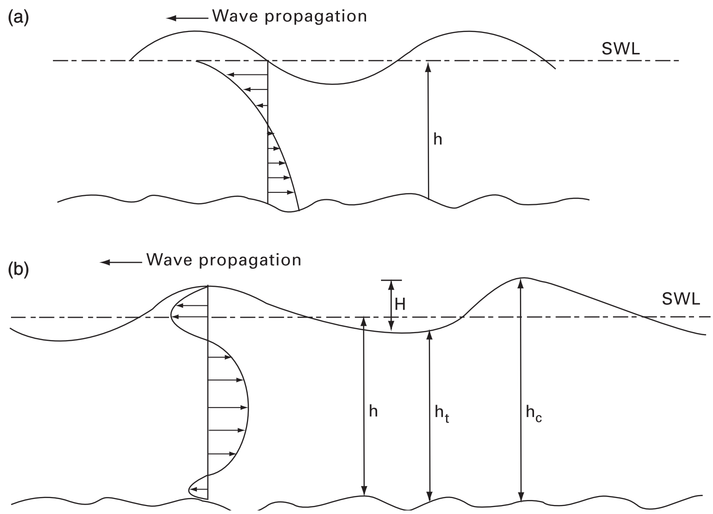

Figure 29 Wave-induced quasi-steady currents: (a) Stokes drift; (b) Longuet-Higgins mass transport

radiation stress (see WAVE) forces a shore-parallel or *longshore current*. Pressure gradients due to water level *set-up* differentials along shore, as well as the shore-parallel component of the onshore *wind stress*, can enhance (or reduce) this forcing. Laboratory and field measurements indicate that the longshore current increases landward from the breakpoint, reaches a maximum around the mid-surf zone and decreases to near zero at the shoreline. Since the radiation stress gradient is a maximum at the breakpoint in the ideal theoretical solution (Longuet-Higgins 1970a,b), a 'smoothing' of the momentum flux across the surf zone, called *lateral mixing*, causes the maximum current to be displaced landward. This mixing also causes *longshore currents* to flow outside the zone of breaking, even though the radiation stress gradients approximate zero. Komar (1998) gives a detailed review of the origins and the spatial and temporal patterns of longshore currents.

Undertow or *near-bed return flow* is a pressure gradient driven, time-averaged, mean current directed seawards near to the bed, everywhere along the shoreline. It is caused by cross-shore differences in the mean water elevation due to wave *set-up* at the shoreline and *set-down* under the breaker zone. *Set-up* and *set-down* result from differences in the local *onshore flux of momentum* by waves (*radiation stress*), which is largest at the breakpoint (where waves are largest) and smallest at the final point of wave dissipation at the shoreline. This gradient forces a displacement of the water from beneath the largest waves towards the shoreline and will be complemented by the onshore *mass flux* of water by the waves, as well as any water moved by *wind stress* acting towards the shoreline. Where nearshore sand bars are present, multiple set-ups and set-downs and associated undertows may be formed by the multiple breaker lines (Greenwood and Osborne 1990). Typically velocities are small, but recordings have been made of undertows up to $0.80\,\mathrm{m\,s^{-1}}$.

Rip currents are discrete, narrow, high velocity *jets* of offshore-directed flow across the surf zone, often forming part of a regular horizontal cellular circulation, with associated shore-parallel to oblique *feeder currents*, and an area of flow expansion, the *rip head*, seaward of wave breaking (Figure 31). *Rips* are often associated with cross-shore oriented depressions (*rip channels*) or breaks in a quasi-shore parallel bar, but are found also on uniformly sloping beaches. Rip currents are not generally *steady state* phenomena, but vary both spatially and temporally.

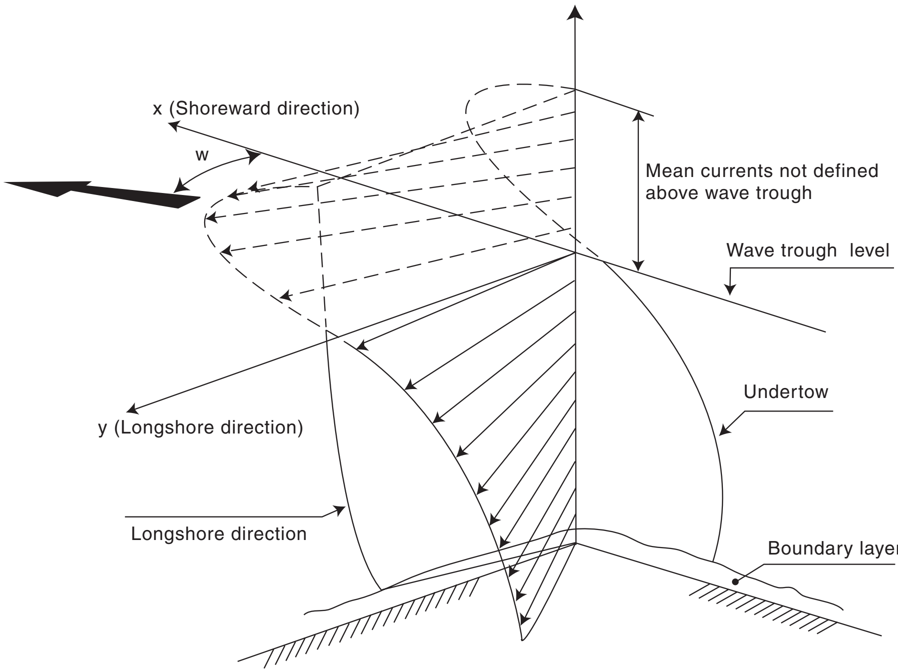

Figure 30 Time-averaged mean velocity vectors in the surf zone (modified after Svendsen and Lorenz 1989)

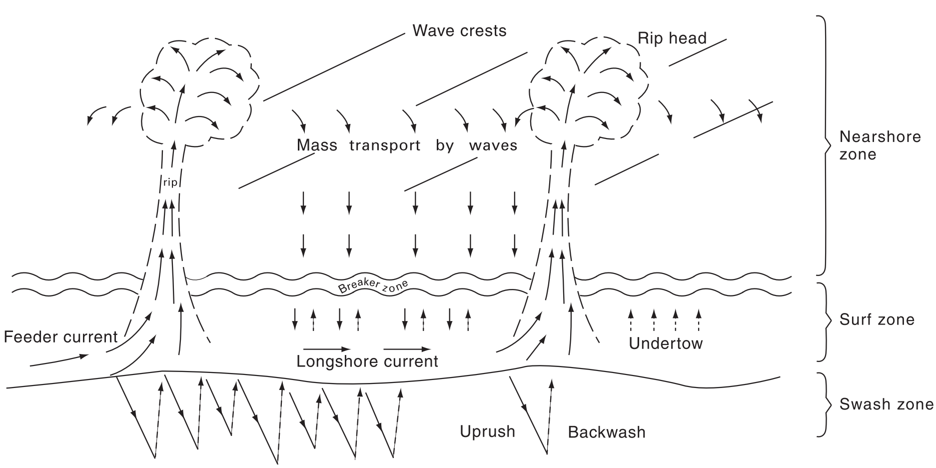

Figure 31 Horizontal cellular circulations in the surf zone

They may be spatially periodic alongshore, with spacing ranging from a few metres (*micro-rips* associated with the embayments of *beach cusps*), to order of 10^2 m (storm cusps), to *mega rips* (Short 1985), which are often single, large-scale, topographically controlled *jets* induced by cellular circulation within the confines of headlands/structures. *Mega rips* may flow offshore for more than one kilometre and can reach speeds $> 3\,\mathrm{m\,s^{-1}}$. Theoretically, rip currents result from a periodic modulation of wave heights and thus *wave set-up* alongshore. This modulation has been related to: (1) rhythmic variations in topography (Sonu 1972; Komar 1971); (2) edge waves (Bowen 1969; Bowen and Inman 1969); (3) interference between two incident wave fields (Dalrymple 1975). Rip currents typically pulse at infragravity wave frequencies (< 0.04 Hz), most likely as a result of the increased *radiation stresses* at the *wave group* frequency. Aagaard *et al*. (1997) measured very long period fluctuations of 5–10 minutes. Current speeds may increase with falling tidal levels, especially where topography increasingly confines the current. Rip currents are capable of transporting large volumes of both coarse bedload (especially in migrating megaripples, e.g. Gruszczynski *et al*. 1993) and suspended load from within the surf zone to seaward. Mega rips may well transport nearshore sediments onto the continental shelf below the level of average *wave base* (depth limit of surface waves).

Swash is the *uprush* and *backwash* currents on the beach face and reflects the ultimate dissipation of incident wave energy. Both currents are turbulent, with the former assuming a flow direction coincident with the angle of approach of the breaking wave; the backwash results from gravity acting on the water on the beach and thus flows down the maximum beach slope. This often results in a 'zig-zag' motion of water and sediment. Swash currents depend on the nature of the incoming waves, the beach face slope and the state of the beach water table. If the beach is not saturated there will be a tendency for infiltration of the uprush into the permeable beach face, and thus a reduction in the amount of water in the backwash. Because water can drain from the beach face water table, the backwash tends to last longer and is usually thinner and flows may be supercritical, with hydraulic jumps common. Van Rijn (1998) and Butt and Russell (2000) provide reviews of *swash* hydrodynamics and sediment transport.

Tidal currents or tidal streams

Currents that result from the tidal wave forced by the gravitational tractive stresses generated by the moon and sun are called *tidal currents* or *tidal streams* and reverse their direction either semi-diurnally or diurnally. They may reach speeds up to $6\,\mathrm{m\,s^{-1}}$ in coastal waters if they are constrained topographically, but are generally much smaller (~$0.05\,\mathrm{m\,s^{-1}}$) and are dominated on open coasts by gravity WAVE oscillations. Tidal currents vary in magnitude in response to the local tidal range and will have a variable phase relationship with tidal elevation depending upon whether the tidal wave is *standing* (maximum flows at mid-tide) or *progressive* (maximum flows at high and low tide). The current direction will be constrained by Coriolis and thus currents generated by the rising (flood) tide will follow a different path than those of the falling (ebb) tide, giving an ellipsoidal pattern of currents. In estuaries, for example, distinct *flood* and *ebb channels* may exist. This creates *residual currents* that may be significant in terms of sediment transport and deposition. In all cases tidal currents vary with depth as a result of bottom friction and a logarithmic velocity profile develops. Tidal currents are most significant in estuaries, inlets and straits and in some sections of continental shelf. Davis and Hayes (1984) discuss the relative role of tides and waves in the development of coastal morphologies.

Other currents in the coastal ocean

Wind-induced currents are formed when wind shear on the surface is transferred into the water column. At the shoreline this will induce a mass transport in the direction of the wind and can be resolved into a shore-normal and shore-parallel component. The former results in an elevation of the mean sea level (*wind set-up*) at the shoreline, which is an addition to the *wave set-up* which causes *undertows*; the latter will enhance the radiation stress driven *longshore current*. However, such currents are often short-lived, as local winds are subject to frequent change in speed and direction. Gradient in *wind set-up* can also generate currents at the scale of the complete coastal ocean boundary layer during large storm events (e.g. hurricanes, intense mid-latitude cyclones). This results in large offshore directed pressure-gradient flows, whose speed and direction are constrained by frictional forces and the Coriolis effect (Swift 1976). In deep water and where wind systems are of long duration (e.g. the Trade Winds, Equatorial Winds, other zonal winds, etc.) they cause large coastal circulation systems (e.g. upwelling and downwelling systems, the Equatorial currents, etc.). The rotational force of the moving Earth (Coriolis) also influences such large-scale flows. Currents tend to move at 45 degrees to the wind at the surface and rotate clockwise (or anticlockwise depending on the hemisphere) with depth, to flow in the opposite direction to the surface wind; this is the *Ekman Spiral*.

Density-induced currents result from density differences due to differences in temperature, salinity or sediment mass concentration. Such gradients force both horizontal and vertical currents, which are also affected by the rotational effect of the Earth. The component of flow driven by the slope of an internal density surface is called the *baroclinic* component; the component driven by the slope of the sea surface is the *barotrophic* component.

Inertial currents are residual currents in large bodies of water, which continue to flow under their own momentum long after the original forcing has ceased.

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BRIAN GREENWOOD

CUSPATE FORELAND

Cuspate forelands are large-scale, tooth-shaped coastal promontories. Although erosion plays a part in their evolution and their form, they are basically landforms of accretion, composed of sorted beach sand deposited from littoral transport (see LONGSHORE (LITTORAL) DRIFT). They often enclose a lagoon (see LAGOON, COASTAL) or marsh. There are two basic types.

Recurved cuspate forelands

At sites where the coastline changes direction abruptly landward, littoral transport slows and deposition occurs, creating over time, a broad elongated shoal. This feature, termed 'spit platform' by Meistrell (1966), is the foundation on which the emergent cuspate foreland grows. The original elongated form is sometimes referred to as a 'flying spit', 'spit with recurves' or 'fleche', in that its growth is in a direction continuous with that of the updrift coast, 'flying' offshore into deeper water. On the leeward side, sand deposits washed over during storms or transported around the tip are subjected to wave action from the opposite direction. The result is a series of concave-seaward recurves, or secondary spits, extending at an acute angle from the tip to the downstream coast. Because of the effect of this bi-directional wave climate, the foreland may range in form from symmetrically cuspate, when the wave effect is fairly balanced on both sides, to asymmetrical and elongated, if wave effect on one side predominates. Examples of this type are Cape Canaveral in Florida, Pointe de la Coubre near the Gironde estuary in western France, and the Toronto Islands of Lake Ontario, Canada.

Dungeness-type cuspate forelands

This is the term originally given by Gulliver (1895) and Johnson (1919) and elaborated by Zenkovitch (1967), to refer to symmetric, accretionary forelands that grow at high angles to the shore. Coakley (1976) demonstrated that they usually form at the site of a pre-existing morphological feature that is transverse to the coastline, e.g. a recessional moraine, or low bedrock ridge. This disruption of the coastal orientation causes the accumulation of the spit platform. These forelands are dynamic and are influenced by periodic reversals in littoral drift direction due to changes in the wind/wave climate. Thus, the foreland may be nourished, and be eroded, from both sides. This results in the classic pointed cuspate form with a well-developed complex of beach ridges. The evolution of the foreland may be studied through the pattern of the preserved beach ridges. Good examples are Dungeness on the south-eastern English coast and Point Pelee, Lake Erie, Canada.

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SEE ALSO: beach cusp; tombolo

JOHN P. COAKLEY

CUT-AND-FILL

Cut- (or scour) and-fill is the local cyclic erosion and deposition of sediment in a river channel, usually over short time periods (hours to years). It occurs as part of the process of sediment transport and development of channel morphology and consequently is associated with spatial and/or temporal changes in flow conditions, such as the passage of a flood wave along a channel. Cut-and-fill is distinct from progressive changes in channel elevation over longer time spans and greater distances that are usually referred to as degradation (erosion) and aggradation (deposition) and may produce substantial accumulation of ALLUVIUM and formation of terraces (see TERRACE, RIVER).

Cut-and-fill occurs in alluvial stream channels whenever bed sediment is moved. It is the result of variation in channel topography related to the normal processes of channel development, sediment transport and the response to, and recovery from, events such as large floods. Consequently cut-and-fill occurs for a variety of reasons including changes in flow hydraulics and sediment transport rate along the stream or during a flood event, development and migration of BEDFORMS, and changes in channel pattern, position or overall morphology. Cut-and-fill related to changes in channel morphology or channel migration is well known from studies of braided streams (see BRAIDED RIVERS) and is associated with the formation and migration of scour pools and bars (see BAR, RIVER) and with channel migration and AVULSION.

A cycle of cut-and-fill may occur at a single cross section during a flood, sometimes associated with rising and falling stages of the hydrograph. For example, in a POOL AND RIFFLE channel, cutting followed by filling may occur in pools while the reverse occurs in riffles because of changes in velocity and bed shear stress as discharge rises and falls. In other cases there are distinct areas of the channel in which only cut or fill occurs during a flood event or over longer time periods. Generally, there is compensating cut-and-fill within a channel reach so that the quantity of erosion at one location is matched by deposition nearby, sediment may be transferred from one to the other and this mass conservation means that there is no overall change in channel elevation (Colby 1964; Ashmore and Church 1998; Eaton and Lapointe 2001).

Observations in large SAND-BED RIVERS have shown cut, and subsequent fill, of the order of two or three metres at particular river cross sections during a single flow event (Colby 1964). In small, GRAVEL-BED RIVERS and streams, measurements indicate that the average depth of cut-and-fill during sediment transport events is of the order of about twice the maximum grain size, but local depths may be much greater than this (Hassan 1990; Haschenburger 1999). The average and maximum depth of cut or fill in a particular stream tends to be greater at higher discharges (greater bed shear stress) as does the area of the channel experiencing cut or fill. Where cut-and-fill is related to the development and migration of bedforms and scour pools the depth of activity is determined by the vertical amplitude of the channel topography.

Common methods for measurement of cut-and-fill are depth sounding, survey of topographic changes over an area of channel, and deployment of scour chains or tracers. Sounding provides very high temporal resolution but may be limited in spatial coverage while repeated surveying provides detailed information on the spatial pattern but may underestimate cut-and-fill amounts and rates if there is both erosion and deposition at a given location between surveys. Scour chains can provide both spatial patterns and also some information about the alternation of cut-and-fill at a point during a flow event. Scour chains are inserted vertically into the stream bed so that the increase in length of chain exposed at the bed after a flow event indicates the depth of cutting, while the depth of fill can be inferred from the depth of burial of the vertical section of chain.

Cut-and-fill is fundamentally and practically significant. Fundamentally, it is the result of the direct connection between channel morphology and sediment transport – spatial and temporal variation in transport rate leads to cut-and-fill and therefore change in channel morphology. Furthermore, the rate of transport of bed sediment during a transport event can be defined as the average depth of cut or fill multiplied by the average velocity of the sediment particles (distance moved divided by the duration of the transport event), which is one method for estimating bed sediment transport rate (Ashmore and Church 1998; Haschenburger 1999). Because cut-and-fill is a significant aspect of stream channel dynamics it is also important in a number of other contexts such as sedimentological interpretation of alluvial deposits (Best and Ashworth 1997), engineering design of river structures, and anticipation of the effects of direct or indirect modification of river channels on channel dynamics and stream habitat.

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PETER ASHMORE

CYCLE OF EROSION

The Cycle of Erosion or 'The Geographical Cycle' was formulated in the latter years of the nineteenth century by W.M. Davis (e.g. Davis 1899). It was the first widely accepted modern theory of landscape evolution (see SLOPE, EVOLUTION). Davis regarded landscapes as evolving through a progressive sequence of stages, each of which exhibited similar landforms. In the Davisian model it was assumed that uplift takes place quickly. The land is then gradually worn down by the operation of geomorphological processes, without further complications being produced by tectonic movements. It was believed that slopes declined in steepness through time until an extensive flat region was produced close to BASE LEVEL, though locally hills called *monadnocks* might rise above it. This erosion surface was termed a *peneplain*. The reduction in the landscape creates a time sequence of landforms progressing through three stages: youth, maturity and old age.

Initially the Davisian model was postulated in the context of development under humid temperate ('normal') conditions, but it was then extended to other landscapes including arid (Davis 1905), glacial (Davis 1900), coastal (Johnson 1919), karst (Cvijić 1918) and periglacial landscapes (Peltier 1950).

Davis's model was immensely influential and dominated much of thinking in Anglo-Saxon geomorphology in the first half of the twentieth century, contributing to the development of DENUDATION CHRONOLOGY. Davis was a veritable 'Everest' among geomorphologists (Chorley *et al*. 1973). The model was largely deductive and theoretical and suffered from a rather vague understanding of surface processes, from a paucity of data on rates of operation of processes, from a neglect of climate change, and from assumptions he made about the rates and occurrence of uplift. However, it was elegant, simple and tied in with broad, evolutionary concerns in science at the time. Nonetheless, by the mid-1960s the concept was under attack (Chorley 1965).

The Davisian model was never universally accepted in Europe, where the views of W. Penck were more widely adopted. Penck's model involves more complex tectonic changes than that of Davis, and regards slopes as evolving in a different manner (slope replacement rather than slope decline) through time (Penck 1953). An alternative model of slope development by parallel retreat leading to *pediplanation* was put forward by L.C. King (e.g. King 1957). Thorn (1988) provides a comparative analysis of the models of Davis, Penck and King. Another evolutionary model of landscape evolution was produced by Büdel (1982), who developed the concept of ETCHING, ETCHPLAIN AND ETCHPLANATION.

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A.S. GOUDIE

CYCLIC TIME

A cycle is a period of time in which events happen in an orderly way. The order repeats itself in time so that there is a recurring series of changes. The term 'cyclic time' is unnecessary and illustrates the confusion that exists in the use of cyclic concepts.

Unfortunately a geomorphological cycle is often regarded as a sequence of *changes* from an initial state, through a series of stages to an ultimate state. In such models it is assumed that the changes taking place are such that the system has a different configuration when observed at different times, in other words, landforms have an observable history.

This emphasis meant that attention was paid to the sequence of changes and the 'stages of evolution' rather than the temporal lengths, frequencies and durations of the cycle and its events. This led to an emphasis on DENUDATION CHRONOLOGY (establishing and dating the stages of change) rather than a real understanding of geomorphological processes, rates of change and event statistics.

The interchangeable use of 'time' as the time during which changes take place, as the sequence of changes or the stage reached in the 'cycle' began with W.M.Davis (1899, 1905). In some passages he uses the correct dictionary sense. After describing the way a river advances through its long life and reduces an uplifted landmass to a PENEPLAIN he states, 'This lapse of time will be called a cycle in the life of a river.'

Unfortunately, he also described the geographical cycle as 'a complete sequence of landforms' but then qualified this as taking place from the uplift (an event) that produced the initial form through a sequence of form changes (responses to process events) to an ultimate form – a plain of low relief. In another passage Davis said that a geographical cycle 'may be divided into parts of unequal duration, each part of which will be characterized by the degree and variety of relief, and by the rate of change that has been accomplished since the start of the cycle'. Davis described how the successive forms of the cycle were dependent on three variable quantities: structure, process and time. He therefore makes it clear that 'time' is the amount of change from the initial form or its stage of development. In other passages the amount of change is 'a function of time' and time is again used as one of the trio of controls.

The period of time involved in a cycle has been poorly thought out. Davis (1899, 1905) estimated that the block mountains of Utah would be peneplained in 20–200 Ma. Wooldridge (personal, communication 1960) estimated up to 100 Ma but stated that the Mio-Pliocene peneplain had been produced in less than 20 Ma (Wooldridge and Linton 1955). Schumm and Lichty (1965) thought in terms of 10^6 years and Schumm (1963) pointed out that the time period to base levelling would be greatly extended by isostatic, erosional rebound. The general conclusion is that denudation cycles involve time spans of geological duration for their completion and recurrence by further uplift.

It is now known that the controls of Earth systems, such as structure, climate and base level, do not remain stable for such long periods of time and that it is preferable to establish the time periods for the frequency of landform creation events, the relaxation times and the landform survival times for the relevant system specifications. Some geomorphologists would argue (Schumm and Lichty 1965) that time can be divided into cyclic, graded and steady time periods. A more recent view (Graf 1977; Brunsden and Thornes 1979; Brunsden 1990) would suggest that the period 'cycle' be dropped in favour of system-based terms.

The name of a cycle (cyclic time?) is taken from the subject matter of the changes involved. General examples are a geographical cycle, a geomorphic cycle, an erosion cycle, a cycle of topographic development, a cycle of denudation, a cycle of life (Davis 1899). More specific uses were the normal cycle (landscapes developed under humid temperate conditions), shoreline development, sedimentation, karst, slope evolution, underground drainage, hydrologic, climatic and cycles of all geomorphological processes regimes.

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DENYS BRUNSDEN

CYMATOGENY

A term introduced by L.C. King (1959) to describe crustal movements intermediate between EPEIROGENY and OROGENESIS. They involve a warping of the Earth's crusts over horizontal distances that range from tens to hundreds of kilometres, and with vertical movements up to thousands of metres. They involve, however, minimal rock deformation. It is thought that the uplift is caused by processes active within the Earth's mantle.

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A.S. GOUDIE

D

DAM

Dams have been used to secure water supplies, to control floods and to generate power for more than a thousand years. The earliest civilizations developed along rivers in arid and semi-arid areas, such as along the Nile, and it is here that the oldest dams were built about 5,000 years ago. Flows in the wet season were stored in reservoirs to supply water for large-scale irrigation agriculture during the dry season. Water security and food security were closely linked and maintained the social, economic and political stability of the developing civilizations.

Today, the flows on most rivers are controlled to some degree by dams (WCD 2000). There are more than 45,000 dams over 15 m high and the largest dams stand more than 200 m high! The first big dam was the 221 m high Hoover Dam on the Colorado River, constructed in 1935. Kariba dam on the Zambezi, closed in 1958, was the first large dam to be constructed in the tropics. Water stored in reservoirs exceeds that stored in natural lakes by more than three times. Major rivers such as the Colorado and Columbia in USA, the Volga in Europe, the Nile in Africa, the Parana in Latin America, and the Murray–Darling in Australia have been intensively developed. Hydro-electric power is a major driver of dam building. Only about 3 per cent of the world's total energy consumption is supplied by water power and some 75 per cent of the hydroelectric power potential of the world's rivers is still to be exploited.

The geomorphological significance of large dams includes reservoir-induced earthquakes that have occurred in a small proportion of cases but have dramatic impacts. Large dams and reservoirs can both increase the frequency of earthquakes in areas prone to seismic activity and cause earthquakes in areas thought to be geologically stable. The mechanism involves the extra water pressure created by the dam and reservoir within faults in underlying rocks. Gupta (1992) records seventy examples of reservoir-induced seismicity. In many cases, the strongest shocks, often exceeding 4 and occasionally 6 on the Richter scale, occurred shortly after the initial filling of the reservoir.

Much more common are the impacts of dams on the fundamental fluvial processes, the flow and sediment transport regimes. These process changes induce adjustments of the size and shape of the river channel, and the form of the floodplain. These changes of the flooding and sedimentation regimes together with the changes to the morphology of the river corridor impact upon plants and wildlife by changing the habitats available for biota.

All dams are designed to capture floodwaters (see FLOOD) and represent perhaps the greatest point-source of hydrological impact. On some rivers reduced flood magnitudes have been experienced for more than 1,000 km below the dam and below the Aswan Dam on the River Nile, the reduction in freshwater flows is seen in the increased salinity of waters offshore of the delta in the south-east Mediterranean Sea. The Colorado River, USA, is dammed along its length, as are its major tributaries, and less than 1 per cent of the virgin flow reaches the river mouth. On the Murray–Darling system in Australia, which is regulated by nine principal storage reservoirs, the natural flow pattern was reversed, with high flows being released from the dams to supply irrigation demands downstream.

The basic concept of flood storage is 'empty space', keeping a reservoir as empty as possible to store floodwaters when they arrive. Water-supply reservoirs need to be kept full to provide water for domestic, industrial or irrigation supplies during dry seasons and dry years. But even when a reservoir is full and spilling over the dam, the flood peak downstream will not be as high as that for the inflow because of temporary storage in the lake as levels rise above the crest level of the overflow weir. Commonly, the size of the mean annual flood below dams has been reduced by between 25 and 50 per cent.

Dams and reservoirs also trap the sediments transported by a river – in many cases permanently storing the entire sediment load supplied by the upstream drainage basin. As the relatively high-velocity, and turbulent, water of a river feeding the reservoir is transferred into the slow-flowing water within the lake the sediment is deposited. Part is deposited in the reservoir itself and part in the channel and valley-bottom upstream, as a result of the backwater effects from the reservoir reducing velocities of river and floodplain flows. The coarser sediments settle out to form a delta. The finer particles, especially the clays, are distributed further out into the lake. Average annual rates of reservoir storage loss are usually less than 0.5 per cent per year but exceptional rates of more than 2 per cent per year have been reported from regions with high SEDIMENT LOAD AND YIELDs. One extreme case is the Heosonghi Reservoir on the Huang Ho, China that lost nearly 20 per cent of its storage capacity within three years of completion.

Flows released from dams or passing the spillway during floods are known as 'clearwater' releases because they are more or less sediment free. However, sometimes the water can appear turbid, not because of suspended sediments, but because of high concentrations of plankton when water is released from the lake surface during summer. This is caused by phytoplankton – algae and diatoms – which can reach high concentrations in relatively warm, surface layers of reservoirs having long retention times. Turbid releases may also be caused by the discharge of deep water during the autumn when stratified lakes mix – the 'overturn'. Such discharges can contain high concentrations of iron, manganese and hydrogen sulphide, giving a bad egg smell. However, in both cases, the quality of the water discharged from a reservoir can be controlled by the selective release of water from different depths within the lake. Occasionally, sediments are deliberately flushed from reservoirs by opening deep valves in the dam, to reduce the rate of storage loss. An example of this operation is the management of the Verbois, Chancy–Pougny and Genissiat reservoirs on the River Rhone in France. During these rare events, suspended sediment concentrations can exceed 1 g l^{-1} but such sudden surges of sediment-laden water can cause problems for water quality downstream.

Clearwater releases and the regulated flow regime below dams induce changes of channel morphology. The size and shape of natural river channels are in regime with the flows and sediment loads. Below dams two general types of change in regime can occur, although in detail there are many variations on these (Brandt 2000). The first type of channel change occurs where the dominant change of fluvial process is the reduction in sediment load. The clearwater releases of sediment-free water from reservoirs into channels with alluvial bed and banks can cause rapid erosion, or degradation, that may extend for many kilometres downstream. Typically bed degradation deepens the channel and the banks may also be undermined and sand and gravel bars eroded. An increase in the size of the sediments on the channel bed, which becomes armoured by the selective removal of the finer particles, and the reduction in channel slope as a result of bed incision may limit the amount of bed erosion. The result is a channel of increased cross-sectional area. Reports of degradation rates of more than 100 mm per year over channel lengths of more than 100 km are not uncommon. Rates decline over time until a new 'regime' condition is reached.

The second type of channel response is to the regulated flows, especially the lower flood levels. This induces a reduction of channel capacity most commonly observed as a reduction of channel width. Flow regulation reduces the capacity of a river to transport sediments supplied by sources downstream from the dam. These sources include tributary catchments and any degrading reaches and the dam and reservoir site during construction. Coarse sediments will be deposited on the channel bed but sediments will also accumulate as bars and benches along the channel margin, sometimes creating a new floodplain. The former floodplain is then converted into a river terrace (see TERRACE, RIVER).

The rate of channel narrowing is highly variable but can be particularly rapid in two situations. First, channel change is often rapid along regulated rivers in semi-arid areas where wide, braided rivers are converted into single channels. In these cases, the growth of vegetation such as willows and poplars, sometimes accelerated by the maintenance of higher baseflows than in the natural river, can result in dramatic reductions of channel width (Merritt and Cooper 2000). The second situation is downstream from tributaries that produce high sediment delivery to the regulated channel. Sometimes, the reduced flood levels within the regulated river can accelerate erosion within the tributary increasing sediment supplies until the tributary has reached regime (Germanovski and Ritter 1988).

Each river comprises a sequence of channel reaches, each having a different channel form reflecting the history of the reach over Quaternary, historical and recent timescales. Channel change involves the movement – erosion, transport and deposition – of large volumes of sediments into and through this series of channel reaches over periods of time ranging from years to centuries. Volumes of up to 1 million cubic metres in a one-kilometre reach are not uncommon. In many cases individual reaches of river channel will show a COMPLEX RESPONSE to impoundment (Sherrard and Erskine 1991; Church 1995). This involves alternating phases of degradation and aggradation as the river network, the main channel and its tributaries, continue to adjust to the regulated flow regime by moving sediment through the sequence of reaches until a new 'regime' channel form is established.

Along rivers that have low sediment loads and stable, cohesive bank materials, adjustments of channel form may be very slow. In extreme cases the timescale for channel change to establish a new 'regime' channel may extend to hundreds of years. In these cases, the existing channel form will accommodate the regulated flows and evidence of upstream impoundment may be limited to local sediment accumulation in pools and backwaters, the growth of moss on large stones, and the marginal growth of emergent aquatic plants. An extreme flood may be required to initiate major channel changes in these reaches.

Geomorphology provides a physical template for river, riparian and floodplain ecology (Petts 2000) (see PHYSICAL INTEGRITY OF RIVERS). Variable river flows and sediment loads, and dynamic channels that change position by the processes of deposition and erosion, creating new floodplain patches and eroding others, sustain a diverse and highly productive riverine ecosystem. The channel pattern (see CHANNEL, ALLUVIAL) determines the range of habitat types found along any river but the frequency of erosion and deposition determine the level of disturbance that rejuvenates ecological successions.

Dams reduce the physical dynamism of the downstream riverine ecosystem, simplifying the physical habitat, and reducing both biological diversity and productivity (Ward and Stanford 1995). Advances in the application of geomorphological knowledge to the operational management of regulated rivers through the development of instream flow models (Petts and Maddock 1994) seek to sustain the ecological integrity of rivers below dams. Such models determine three levels of flows that need to be sustained along a regulated river to maintain the physical and, therefore, ecological dynamism of the river corridor. These flows are the floodplain maintenance flow, the channel maintenance flow (usually the BANKFULL DISCHARGE), and flushing flows to prevent the siltation of the channel bed and to prevent vegetation encroachment into the channel.

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The journal *River Research and Applications* (until 2002 *Regulated Rivers*) has a focus on rivers below dams.

GEOFFREY PETTS

Plate 31 A broad, flat-floored, grassy dambo in west central Zambia

DAMBO

A headwater valley in areas of low relief, particularly in the seasonal tropics, that is channelless and in humid areas may contain swamps. Dambos are also known as *vleis* in southern Africa, *matoro* in Zimbabwe, *baixas* in Amazonia, *bolis* in Sierra Leone, *mbuga* in East Africa and *fadama* in northern Nigeria. German geomorphologists (e.g. Büdel 1982) have called them 'Spülmulden' or *wash depressions*. True dambos tend to be restricted to climates with present-day rainfalls between 600 and 1,500 mm, but the *bolis* of Sierra Leone are found where annual rainfall approaches 2,500 mm. They are also probably best developed on ancient planation surfaces. They occur on a wide range of rock types from unconsolidated Kalahari Sand through to shales, quartzites, schists, gneisses and granites (Thomas and Goudie 1985, Plate 31).

Their hydrology has been described by Bullock (1992), and they are a major source of water supply in rural areas in countries like Zimbabwe. Many of them are now being exploited for agricultural reasons and are suffering degradation, including gullying, as a consequence. Indeed, dambo is a Bantu word meaning 'meadow grazing,' for they are often grass covered and have no true woodland vegetation (Mäckel 1974).

Dambos tend to have low gradients (usually less than 2°). They receive their water either from direct precipitation onto the dambo or by subsurface flow from the surrounding high ground. With regard to the processes that lead to their formation, two main schools of thought exist (Boast 1990). The fluvial school envisages dambos as the simple extensions of the channelled drainage network. Rivers erode their head valleys which may subsequently be infilled by slope colluviation and by channel alluviation. Sheet-wash processes under seasonal rainfall regimes may be especially important. The other school of thought advocates differential chemical and biochemical corrosion or sapping rather than mechanical erosion as the main process. It sees dambo morphology as breaking 'too many fluvial rules' to be explicable in simple fluvial terms. That fluvial processes have operated in some dambos is made clear by the stratigraphy of their floors, which can reveal old alluvial fills. It is evident in many parts of central Africa that the balance between colluviation and alluviation has varied repeatedly in response to climatic changes. However, the two schools of thought are not necessarily mutually exclusive and Thomas (1994: 279) believes that 'Opposition between sapping (or etching) processes and sedimentation in dambos is misplaced.'

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A.S. GOUDIE

DATING METHODS

Stratigraphic relationships between landforms or within depositional sequences provide the most common, and simplest, means of deducing the age. Other than under exceptional circumstances, younger landscape features or sediments overlie older ones. However, this approach does not enable rates of processes to be deduced, nor give any idea of the relative or absolute timing of events. A number of dating methods exist based on chemical and biological changes that occur through time. The formation of CHEMICAL WEATHERING rinds and DESERT VARNISH on exposed rock are examples of the former, while amino-acid racemization and LICHENOMETRY are examples of the latter. These are all relative dating methods (i.e. indicating that one landform is approximately twice as old, or three times as old, as another). Another class of dating methods is that based on the correlation of events. For example, periodically the Earth's magnetic field is reversed, with the positions of the north and south magnetic poles switching. The last time that such a reversal occurred was 780,000 years ago. This event is recorded in a number of sedimentary and volcanic records and provides a synchronous marker across the globe, thus allowing one site to be correlated to another. In order to define the numerical age of this event (i.e. the age expressed as the number of years before present) a different class of dating methods are required – absolute age methods.

The discovery of radioactivity at the end of the nineteenth century provided the foundation for a suite of absolute dating techniques collectively known as radioisotopic methods. These all rely upon the fact that the rate at which a radioactive isotope of an element undergoes decay to produce another isotope (known as the daughter product) is constant, unaffected by any external controls such as temperature or pressure.

Radiocarbon dating was the first radioisotopic dating method to be widely applied starting in the 1950s. Carbon occurs as three isotopes, ^{12}C, ^{13}C and ^{14}C. The first two are stable isotopes, while the latter is radioactive, but all react chemically in identical ways. Radiocarbon (^{14}C) is generated in the upper atmosphere by the interaction of high energy cosmic rays with nitrogen atoms. The ^{14}C generated in this way is rapidly oxidized to form carbon dioxide which enters the carbon cycle. Radiocarbon has a half-life (the time taken for half of the atoms of ^{14}C within a sample to undergo radioactive decay) of $5{,}730 \pm 40$ years, and the concentration of ^{14}C in the atmosphere is a balance between the rate of production and the rate of decay. All living things exchange carbon with some part of the carbon cycle, and thus contain ^{14}C. After death this exchange ceases. The ^{14}C continues to decay according to its half-life, but it is no longer replaced by exchange with any part of the carbon cycle. Measurement of the ^{14}C remaining in a sample allows calculation of the period of time since death. Radiocarbon dating is most appropriate for organic materials, but can also be applied to some carbonates. The method assumes that the concentration of ^{14}C in the various reservoirs of the carbon cycle has remained constant through time. Measurement of the ^{14}C activity of tree rings of known age for the last 11,000 years shows this not to be the case, but these results allow ^{14}C ages to be calibrated to calendar years (Aitken 1990: 98). Between 11,000 years and ~40,000 years, the limit of the method, the ^{14}C calibration is less well known and the uncertainties on the ages larger.

In addition to ^{14}C, a wide variety of other isotopes (^{10}Be, ^{26}Al, ^{36}Cl) are generated both in the atmosphere and at the surface of the Earth by the interaction of cosmic rays. A suite of dating methods based on these cosmogenic isotopes have recently been developed (see COSMOGENIC DATING).

Other radioisotopic methods rely upon the very long half-lives of certain isotopes. Uranium occurs naturally as several isotopes (^{234}U, ^{235}U, ^{238}U). ^{238}U has a half-life of 4.47×10^9 years, comparable with the age of the Earth, and thus a significant quantity persists in the natural environment. Unlike ^{14}C, whose daughter product (^{14}N) is stable, the decay of ^{238}U produces ^{234}Th, which is itself radioactive. This in turn decays to produce ^{234}Pa, which decays to produce ^{234}U, ^{230}Th, ^{226}Ra and so on, producing a decay series until a stable isotope, ^{206}Pb, is produced (Figure 32). Over time the concentration of the different isotopes within the decay chain will alter until a state is reached where the number of decays per unit time from each isotope is identical – this state is termed secular equilibrium. The different chemical characteristics of the elements within the decay series provide a number of radioisotopic dating methods. For instance, when calcite is deposited in KARST environments, trace quantities of uranium are also deposited.

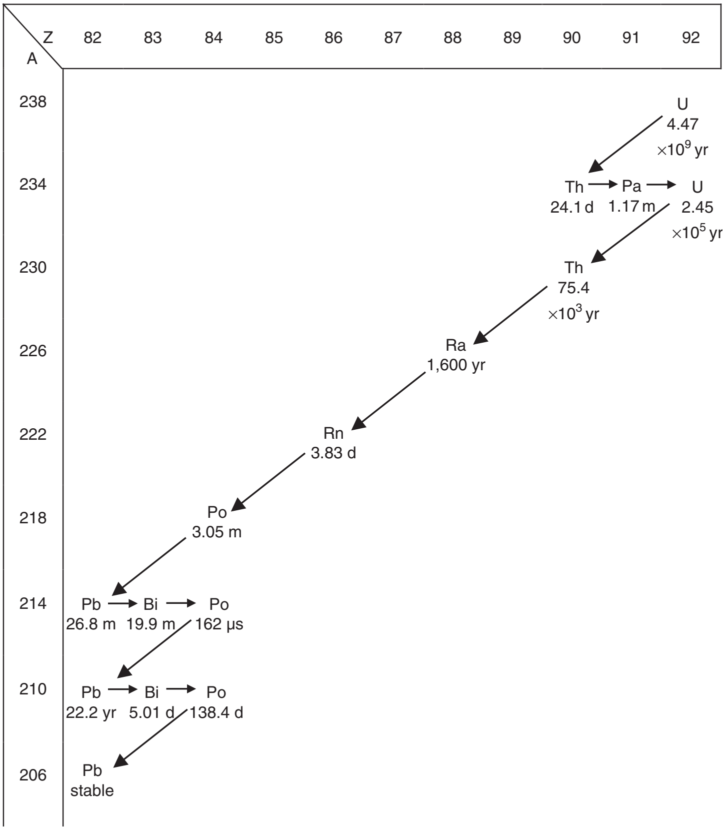

Figure 32 Decay series for ^{238}U. The half-life of each isotope is shown under the isotope. A is the atomic mass, while Z is atomic number

However, little or no thorium is deposited because it is relatively insoluble. Thus within the calcite, uranium will occur but without its thorium daughter products – it is said to be daughter-deficient. Over time, as the uranium undergoes decay, the concentration of thorium will increase. At the time of deposition, the $^{230}Th/^{234}U$ ratio will be zero, and will increase in a predictable manner allowing the age of formation of the calcite to be determined. This process can be used to date the precipitation of calcites over the last 350,000 years. As well as calcite in karst environments, another excellent target for Th/U dating is coral (Muhs 2002). Another part of the uranium decay series ^{210}Pb has a much shorter half-life (22 years) and can be used to provide ages over the last 100 years. In this case, the method is most commonly applied to lake sediments and nearshore marine basins (Appleby and Oldfield 1992). The method relies on the fact that one of the isotopes within the ^{238}U decay series, ^{222}Rn (radon), is a gas. This escapes to the atmosphere where it will undergo decay via a series of short-lived daughter products to produce ^{210}Pb. This falls from the atmosphere, producing a near constant supply to the surface of lakes and the nearshore. This ^{210}Pb is incorporated into the sediment accumulating under the water body, but none of its parent isotopes are present – thus there is a daughter excess.

Like uranium, potassium has an isotope with a long half-life (1.25×10^9 years). A small, but significant, proportion (0.01167 per cent) of all potassium is the radioactive isotope ^{40}K. This forms the basis for the techniques of potassium–argon (K–Ar) and argon–argon (Ar–Ar) dating of volcanic rocks (see VOLCANO). ^{40}K undergoes radioactive decay to produce either ^{40}Ca or ^{40}Ar. Argon is an inert gas, and while magma is molten any ^{40}Ar produced will be driven off, eventually making its way into the atmosphere (where it constitutes ~1 per cent by volume). Once crystallization occurs at the time of eruption, argon is unable to escape and begins to accumulate within the minerals crystal structure. Thus the ratio of the parent isotope (^{40}K) to the daughter product (^{40}Ar) provides a means of dating the volcanic eruption – this is the K–Ar method. The ratio of the parent and daughter isotopes can be measured more precisely by irradiating the sample of volcanic tephra or lava in a neutron beam in a nuclear reactor. This causes a proportion of the potassium to transform to ^{39}Ar, an isotope not found in nature. The age of the sample can then be found by measuring the ratio of two argon isotopes, ^{39}Ar (which is now a measure of the potassium concentration) and ^{40}Ar. Measuring this isotopic ratio is a more precise analytical process than measuring potassium and argon separately. Equally importantly, both argon isotopes are measured on the same subsample, thus allowing samples as small as single tephra crystals to be dated. Using the Ar–Ar method, ages as recent as a few thousands of years can be obtained (e.g. Renne *et al.* 1997, Figure 33).

An alternative approach to dating is not to measure the concentration of radioactive isotopes directly, but instead to look at the effect that the radioactivity has on materials in the natural environment – these are radiogenic methods. One such method is fission track dating. The most common way in which uranium decays is by the emission of an alpha particle (consisting of two neutrons and two protons). However, ^{238}U may also undergo fission, whereby the nucleus (consisting of 92 protons and 146 neutrons) splits into two new nuclei of almost equal masses. A significant amount of energy is released at the same time, and the two nuclei (the fission fragments) recoil away from each other. This leads to ionization of the crystal along these tracks – this damage can be made visible by etching the crystal surface using acids, and the number of fission tracks counted. The method is most commonly applied to volcanic rocks, including far-travelled tephra, and dates the formation of the crystals. Zircons have the advantage of high uranium concentrations (typically between 10 and 1,000 ppm) meaning that the number of tracks produced in a given time will be high. Glass has a much lower uranium concentration (~1 ppm) but is the most abundant component of tephra, and it too can be used for fission track dating providing that a method such as Isothermal Plateau Fission Track Dating (ITPFT) is used which compensates for the ability of glass to naturally anneal fission tracks (Westgate 1989).

Luminescence techniques are also based on the effects of radioactive decay. Alpha, beta and gamma radiation, resulting from the decay of various radioactive elements in the Earth's crust, is ubiquitous. When this radiation is absorbed by commonly occurring minerals such as quartz and feldspar, the energy from the radiation may be used to trap electrons at excited sites within the crystal. In effect, the mineral grains act as dosemeters, integrating the total amount of radioactivity that they are exposed to. In the laboratory, these mineral grains can be stimulated, allowing the trapped electrons to release their stored energy. The energy is released as light emitted from the quartz or feldspar grains – it is this light that is called luminescence. If the mineral grains are stimulated by heating (typically up to 500 °C) then this is termed thermoluminescence (TL). For geological materials it is normally more appropriate to stimulate them using light of a fixed wavelength (e.g. 532 nm from a $NdYVO_4$ laser) in which case optically stimulated luminescence (OSL) is observed. The luminescence signal is light sensitive, and exposure to natural daylight reduces the luminescence signal to a low level. Many subaerial transport processes will entail exposure of mineral grains to daylight (e.g. AEOLIAN PROCESSES) and the sediments deposited by these processes (e.g. DUNE, AEOLIAN; LOESS) are ideally suited to luminescence dating (Stokes 1999). Upon burial the continued exposure to radiation from the natural environment causes the trapped electron population to increase with time. The OSL signal is reset by exposure to daylight more completely than the TL signal, and hence the use of OSL has allowed more precise ages to be obtained and has allowed younger samples to be dated. In environments where the exposure to daylight at deposition can be assumed, events as recent as the last 50–100 years can be routinely dated.

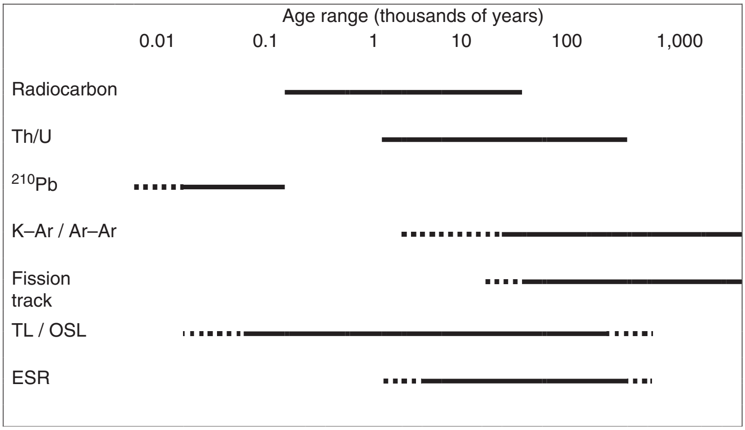

Figure 33 Age ranges over which various radioisotopic and radiogenic dating methods can be applied. The exact limits are often determined by the nature of the material being dated, and the dashed lines reflect this variation from one application to another

Electron spin resonance (ESR) dating is another technique based on measurement of the charge trapped in materials due to radiation from the environment. While TL and OSL are applicable to quartz and feldspar in sediments, ESR can be applied to stalagmites, tooth enamel, corals and sometimes bones.

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SEE ALSO: cosmogenic dating; dendrochronology; lichenometry

G.A.T. DULLER

DAYA

Small, silt-filled, closed solutional depressions found on limestone surfaces in some arid areas of the Middle East and North Africa. They are a type of PAN.

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A.S. GOUDIE

DEBRIS FLOW

Debris flows are MASS MOVEMENT phenomena transitional between LANDSLIDES and sediment-laden water floods. They occur commonly in tectonically active regions subject to rapid uplift and erosion. Typically debris flows consist of churning, water-saturated mixtures of poorly sorted sediment and miscellaneous detritus, which rush down slopes and funnel into channels when they reach valley floors. Debris flows generally form abrupt surge fronts, attain peak speeds greater than 10 metres per second, and include up to 70 per cent solid particles by volume. As a consequence, debris flows can denude slopes, damage structures, drastically alter stream channels and endanger human life. Notable debris-flow disasters include those in Armero, Colombia, 1985, and Vargas state, Venezuela, 1999, each of which resulted in more than 20,000 fatalities.

Debris flows have some alternative names. For example, LAHAR is a commonly used Indonesian term for a debris flow that originates on a volcano, and mudflow describes a debris flow that consists predominantly of silt and clay. Such fine-grained flows are rare in SUBAERIAL settings but more common in submarine (see SUBMARINE LANDSLIDE GEOMORPHOLOGY) environments.

Most subaerial debris flows commence as rapid landslides triggered by intense rainfall or rapid snowmelt. A flow may originate from a single, discrete landslide source or from numerous, distributed sources from which debris issues and coalesces. Source areas generally slope more steeply than 25 degrees, but debris flows commonly scour bed and bank sediment from channels that slope as gently as about 8 degrees. On flatter slopes debris flows typically decelerate and form lateral LEVEEs and lobate deposits that are very poorly sorted and readily distinguished from fluvial deposits. Many ALLUVIAL FANs in tectonically active regions are composed largely of debris-flow deposits.

Debris flows have a remarkable ability to flow quite fluidly, despite having grain concentrations comparable to those of static soil. The fluidity of debris flows results principally from a phenomenon called LIQUEFACTION, which occurs when pressure in the intergranular pore water rises to levels sufficient to support the weight of the overlying debris, thereby reducing friction at grain contacts. The reduced friction allows grains to move smoothly past one another, facilitating downslope flow. Liquefaction commences when debris flows begin to mobilize during landsliding of loosely packed soil or sediment, which contracts during shear deformation and transfers pressure to the intergranular pore water. Liquefaction persists in debris-flow bodies because silt and clay-sized sediment impedes pore-pressure dissipation, even if the fine sediment comprises just a few per cent of the debris-flow mass.

Effects of liquefaction are reduced or absent at the heads and lateral margins of debris-flow surges, where high concentrations of coarse debris accumulate. Debris-flow deposition occurs because coarse-grained marginal debris lacks high pore-water pressures and exerts strong frictional resistance to motion.

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RICHARD M. IVERSON

DEBRIS TORRENT

Debris torrents are a regional phenomenon, extensively documented in the coastal Pacific north-west of the United States, British Columbia and south-east Alaska. A debris torrent is defined as 'a mass movement event that involves water-charged, predominantly coarse grained inorganic and organic material flowing rapidly down a steep, confined, pre-existing channel' (Van Dine 1985; Slaymaker 1988). This is a North American usage which contrasts with the European usage of the term torrent (*torrent* in French; *torrente* in Italian and *wildbach* in German). In Europe, torrent is descriptive of mountain stream morphology and not of a debris discharge event (Aulitzky 1980). Descroix and Gautier (2002) describe the appearance and disappearance of torrents (in the sense of a distinctive morphology) in the southern French Alps as a function of climate and land use changes.

Swanston (1974) and Hungr *et al.* (1984) have argued that the term 'debris torrent' is highly descriptive and well suited to the particular character of coarse-grained, channelized mass movement events of the Pacific maritime mountains. Slaymaker (1988) has argued that the case for debris torrents as a separate category is that they are a form of channelized debris flow which lack a fine-grained fraction, particularly clay, and have a relatively large organic debris content.

Debris torrents tend to occur in small drainage basins, from 0.1–10 km^2 (Mizuyama 1982); have steep channels, with an initiation zone greater than 25°, an erosion/transport zone (10–25°) and a depositional zone (5–12°); occur in high runoff intensity zones and require substantial amounts of organic and inorganic debris available for mobilization. Triggering mechanisms include storm and/or snowmelt runoff, water release from subglacial or lake storage, log jam bursts, rockfall, debris or snow avalanches from upslope or seismic shaking. The history of sediment accumulation in the channel is also critical (Bovis and Dagg 1987). Little cohesive material is present in debris torrents, a high proportion is gravel and boulders and wood and organic mulch is prominent. A frontal and lateral 'macrostructure' consists of framework supported boulders which are pushed forward by a turbulent slurry. The slurry is extruded through the macrostructure, effectively producing a two-phase flow.

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SEE ALSO: debris flow; mass movement

OLAV SLAYMAKER

DECOLLEMENT

A fault surface marking where crustal deformation occurs in a parallel fashion, usually between an upper mechanically weak horizon, layer, or boundary, and a lower undeformed boundary. Decollements or decollement surfaces are formed by the upper rock series sliding over the lower during folding, and so is associated with overthrusting. They are typical between crystalline basement rock overlying sedimentary rock, often in thrust faulted regions such as the Alps, the Jura Mountains and the Zagros Mountains of Iran.

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SEE ALSO: crustal deformation

STEVE WARD

DEEP-SEATED GRAVITATIONAL SLOPE DEFORMATION

Deep-seated gravitational slope deformations (DGSDs) are gravity-induced processes which evolve over a very long time interval and usually affect entire slopes, displacing rock volumes up to hundreds of millions of cubic metres over areas of several square kilometres with thicknesses of several tens of metres. The main feature of these processes is the probable absence of a continuous surface of rupture and the presence, at depth, of a zone where displacement takes place mostly through microfracturing of the rock mass (Radbruch-Hall 1978). Before the definition in literature of DGSDs, Terzaghi (1950: 84) contributed to this subject significantly by clarifying the difference between a 'creep' and 'LANDSLIDE' with a statement that is applicable also to deep-seated phenomena:

> A landslide is an event which takes place within a short period of time as soon as the stress conditions for the failure of the ground located beneath the slope are satisfied. By contrast, creep is more or less a continuous process. A landslide represents the movement of a relatively small body of material with well-defined boundaries, whereas creep may involve the ground located beneath all the slopes in a whole region and no sharp boundary exists between stationary and moving material.

Thus the deformation phase may be naturally followed by a sliding phase within which shear planes are recognizable, though the evolution time of these processes is hard to predict and generally extremely long.

DGSDs, thus defined by Malgot (1977), have been documented almost everywhere in the world since the end of the 1960s and described by different authors with different terms, such as sackung, gravity faulting, depth creep of slopes, deep-reaching gravitational deformations, deep-seated creep deformations, gravitational block-type movements, gravitational spreading and gravitational creep (see MASS MOVEMENT). In spite of the variety of terms used, at present the terms most frequently used to identify the main DGSD types are *sackung* and *lateral spreading*.

Sackung

SACKUNG can be described as a sagging of a slope due to visco-plastic deformations taking place at depths which affect high and steep slopes made up of homogeneous, jointed or stratified rock masses showing brittle behaviour (Zischinsky 1969; Bisci *et al.* 1996). Typical morphological features are twin ridges, trenches, gulls and uphill facing scarps in the upper part of the slopes whereas the middle and lower parts of the slopes tend to assume a convex shape because of bulging and cambering. At the foot of the slope sub-horizontal joints can be found. The displacement mechanism, though, has not been well defined. It is thought that the rock mass behaviour at depth is different from that at the surface, owing to the high confining pressure acting all over the material. Two main displacement models have been defined. Most researchers (e.g. Mahr 1977) assume that at depth, in correspondence with the central portion of the slope, a high confining pressure does not allow the formation of well-defined surfaces of rupture, permitting only viscous deformations (non-shearing model). On the contrary, at the top and toe of the slope, where these pressures are lower, such surfaces might develop. Savage and Varnes (1987) assume instead that the zone subject to ductile deformation is indeed interrupted along a shear surface located at the base of the unstable rock mass (plastic failure model).

Lateral spreading

Lateral spreading consists of lateral expansions of rock masses occurring along shear or tensile fractures. Two main types of rock spreading, occurring in different geological situations, can be distinguished (Pasuto and Soldati 1996):

1. *Lateral spreading affecting brittle formations overlying ductile units*, generally due to the deformation of the underlying material. They are characterized by prevalently horizontal movements along tensile fractures or subvertical tectonic discontinuities. Trenches, gulls, grabens, karst-like depressions in the competent rocks and bulges in the clayey material are common features in this type of deformation. The overburden of the rock slabs is generally assumed as the cause of long-term displacements affecting the underlying formations which result in the squeezing out of the weaker rock types and rock block spreading due to tensile stresses. The process may be accelerated by water percolation through the fissures and consequent softening of the clay shales. Downcutting of valleys may then induce rotational slides and rock falls, together with block tilting and rotation which may prepare the way for block slides. The process may continue and cause progressive spreading and dismembering of the rock slab. The spreading may extend for several kilometres back from the edges of plateau.
2. *Lateral spreading in homogeneous rock masses* (usually brittle) without a recognized or well-defined basal shear surface or zone of visco-plastic flow. Typical morphological evidence is given by double ridges, uphill-facing scarps, ridge-top depressions and infilled troughs. This phenomenon has been recognized as prevalent in high mountain areas. The pre-existence of cracks in the rock mass and a high relief energy are considered as favouring factors but the mechanics of the deformation have not yet been well defined.

The evolution scenarios of sackung and lateral spread are different. The former may be considered as an initial stage of rotational– translational slides, with the tendency to evolve into rock or debris avalanches, i.e. processes which may induce high geomorphological risk situations. On the other hand, the latter may correspond to an early phase in the development of block slide-type phenomena, which are usually subject to a slow evolution of displacements.

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MAURO SOLDATI

DEEP WEATHERING

Weathering studies have enjoyed a precarious role in geomorphology, at once central and yet often neglected. Rock decay due to chemical and biochemical processes mediates the rate of erosion and destruction of relief in almost all climates, and the dominance of quartz sand in clastic sediments demonstrates its effectiveness. Soil clays are products of these processes and are universally recognized without special comment. However, weathered materials frequently extend well below the classic soil profile to depths of tens of metres, and not infrequently to more than 100 m. The transition from surface soil to fresh rock is described as the REGOLITH or *weathering profile*. While there is no formal definition of what

Plate 32 Deep weathering profile (>50 m) in granite with corestones in east Brazil

constitutes *deep weathering*, some authors use the term to describe 'exceptional' depths of rock decay (Taylor and Eggleton 2001), but this reflects experience from outside the humid tropics where weathering depths exceeding 30 m are common (Plate 32). A different approach refers to denudation being 'weathering limited' where altered materials are removed more or less instantaneously following breakdown (by chemical and mechanical processes), or 'erosion limited' where stores of non-cohesive, weathered material underlie the landsurface. In the latter case the products of weathering have remained *in situ* for an unspecified period, and it implies that during this time rates of weathering have exceeded rates of erosion. It is these circumstances that lead to the formation of deep weathering profiles, often over periods of 10^6–10^7 y. In the upper zones of many deep weathering profiles the rock has been largely reduced to a mixture of clays, Al and Fe oxides and quartz sand, through which traces of the rock structure can still be seen. This material is termed SAPROLITE.

It is often stated that deep weathering is mainly associated with ancient landsurfaces of low relief and is the product of a humid tropical or subtropical environment. This reasoning is commonly applied to occurrences of deep weathering found in high latitudes, which are explained as relicts of a formerly extensive mantle of weathered rock formed at the end of the Mesozoic or in the early Cenozoic, when warm moist conditions prevailed to perhaps 60°N. In support of this view, extensive deep weathering of the Scandinavian shield rocks is found below Cretaceous sediments in South Sweden (Lidmar Bergström 1989), and 5–10 m of advanced alteration is found between Palaeogene lava flows in Northern Ireland (Smith and McAlister 1995). Deep saprolites are widely encountered throughout Western Australia, to depths of 100 m in places, and oxygen isotope and other methods have indicated ages from Permean to Miocene, when the Australian plate was far south of its present position and never in tropical latitudes (Bird and Chivas 1988). This led Taylor *et al.* (1992) to argue that time rather than climate might be the main determinant of advanced rock decay to great depths. However, deep saprolites exhibiting advanced weathering are found in Neogene terrain in the humid tropics, and have been cited from Borneo and New Guinea (Thomas 1994; Löffler 1977). Extensive planation is not recorded in these areas, so the profiles indicate high rates of weathering combined with low rates of erosion in a landscape of moderate relief in an equatorial climate under rainforest. In contrast, many deep weathering occurrences in high latitudes present features indicative of incipient rather than advanced decay. These materials are sandy, with a low clay content (typically 2−7 per cent), and are described as *arènes* (French) or GRUS (German). Occurrences of grus are found worldwide in temperate climates, and a similar material may be found at depth beneath clayey saprolites in the tropics. Grus depths are usually <15 m and commonly 3–6 m, but are not confined to landscapes of low relief (Migoń and Lidmar Bergström 2001). When all types and degrees of rock alteration are grouped together, deep weathering is found to be very widespread. It is comparatively rare in hot and cold deserts, and in areas of recent or active tectonics. Most of the regolith mantle has also been removed where there has been severe Pleistocene glacial scour. But deep profiles have been found in north-east Scotland (Hall 1985) and northern Scandinavia, where ice sheets were either cold based and non-erosive or had extended on to low ground.

The formation of deep weathering profiles poses difficult problems. For example, weathering processes are advanced by renewal of ground water and removal of minerals in solution, and will be inhibited by rising concentrations of solutes. Very deep profiles beneath ancient plateaux must, by this reasoning, require very long periods to form and need some means to export minerals in solution. Low solute concentrations in tropical rivers draining weathered landscapes are often cited in support of low weathering rates in these landscapes. The formation of a thick layer of saprolite is, therefore, considered by many to be a self-limiting system experiencing negative feedback. However, we know little about either the deep circulation of water or the potential for long distance migration of ions by diffusion processes. Arguments have been advanced in favour of hydrothermal processes being responsible for much deep rock decay, especially in granites. But many analyses have adduced oxygen isotope evidence for low temperature alteration (70 °C) as at St Austell, south-west England (Sheppard 1977), and hydrothermal mineralization is usually very restricted in extent (Ollier 1983). It is necessary to recognize the importance of interactions between meteoric water penetrating from the Earth's surface and juvenile waters generated by magmatic processes. Both are part of the global water cycle, and rock decay is ultimately a process of adjustment of mineral species to atmospheric conditions at the Earth's surface.

The existence of an extensive mantle of residual weathering products has great significance for engineers, as well as for geomorphologists and pedologists. But the nature of the weathered material is equally important. Grus behaves very differently from a clay-rich saprolite, for example. The transition from fresh rock, upward through the weathered rock towards the surface soil can be complex, but models have been developed to describe the *weathering profile*, as distinct from descriptions of soil profiles (Figures 34, 35). At the base of the profile is the WEATHERING FRONT, often described as the *basal weathering surface* because of the commonly observed, abrupt transition from sound rock to a disaggregated and altered 'saprock'. Very little chemical change is required to cause expansion of rock minerals by hydration and partial hydrolysis, leading to a disruption of the rock fabric. The most commonly described weathering profiles (Figure 34) are based on examples in jointed granites, and similar features are found in basaltic lavas and in feldspathic sandstones. But in banded and foliated metamorphic rock, such as schists, profile subdivisions may be indistinct.

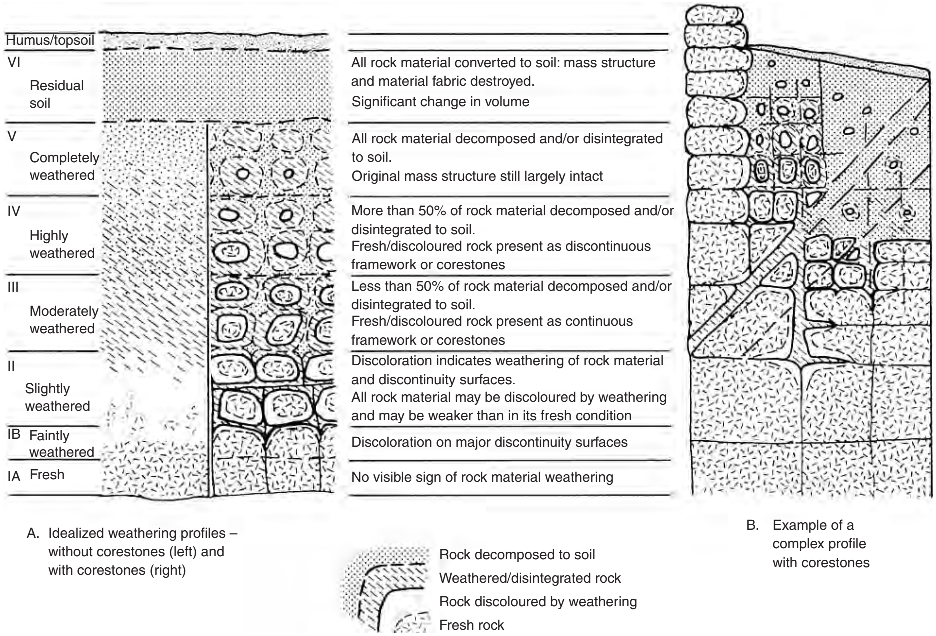

Figure 34 Characteristic weathering profiles with commonly used weathering grades shown in far-left column. Compiled by the author for Fookes (1997)

The 'granite model', first formally described from Hong Kong (Ruxton and Berry 1957), has been refined for the use of engineers (Fookes 1997); other models have been developed to describe mineralogic changes or the occurrence of specific weathering zones, including *laterite* (Figure 35). Chemical and mineralogical changes down profile are important in mineral prospecting, and the nature of the clays predicts engineering behaviour. The understanding of complete regolith profiles can be difficult due to problems of sampling, complexities of rock structure, and the mineral transformations caused by changing hydrologic conditions over long periods. But the issue is important if partly eroded (truncated) profiles, often found in the field, are to be correctly described and understood. The properties of soils in areas of deeply weathered rock, are strongly influenced by the degree of pre-weathering, which limits the availability of cations for plant growth. In many parts of the tropics, several generations of soils may have been formed, lost by erosion and re-formed within deeply weathered parent materials (Ollier 1959).

In TROPICAL GEOMORPHOLOGY, the role of the weathered mantle in determining landscape forms has been widely discussed (Thomas 1994). The balance between the rate of weathering and the rate of erosion is central to questions about the degree of alteration of near-surface weathering products on the one hand and the exposure of fresh rock forms on the other. Estimated rates of weathering on silicate rocks range from 2–50 m Ma^{-1} (mm ka^{-1}). Although surface erosion rates may exceed the highest value by two orders of magnitude, many forested slopes of moderate inclination in the tropics erode at rates less than 5 mm ka^{-1}. But data are sometimes contradictory and it is difficult to generalize. Circumstantial evidence for low rates of erosion in undulating, forested terrain comes from the partial conformity of weathering zones with present-day relief, which often exhibits multiconvex weathered compartments (*demi-oranges,*

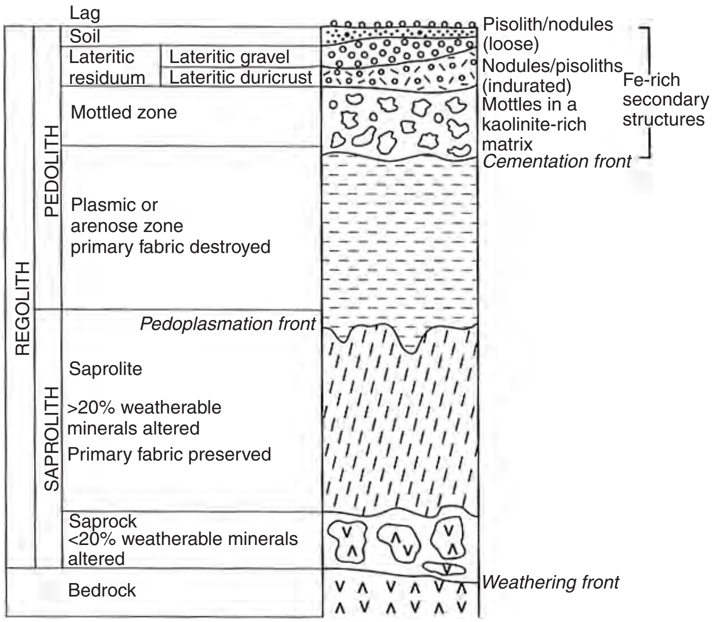

Figure 35 Scheme for regolith terminology in a profile with laterite. From Eggleton (2001)

French; meias laranjas, Portuguese). In the more arid areas of Africa and Australia, weathering profiles have been widely truncated, leaving mesa-shaped tabular hills capped by FERRICRETE or SILCRETE DURICRUSTS or landscapes with shallow, sandy regoliths and frequent outcrops (boulders, TORS, INSELBERGS). In central Australia, reference is made to 'the weathered landsurface' (Mabbutt 1965), and to the varied results of partial stripping of the regolith. Such landscapes have been described as varieties of 'etched plain' (see ETCHING, ETCHPLAIN AND ETCHPLANATION). A wider interpretation views these characteristics as the result of a 'cratonic regime' (Fairbridge and Finkl 1980) involving landscape stability and advanced weathering lasting perhaps 10^7–10^8 y, alternating with periods of erosion and regolith with a duration of 10^5–10^6 y.

In detail, the patterns of deep weathering can be shown to follow petrographic variations and structural weaknesses within rock masses. Ferromagnesian minerals and plagioclase feldspars decay more rapidly than orthoclase and mica in granites, and adjacent plutons containing different mineral suites often show contrasts in weathering. Potassium-rich intrusive rocks and silicified metamorphic gneisses, in particular, resist chemical attack. Intersecting joint patterns often outline basins of deeper rock decomposition. Deep erosion into ancient granite plutons exposes massive compartments under compressive stress that resist weathering although subject to spalling, while younger, higher level intrusives are usually subdivided along many joint directions. Where geology is uniform, patterns of weathering often respond to the relief, deeper weathering being found beneath convex summits in forested environments. While this can result from dissection into an extensive, deep saprolite mantle, the better drained conditions beneath upper slopes contribute to more rapid decay. Most perennial rivers flow in bedrock channels, but some channels in plateau landscapes alternate between anastomosing reaches containing rapids marking exposed fresh rock and meandering channels where the river flows above saprolite.

Deeply weathered landscapes were formally very extensive in Europe (and elsewhere). So-called 'lateritic' weathering covers were partially stripped from the Hercynian massifs of Europe during the Cenozoic, and are found today in the deposits of the Aquitaine, Paris and many other sedimentary basins. These were described by Millot (1970) as 'siderolithic facies', and elsewhere as 'laterite derived facies' (Goldberry 1979) and 'red beds'. During the Neogene a renewal of the regolith cover occurred patchily in the broken relief, resulting from Alpine tectonics. But the short duration of this period and the cooler climates of higher latitudes resulted in thinner, poorly differentiated sandy 'grus'. In the tropics, the breakup of Gondwanaland and drifting apart of the continents during the last 100 Ma also led to deep dissection and the infilling of downwarped and faulted sedimentary basins with the detritus of Mesozoic weathering. These are known as the Continental Terminal in west Africa and the Barreiras Formation in South America. Climatic vicissitudes have involved aridification of many tropical areas after the mid-Miocene, halting the advance of the weathering front in some drier regions. Elsewhere the warmth, humidity and biological productivity have combined to produce younger saprolites with well-defined profiles.

Paradoxically, the most rapid weathering probably takes place in tectonic regions, where a combination of high rainfall, the occurrence of marine sediments (limestones, greywackes) and epithermal igneous rocks, plus stress fracturing of nearly all formations, promote weathering penetration and contribute to high erosion rates (Stallard 1995). However, the steep slopes erode rapidly and become weathering limited, deep weathering is therefore rare. In the humid tropics, steep terrain is subject to frequent landsliding, and the regolith becomes unstable when depths of 5–6 m are reached. As relief and slope are reduced, weathering profiles deepen and there is a need to research the thresholds governing this balance. Observations suggest that, where slopes are reduced below *c.*20°, weathering rates under forest can keep pace with the rate of regolith loss by erosion.

The phenomenon of deep weathering is, therefore, an expression of the formation and survival of materials in equilibrium with near-surface Earth environments. It involves the decay of minerals contained in rocks formed under pressure and in the absence of atmospheric gases, organic acids and micro-organisms, all of which are agents of chemical change. It is also an expression of the fluctuating rates of denudation in time and space. The great stores of saprolite that occur on the continents are, in some areas at least, relicts of the remote past, but weathering processes are continuous and deepening of the weathering mantle occurs where weathering is favoured and rates of denudation low.

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MICHAEL F. THOMAS

DEFLATION

The process by which the wind removes fine material from the surface of a beach or desert. Large particles are left behind as a deflation lag and can cause ARMOURING and STONE PAVEMENT formation. Deflation from largely vegetation-free surfaces can create DUST STORMs and contribute to the formation of various features of wind erosion, including PANs and YARDANGs.

A.S. GOUDIE

DEGLACIATION

Deglaciation means the time period of uncovering of land or water by ice due to glacier retreat, normally forced by a climate change, in contrast to the term glaciation meaning the period of covering of land by ice. Deglaciation is related both to the major retreat of continental or regional-scaled glacier ice masses (large-scale deglaciation), especially during the glaciation phases in the Pleistocene, and for glacier shrinkage during the Holocene neoglaciation, e.g. since the Little Ice Age (small-scale deglaciation).

Deglaciation is triggered through climate changes (long-term) or climate variations (short-term) which impact the GLACIER mass balance due to changes in snow precipitation (accumulation) during winter and energy balance (e.g. temperature, radiation, latent heat release) during summer (ice melting). Increased summer energy input will increase ablation and cause an immediate response as retreat of the glacier front. The dynamic response of the glacier due to positive or negative mass balance of the glacier causes changes in the ice flux (mass transport) from the accumulation area down to the lower ablation area and will result in an advancing or retreating glacier front. The glacier front position will react on this forcing after a certain time period, known as the reaction time and the response time. The reaction time is given as the time lag between when the changes in mass balance occur and the first visible dynamic response of the front, and the longer response time defines the period until the glacier has stabilized to the new mass balance. These timescales are related to the dynamics of the glacier, and the glacier geometry, and can vary from a few years on a small valley glacier to several hundred or even thousands of years on large outlets from an inland ice sheet. The geometry and hypsometry (area-altitude distribution) of the glacier is important for the response. For example, if the Equilibrium Line Altitude (ELA) is raised by 100 m due to a warmer climate the increased area affected by more melting will be larger on a flat, wide glacier, and smaller on a steep, narrow glacier. Higher summer energy input, giving an immediate glacier front retreat and gradually lower ice transport over time, almost always causes deglaciation. During the last decades several energy balance models coupled to glacier-dynamical models have been developed, allowing the spatial and temporal simulation of glacier retreat due to different types and magnitudes of climatic forcing (e.g. Oerlemans 2001) (Figure 36).

The rate of deglaciation thus depends on climatic and topographic factors. Glaciers ending on land will usually get thinner and flatter during deglaciation. Calving glaciers will keep their steep, calving front, but get thinner and retreat much faster than glaciers ending on land. If the glacier is grounding in water, the reduction of mass flux may trigger buoyancy forces to lift up parts of the glacier front, which in turn leads to a massive up-calving of the glacier front. Such a rapid ice retreat occurred in deep fjord areas in western Scandinavia during the deglaciation of the Weichselian ice sheet (e.g. Sollid and Reite 1983).

The high temperature variability during the Pleistocene has caused numerous deglaciation phases in the northern hemisphere (Figure 37). According to present knowledge, there have been more than forty phases of glaciation and deglaciation during the Pleistocene. The last deglaciation was forced by a rapid increase of temperature. The mean Holocene temperature is about 10–13 °C

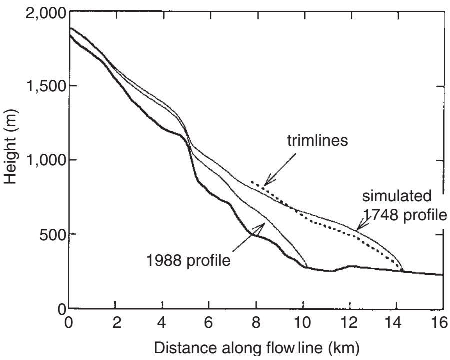

Figure 36 Glacier front positions at the outlet glacier Nigardsbreen, southern Norway (from Oerlemans 2001). The positions are obtained by applying a combined mass balance and dynamic glacier model

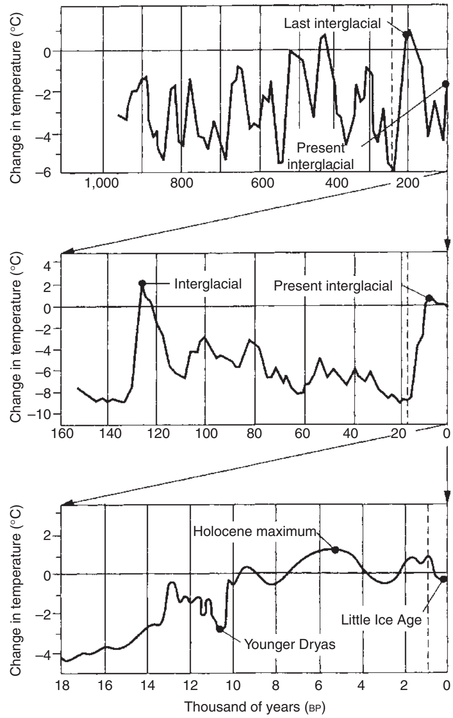

Figure 37 Temperature variation during the Pleistocene and Holocene, indicating glaciation and deglaciation phases (changed from Siegert 2001)

warmer than the mean temperature during full glacial conditions (Figures 37, 38). From the Greenland ice cores several rapid changes, called Dansgaard–Oeschger events, have been observed during the last glaciation (Dansgaard 1993). They all show an extremely rapid temperature increase of about 10° over only a hundred years followed by a slow cooling over several hundred years. At about 10,000 BP the temperature increased quickly again and stabilized at the Holocene temperature level, causing a rapid deglaciation. The forcing mechanisms for these large temperature changes over short time periods are discussed but not yet known. During Holocene the warmest period was in early to mid-Holocene and in many mountain regions the glaciers were probably melted away in the period 8,000–6,000 BP. The climate became colder from about 3,000 BP, starting the increase of glaciers in high mountain areas of the world (NEOGLACIATION), with a culmination in the period between the thirteenth century and about 1750 in Europe (e.g. Nesje *et al.* 2000), or about a hundred years later in some regions of the world. This period is known as the Little Ice Age. Since then deglaciation has prevailed until recent times in most glacial environments of the world. In some mountain regions with valley and cirque glaciers the mass loss has been massive, as for example in the Alps where the glacier retreat has resulted in a nearly 50 per cent reduction of the ice volume since the mid-1800s (e.g. IAHS(ICSI)/UNEP/UNESCO/ WMO 2001). Since the 1990s an accelerated deglaciation is observed in many alpine and in some arctic areas (Arendt *et al.* 2002; Meier and Dyurgerov 2002) and has been attributed to global warming.

Large-scale deglaciation led to a reduced weight on the landmasses, forcing a land heave. Simultaneously, the melting of glacier ice forced sea level to rise. Thus large-scale deglaciation is always related to land emergence and sea-level rise (see ISOSTASY; EUSTASY). During the maximum Weichselian ice extension global sea level was approximately 120 m lower than today. During the last deglaciation of the Weichselian ice sheet, the deglaciated area showed a net land heave, e.g. in Scandinavia. The central Bothnian area has emerged by more than 800 m since the deglaciation. Land heave produces continuously new coastlines with corresponding landforms, such as BEACH RIDGES and coastal ABRASION platforms, indicating the marine limits during a period of time. The spatial relationship between land heave rates and marine limits has been used for relative dating (see DATING METHODS) of ice-recessional landforms. Furthermore, land heave results in subaerial exposure of the former sea bottom, covered by mainly marine clay-rich sediments. Areas covered by marine clays are abundant in areas of deglaciation, such as in Scandinavia and Canada. Having a high nutrient content and being easily arable these areas were of interest for early settlement and agriculture. However, these sediments are highly unconsolidated, and since deglaciation subject to severe GULLYing and prone to landsliding (see LANDSLIDE; QUICKCLAY).

The period of deglaciation is also the period of sediment accumulation by glacier ice and meltwater. Deglaciation is not a continuous process. Especially during the early phase of deglaciation the recession of ice margins frequently halted ('stagnation') or smaller re-advances occurred due to short-term climatic deterioration. The geological time periods of the Younger Dryas and the Preboreal are examples of such short-term climatic variations, which morphologically can be followed by glacial landforms almost throughout southern Finland, central Sweden and coastal Norway (Figure 38).

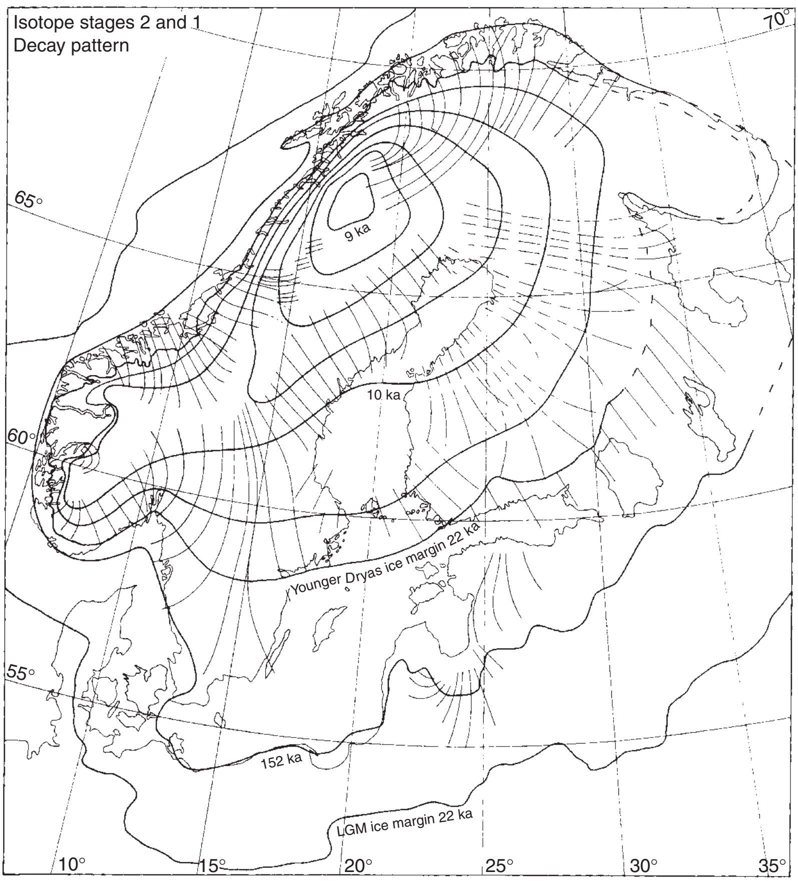

Figure 38 Glacial decay pattern of the Weichselian glaciation in Fennoscandia since the last glacial maximum (LGM, 22 ka) and the onset of the Holocene (*c.*10 ka) (adapted from Kleman *et al.* 1997)

The type of glacial land system (see Plate 33) during deglaciation phases is dependant upon whether the glacier front ends in water (marine/lacustrine) or on land (terrestrial), how fast the land is uncovered (deglaciation rate) and what glacier temperature regime prevailed during deglaciation (glacier thermal regime) (see also Benn and Evans 1998). A fast retreat of glacier tongues or lobes results in an uncovering of subglacial landforms such as fluted surface or drumlinoide forms. These subglacial landforms normally show the very last glacier movement direction. If the deglaciation happens in a permafrost environment, parts of the glacier marginal areas may be cold based. In such environments, glaciers may preserve sediments and landforms derived from earlier glaciation and deglaciation periods. Slow retreats and/or temporary stagnation of the glacier front produce landforms of accumulated glacigenic material, terminal moraines. During stagnation phases the glaciers advance some metres during winter and retreat during summer due to variations of ablation. The winter advances produce small annual moraines. Produced under water they are called DeGeer-moraines in marine environments and cross-valley moraines in mountainous lacustrine environments. Such moraines often build a sequence of landforms, used in deglaciation reconstruction. Glacier margins in permafrost environments are cold based. In this setting, a net-freezing condition along the glacier base prevails (e.g. Boulton 1972), forcing glacigenic material to accumulate in the glacier front area. Ice motion

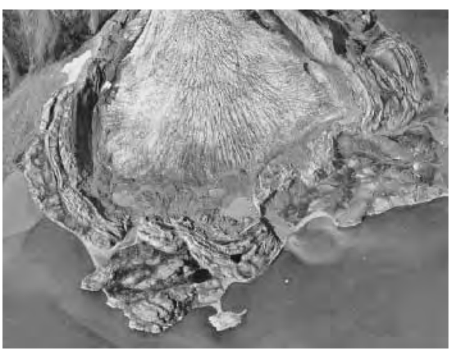

Plate 33 Oblique air photo of recently deglaciated terrain, Erikbreen, northern Spitsbergen (from Etzelmüller *et al.* 1996)

and surface melting leads then to an accumulation of glacigenic sediments on the glacier surface. If the thickness of this layer is larger than the ACTIVE LAYER thickness of the environment (see PERMAFROST), the ice below the layer is preserved, preventing further ablation. These ice-cored moraines can persist over long periods during deglaciation. If the active layer thickness increases or is eroded by e.g. fluvial action, the ice core melts and produces a hummocky moraine terrain proximal to the glacier front area and a clear distinctive border distal from the glacier. Deposited on steep terrain, the ice-cored moraines may continue creeping, producing ROCK GLACIER-like landforms. If the glacier front ends in water, most sediment transported in glacial meltwater is deposited in the vicinity of the glacier front, building up ice-contact deltas. When built up to the water surface, they give an indication of sea level during the deglaciation phase. Especially during the last stages of deglaciation, parts of the ice streams may become decoupled from the active part of the glaciers. This results in the loss of glacier flux and thus dead-ice wastage occurs. In this situation ice can be buried by e.g. glacifluvial sediment. Melting of these ice bodies results in typical dead-ice landforms, consisting of hummocky irregular terrain and KETTLES AND KETTLE HOLES. On cold ice caps, meltwater is often routed along the glacier margins, forming channels that mark the ice surface during phases of deglaciation. Swarms of subsequently lower channels along mountain slopes indicate the lowering of the ice surface and the glacier surface slope during different phases of deglaciation.

On continental ice sheets the ice divide did not necessarily correspond with the position of the topographic water divide of the underlying relief. During deglaciation topographic water divides often became ice free before all ice disappeared. This ice could occasionally block the drainage and thus formed ice-dammed lakes that drained over local or regional water divides. Like terminal moraines, lacustrine sediments and landforms such as shorelines bear witness to periods of deglaciation. Sudden outbursts of glacial lakes are called by the Icelandic word *jokullhaup*. In many high mountain areas, e.g. in central Asia, deglaciation of valley glaciers leads to damming of lakes between the glacier front and terminal moraines, which often are ice-cored. These lakes are unstable, and outbursts, often called GLOFs – glacier lake outburst floods – are a potential hazard for lower lying valleys and human settlements and infrastructures. The same risk applies to the situation where valley glaciers block the drainage from minor side valleys.

Research on deglaciation is concentrated on (1) determining the start of deglaciation, (2) the deglaciation rate, and (3) the change of the spatial distribution of the ice body during different phases of deglaciation. Traditionally, scientists concentrated on determining the start and deglaciation rate, by dating of landforms associated to terrestrial or marine/lacustrine glacier margins (see DATING METHODS) and analysing sediment succession building up these landforms. Coring of marine sediments on continental margins and deep ocean basins revealed continuous information on glaciation and deglaciation phases during the Pleistocene (e.g. Elverhøi *et al.* 1995). The Holocene NEOGLACIATION chronology is depicted by core analysis of local sediment sinks such as lakes fed by meltwater from glaciers (e.g. Karlén 1976). The spatial distribution of ice bodies during deglaciation phases is often determined through GEOMORPHOLOGICAL MAPPING of glacial landforms combined with dating methods. The past vertical extent of an ice sheet in its accumulation area is difficult to obtain because of the lack of marked landforms due to low glacier velocities and often cold ice in culmination zones of ice sheets. Recently, exposure dating using cosmogenic isotopes has proved to be a helpful tool in this respect.

Deglaciation has become an important problem for human settlement and sustainable development in many high mountain environments. Especially in many semi-arid areas, such as the eastern slopes of the Andes Mountains and in central Asian mountain ranges, glaciers act as freshwater reservoirs, since meltwater from the glaciers is important for irrigation and water supply. Deglaciation leads to a periodically enhanced runoff. However, due to shrinkage of the water reserve (the glaciers), water availability will be reduced over the long term.

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SEE ALSO: dating methods; eustasy; glacier; ice dam, glacier dam; isostasy; moraine; neoglaciation

BERND ETZELMÜLLER AND JON OVE HAGEN

DELL

Small headwater valleys which are characteristically sediment-choked and swampy. Dells frequently occur at the head of deep gorges on plateaux surfaces and may be analogous to DAMBOS. Notable dells have developed on sandstone on the Woronora Plateau of New South Wales, Australia (Young 1986).

They are also known from Eocene beds in the New Forest of southern England, where they may have a periglacial origin and form tributaries to small dry valleys (Tuckfield 1986).

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A.S. GOUDIE

DEMOISELLE

A French term used to describe a needle-shaped Earth pillar composed of eroded rock, and capped by a large boulder. The overlying block is typically more resistant than the underlying material, and so tends to protect it from erosion (predominantly by water), as well as helping to maintain its vertical integrity. Demoiselles are commonly found in Alpine areas of highly weathered volcanic breccia or of glacial till. The term is derived from French for 'young lady', and is commonly employed throughout the French Alps. The term *cheminée de fees* (fairies' chimney) is also used frequently in place of demoiselle, while the American synonym of demoiselle is HOODOO.

STEVE WARD

DENDROCHRONOLOGY

Dendrochronology is the study of annual tree rings, with studies based on the measurement of variations in ring widths caused by variations in climate and environment at the time of ring formation. Ring counts and ring-width measurements provide precise calendar dating for ring formation years and a basis for numerous research applications.

The main disciplines included in these applications are: dendroclimatology, dendroarchaeology, art historical research, history, dendroecology and the related disciplines of DENDROGEOMORPHOLOGY, dendroglaciology, dendrohydrology, dendroniveology (snow and ice research), dendropyrology (fire events) and dendroseismology. Tree-ring chronologies have also been used to calibrate the radiocarbon timescale.

Douglass explored the potential of tree rings for climate analysis in 1919, but it was not until 1953 with the discovery of the 4,000-year old Bristlecone pines that the technique drew widespread attention. By the end of the century, chronologies covering nearly 10,000 years had been constructed from matching ring-width patterns of living and dead trees, and dendrochronology was being applied worldwide in research on global change, although as yet work in the tropics has been limited.

Procedures entail either cutting discs from a stem or, more usually and less destructively, the collection of wood cores with a 5 mm increment borer. The core is usually glued to a wood support, with its grain perpendicular to the support, and polished to reveal the ring structures ready for ring-width measurement.

Coring causes mechanical injury. Thus, it is important to obtain permission before taking samples and never to core anything that could be valuable as timber. However, trees compartmentalize wounds and often produce anti-fungicidal substances that generally limit damage; injuries stimulate local growth and holes are callused over in a few years. Studies have indicated that the core hole should be left open and untreated since this could introduce foreign organisms and impede the healing process.

An annual tree ring usually has two growth phases. At the beginning of the growing season conifers produce large, pale, thin-walled early-wood cells; towards the end of the season increasingly small diameter, dark, thick-walled, latewood cells develop. Hardwood trees have a variety of ring forms. Healthy, unstressed trees will produce concentric rings with approximately equal ring widths while stressed trees will form eccentric ring patterns and show narrow, variable ring growth. Trees on slopes are frequently bent, with deciduous species having their central pith displaced towards the downslope side of the stem and conifers to the upslope side (a point to be remembered when coring bent trees).

Problems for the technique, apart from those caused by growth eccentricities, are introduced by non-uniform cell growth due to adverse conditions, with normal cell formation either halted or present over only part of a stem resulting in missing rings. Alternatively, false rings can be produced by late frosts, droughts or other growth-inhibiting events resulting in darkened cells followed by resumption of normal growth before true dark latewood growth marks the end of the growing season. These complications can be mitigated by crossdating.

Crossdating is achieved by matching sample ring-widths using visual and statistical tests. At least two cores are usually collected from each tree, so that they can be crossdated to check for missing or false rings and, where there are none the radii are averaged to show mean annual ring growth. Graph plots of the means of individual trees are then compared and crossdated and a site masterplot created.

Sample depth (number of trees sampled) will change through time affecting the quality of a chronology. Consequently, chronologies may be truncated where there are less than three trees to support a mean curve and, for valid climatic results, curves should contain an absolute minimum of ten trees per site, but thirty or more are desirable. Where sample depth is important, this information should be included on ring-width graphs.

The technique was revolutionized by the advent of computer processing enabling the digitizing of ring-width measurements; rapid plotting and comparison of graphs; rapid application of multi-variant statistics, and radio-densitometric determination of wood density. This latter approach is based on X-ray analysis of changes in cell densities; it is used particularly in climate studies to highlight sensitive reactions of cell densities to temperature variations. It is also used for analysis of tropical species' growth, since these species, rather than always forming annual rings, may produce growth zones reflecting aperiodic precipitation or drought.

Prior to computerization, measurements were made by hand and one approach to crossdating was the use of 'skeleton plotting' based on ring counting, visual assessment of relative per cent ring widths, and identification of event years. Skeleton plots provide dating for, and a visual summary of, the effects of environmental events without the need to make precise ring-width measurements. It is a useful procedure showing major events and growth trends where rapid assessment of limited sample numbers is required.

Apart from the effect of sudden events on tree growth, slow growth changes may occur due to gradual variations in climate or natural reduction in ring width as a tree ages. This latter effect is routinely removed by standardizing (detrending) ring widths using various techniques. Standardization, apart from eliminating age trends, emphasizes event years while removing climatic trends shown by moving averages of the mean ring-width data.

Further reading

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VANESSA WINCHESTER

DENDROGEOMORPHOLOGY

Dendrogeomorphology is based on analysis of the annual growth rings of trees or woody plants and their growth forms. It is used to investigate spatial and temporal aspects of Earth surface processes operating during the Holocene at annual to centennial timescales.

The technique is closely associated with dendroclimatology and employs largely the same methods as DENDROCHRONOLOGY. Applications include dating and establishing rates of change and frequencies of storms, FLOODs, LAKE outbursts, river channel changes, frost events and ICE surges, GLACIER movements, snow avalanches (see AVALANCHE, SNOW) MASS MOVEMENTs and FIREs, and show the relationships of events to climate. In addition, tree rings can provide records of events unrelated to climate: volcanic eruptions; earthquakes and TSUNAMIs; environmental management; COMPACTION OF SOIL; water table variations, changes in pollution and saltwater ingression.

Methods, other than those used in dendrochronology, include studies of the age, anatomy, morphology, and structures of tree roots, stems and crowns. Root ring patterns can be used to date sediment aggradation or degradation. Trees respond to increases in soil depth by producing adventitious roots; soil movements cause root structures to bend while degradation leaves roots exposed. Ring counts supply dates for root structures, bends and stem age at ground level while age and distances between features show the scale of events. Eccentric ring patterns develop where roots are part-exposed or when denudation brings them close to the surface. Changes in patterns supported by changes in cell structures can be dated. Before sampling buried or exposed roots, records should be made of all relevant features: positions and orientations of main and adventitious root systems; distances to the ground surface; vegetation cover and soil type.

Stems deformed by site changes produce eccentric ring patterns. Stem discs or cores taken both in the direction of stress and at right angles, show when eccentricity begins and the orientation of patterns provides information on the direction of changes. Injuries to stems or roots produce scarring with local growth being stimulated. A core from an undamaged area near the wound (but avoiding re-growth tissue) will show the number of years elapsed since the event.

Crown development provides information on competition, wind and storm events, snow cover and tree health.

The main problems for dendrogeomorphology where surface age is the focus of interest are to establish the total age of a tree and the length of time taken for a tree to colonize a freshly exposed surface. Core ring counts only show the age of a tree above the coring point; thus to find total age an estimate of the number of years growth below this is required. One method is to cut stem discs near ground level of a number of small trees growing in a range of local microenvironments, correlate the height of the trees with their age and calculate the mean growth to height ratio for the location. Verification of ecesis (colonization) times requires an alternative dating source.

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VANESSA WINCHESTER

DENUDATION

On Earth, two forces counterbalance: uplift (i.e. creation of relief) and denudation. Denudation includes all processes that remove the relief at the surface of the Earth. Denudation acts chemically or physically. Chemical denudation, also termed chemical weathering or chemical erosion, is the slow complete or partial dissolution of rock minerals. Physical denudation or mechanical weathering processes correspond to the removal of solids from the land surface. Quantification is generally expressed by the mean of chemical or physical fluxes (or rates) of denudation, expressed most often in $\mathrm{t\,km^{-2}\,a^{-1}}$.

The adaptation of rock minerals to the conditions at the surface of the Earth (Ahnert 1996) releases the most soluble elements and leads generally to the formation of residual minerals, usually clays and hydrous iron or aluminium oxides, that accumulate at the interface between the atmosphere and the lithosphere (the REGOLITH). Examples of chemical denudation reactions are the weathering of calcite, which does not leave any residue

$$CaCO_3 + H_2O + H^+ = HCO_3^- + Ca^{2+}$$

and the weathering of albite, that produces a secondary phase, for example kaolinite:

$$\begin{aligned}&2NaAlSi_3O_8 + 9H_2O + 2H^+ =\\ &\quad Al_2Si_2O_5(OH)_4 + 2Na^2 + 4H_4SiO_4\end{aligned}$$

These equations show that protons are necessary to attack rock minerals. At the surface of the Earth, these protons are mostly derived from the dissolution of the atmospheric or soil CO_2 in water

$$CO_2 + H_2O = H_2CO_3 = H^+ + HCO_3^-$$

Weathering reactions therefore pump CO_2 from the atmosphere and convert it into bicarbonate ions. Ultimately, these ions, combined with the Ca and Mg ions liberated by rock weathering, will lead to the precipitation of calcite in the ocean by living organisms, allowing the sequestration of atmospheric-derived carbon. Other possible origins for protons include production by organic molecules derived from the degradation of organic matter in soils (humic and fulvic acids) and the oxidation of sulphide minerals producing sulphuric acid.

Like minerals, rocks do not weather at the same rate. A good way of estimating the rate of chemical weathering is by analysing rivers draining a single type of rock, provided that the river dissolved load is corrected from the inputs that do not derive from rock weathering (atmosphere, pollution, biomass). Another approach is based on soil mass budgets. Rates of chemical denudation are extremely variable, ranging from less than $t\,km^{-2}\,a^{-1}$ in high latitude granitic catchments or in the low-lying regions of central Africa, to more than $100\,t\,km^{-2}\,a^{-1}$ for rivers draining basalts at Réunion or Java Island (Louvat and Allègre 1997). Basaltic lithologies thus weather 10 to 100 times faster than granites. At a global scale, it has been shown by Dessert *et al.* (2002), that, even if the outcrops of basalts represent 5 per cent of the emerged surface of the Earth, the flux of CO_2 uptake by basalt weathering is as high as 35 per cent of the total consumption flux by rock weathering. Basalt weathering therefore appears as a major mechanism of atmospheric CO_2 regulation. Carbonate rocks also have high denudation rates, ranging from 10 to $200\,t\,km^{-2}\,a^{-1}$. Saline rocks have the highest chemical denudation rates because they are highly soluble in water. From large river systems, chemical denudation rates (Figure 39) ranging from a few $t\,km^{-2}\,a^{-1}$ for the Zaire, Nile and Siberian rivers to about $50\,t\,km^{-2}\,a^{-1}$ for rivers such as the Mekong, Mackenzie or Brahmaputra have been determined (Summerfield and Hulton 1994). These rates are strongly correlated to the abundance of carbonates within the drainage basin, simply because carbonates and evaporites weather at a faster rate than silicates.

Although lithology is the first controlling factor on chemical weathering rates, other parameters exert a control on chemical denudation. At both small and large scales, chemical weathering rates increase strongly with runoff and temperature. This is especially true for basalt weathering rates which respond at a global scale to an Arrhenius-type law (Dessert *et al.* 2002). At a continental scale, however, several authors have pointed out that the highest chemical denudation rates of silicate rocks are not found in the regions of highest rainfall and temperatures (Edmond *et al.* 1994). The Zaire river has the same chemical denudation rates as the Yenisey (Gaillardet *et al.* 1999). The low chemical denudation rates found in the flat and humid tropical areas contrast with the highly weathered nature of soil material (laterites) that characterize these regions. At a global scale, intensity and flux of chemical denudation of silicates are inversely correlated. This paradox will be explained later.

Physical denudation rates can be estimated by different means: by using rivers, sediment accumulation in reservoirs or sedimentary basins and

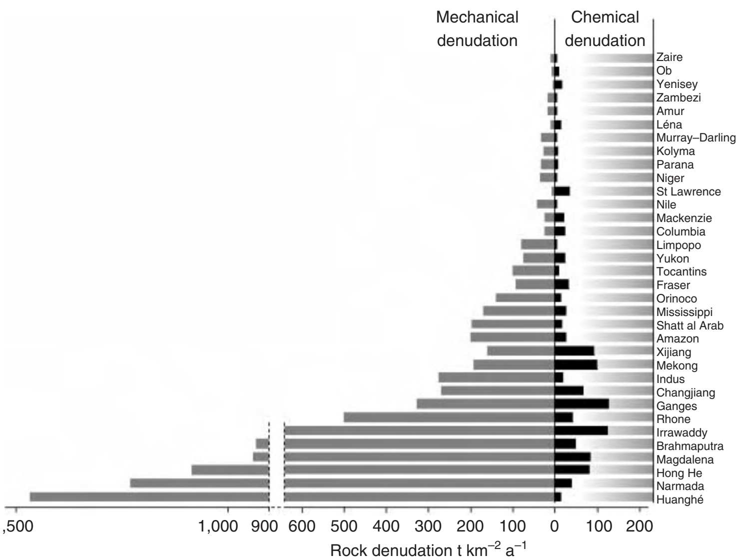

Figure 39 Physical vs chemical denudation rates for the world's largest river basins. Solute load is corrected from atmospheric inputs. Based on Summerfield and Hulton (1994) and Gaillardet *et al.* (1999)

cosmogenic isotopes. Rivers are probably the easiest way of calculating mechanical denudation rates, but this approach suffers from various limitations. Unlike dissolved load, which often fluctuates by a factor of two or so between high water and low water stages, the quantity of solids transported by rivers can vary drastically. A single daily flood event can transport as much sediment as several years of regular sediment flux. In addition, the amount of material transported as bottom sand is generally unknown. Rivers, especially large rivers, store sediments in their floodplains or alluvial fans. For example, it has been shown that two-thirds of the sediments removed from the Andes by the headwaters of the Rio Madeira in South America are stored in the foreland floodplains and never reach either the Amazon, or the sea (Guyot 1993). Finally, the influence of anthropogenic activities has usually resulted in a strong increase of river sediment yield. The best example is the Huanghe river, that transports, today, up to $20\,g\,l^{-1}$ of sediments to the ocean, while the long-term pre-anthropogenic estimate of Holocene sediment volumes deposited by the river indicates an average suspended load concentration one order of magnitude lower (Milliman and Syvitski 1992).

Mechanical denudation rates (Figure 39) have been computed for the largest river systems by a number of authors (Pinet and Souriau 1988; Milliman and Syvitski 1992; Summerfield and Hulton 1994). They range from numbers lower than $10\,t\,km^{-2}\,a^{-1}$ for rivers such as the Siberian (Ob, Yenisey, Kolyma) and tropical (like the Zaire), to numbers higher than $700\,t\,km^{-2}\,a^{-1}$ for the largest rivers of Asia (Brahmaputra, Ganges). The world average value is estimated to be $200\,t\,km^{-2}\,a^{-1}$, corresponding to $20.10^{9}\,t\,a^{-1}$ (Milliman and Syvitski 1992).

The dominant factors that influence mechanical denudation are the erodibility of rocks and relief. High relief areas tend to have high mechanical denudation rates. This is mainly due to slope instabilities and to glacial abrasion. There does not seem to exist any clear relationship between climate (runoff or temperature) and physical erosion, at least on a global scale (Pinet and Souriau 1988; Summerfield and Hulton 1994). Regions of high precipitation and high seasonality of rainfall seem however to exhibit higher mechanical denudation rates (see Goudie 1995). An inverse correlation between suspended yields and basin area is reported by an extensive study of Milliman and Syvitski (1992), possibly showing the importance of sediment storage. The same authors showed that humans are also a major controlling parameter, as fluvial denudation rates during the Holocene are estimated to be less than half the present-day rates. However, for the largest rivers of Asia, there is a remarkably good agreement between present-day fluvial physical denudation rates and long-term rates based on sediment volume (Métivier and Gaudemer 1999) accumulated in the sea. Based on the denudation rates computed on large rivers, Gaillardet *et al.* (1999) have shown that chemical denudation rates of silicate rocks are positively correlated to physical denudation rates. Such a relation is confirmed by cosmogenic nuclides measurements (Riebe *et al.* 2001). This global coupling between chemical and physical fluxes of silicate denudation is explained as follows. With a low mechanical denudation regime, soil development is favoured, leading to a shielding effect of the soil and a negative feedback on the interaction between water and mineral surfaces. This is the transport-limited regime. Conversely, in regions of high mechanical denudation mineral surfaces are continuously exposed, and even if chemical weathering is not intense, the fluxes of released solutes are increased. This is the weathering-limited regime. Overall, the present-day Earth is under the weathering-limited regime.

Total denudation rates are calculated by adding the chemical and physical denudation rates. Using a mean crustal density of $2{,}700\,kg\,m^{-3}$, these rates are usually translated in $mm\,a^{-1}$ (Figure 40). For large rivers, landscape downwearing ranges from about $10\,mm\,a^{-1}$ in regions such as the shield low-lying areas of the Congo craton, Niger basin, Brazilian shield and Australian shield to 100–200 mm per 1,000 years for the Ganges, Indus, Changjiang and Amazon rivers. The Brahmaputra river has the highest rate of total denudation, close to 700 mm per 1,000 years. The mean value for the Earth's surface is 61 mm per 1,000 years (areas of internal drainage excluded). Total denudation rates calculated by mass budgets of riverine products are in good agreement with cosmogenic isotope measurements. Cosmogenic isotopes are produced during the bombardment of cosmic rays and their abundance is a function of the production rate and total erosion rate. This technique has been applied by a number of authors and gives denudation rates, integrated over tens of millions of years in the case of ^{10}Be, which are in general agreement with other techniques. For example, the ^{10}Be derived denudation rates of the Loire, Meuse, Neckar and Regen basins are in the order of magnitude of those found by conventional techniques (Schaller *et al.* 2001).

As a global average, 10–15 per cent of the total denudation is chemical denudation, 75–80 per cent is physical denudation (Figure 39). The ratio of mechanical over chemical fluxes fluctuates widely from about 1 for the Siberian and tropical rivers to 10–20 for mountainous rivers. The present world is therefore dominated by physical denudation, but it may not have been the case for geological periods of low relief.

Atmospheric dust transport from the continent to the ocean is also responsible for the denudation of continents. Aeolian denudation is extremely difficult to quantify and estimates vary between 0.1 to $5.10^9\,\mathrm{t\,a^{-1}}$. In deserts, aeolian denudation may be the only process of relief denudation.

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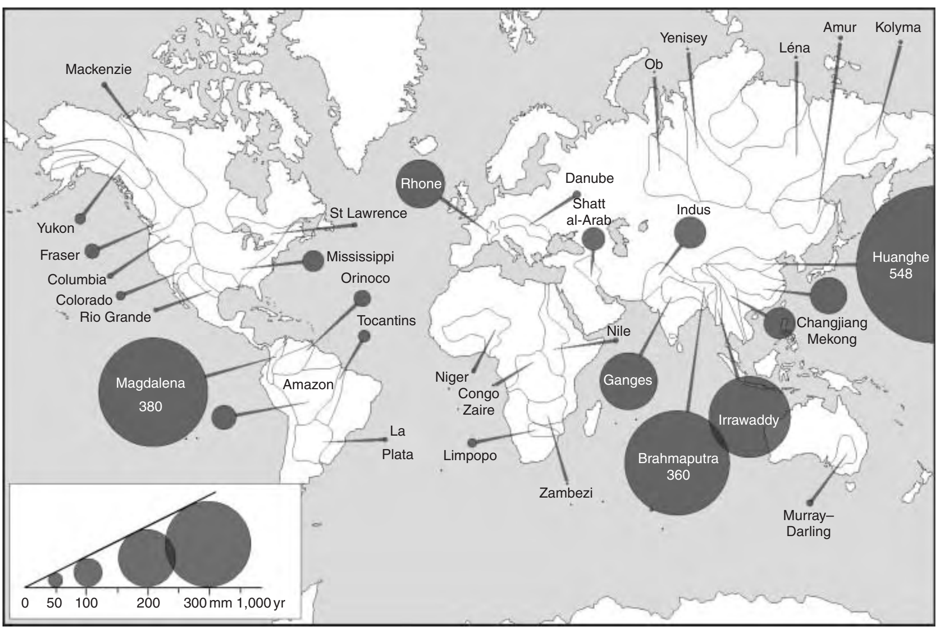

Figure 40 Total denudation rates for the largest drainage basins calculated with a mean density of $2.7\,\mathrm{g\,cm^{-3}}$. Based on Summerfield and Hulton (1994) and Gaillardet *et al.* (1999)

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SEE ALSO: cosmogenic dating

JÉRÔME GAILLARDET

DENUDATION CHRONOLOGY

The explanation of how topographic landscapes came to attain their present form has always been a prime objective of geomorphology. Prior to the 1960s, most workers adopted an essentially historical approach to landscape evolution. The aim was to identify the sequence of episodes, or stages, of erosional development that demonstrated how contemporary landscapes had been sculptured from hypothetical initial topographies that were usually considered fairly uniform and featureless. This sequential approach to topographic evolution, with its focus on DENUDATION, came to be known as denudation chronology. A significant proportion of such studies were based on the Davisian model of landscape evolution and can be termed classical denudation chronology, an historic element of a broad field of study currently known as long-term landscape development or evolutionary geomorphology.

Classical denudation chronology sought to identify evidence of past PLANATION SURFACES and erosional levels in a landscape and to place them in a chronological sequence. Two key concepts were employed. First, that topographic 'flats', bevels and benches, together with accordant ridge and summit levels, represented the remnants of marine platforms, SUBAERIAL low-relief surfaces or terraces produced during past periods of relatively stable BASE LEVEL or 'stillstand'. Second, that there had been a progressive but episodic fall in base level with time, so that the highest features (in terms of elevation) were the oldest. The resulting 'geomorphological staircases' often rose via terraces and benches to the more fragmentary remains of 'summit surfaces' preserved on ridges and CUESTAs (e.g. the 'Schooley Peneplain' of W.M. Davis (Figure 41), the 'Mio-Pliocene Peneplain' of Wooldridge and Linton (Figure 42)), or to even higher surfaces whose former existence was postulated on the basis of the summits of residual hills or 'monadnocks' (Figure 42), or by the projection of the planes of outcropping unconformities (e.g. the 'Fall Zone Peneplane' rising above the Appalachians (Johnson 1931) (Figure 41), the 'Sub-Eocene Surface' of south-east England (Wooldridge and Linton 1955) or the 'Sub-Cretaceous Surface' rising over Wales (e.g. Brown 1960)).

The identification and delimitation of such surfaces was usually based on visual observation, augmented by field mapping, profiling and various kinds of cartographic analysis, including the use of

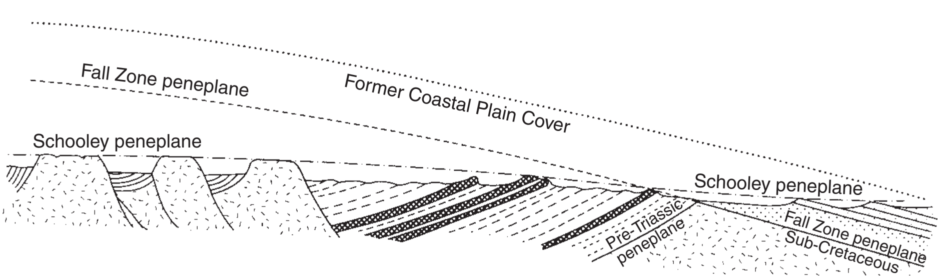

Figure 41 Scheme of landscape development for the eastern USA first advanced by D.W. Johnson in 1928

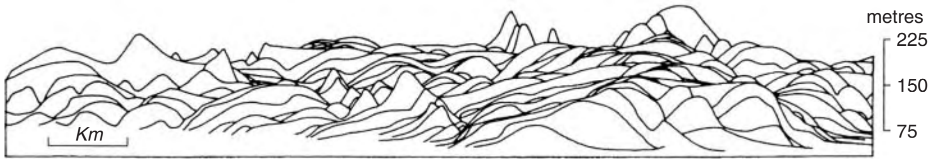

Figure 42 Projected profiles on the Chalk, western margins of Salisbury Plain, southern England, published by C.P. Green in 1974. The residual hills indicate the altitude of the former Sub-Eocene (now Sub-Palaeogene) surface and rise above the Mio-Pliocene Peneplain of Wooldridge and Linton, which Green subdivided into Summit Peneplain: Higher Surface and Summit Peneplain: Lower Surface, dated as Miocene and Pliocene respectively (see Jones 1981)

superimposed and projected profiles (Figure 42), and the resulting sequences could be extended back into the geological past by use of the basic rules of stratigraphy, such as the laws of superposition and of original horizontality and the nature of unconformable relationships. Hence many denudation chronologies were extended through the Neogene into the Palaeogene and even into the Mesozoic (Figure 41). Although this was often claimed to be a strength, because such studies provided a bridge between Quaternary landscape development and stratigraphy, in reality the resultant 'histories' had much closer affinities with historical geology than with geomorphology, especially once contemporary process studies had developed to dominate geomorphology.

Relatively little emphasis was placed on the study of surficial deposits, largely because of the limited availability of analytical techniques and the complexity of such deposits. Often the studies that were undertaken focused on whether or not the identified erosional remnants were of subaerial or marine origin. By contrast, drainage patterns and drainage evolution figured prominently. Drainage-structure relationships, together with the existence of cols, wind-gaps and UNDERFIT STREAMS, were all used to recreate former drainage patterns associated with particular stages of landscape evolution (see Brown 1960). Often the aim of such studies was to recreate the original pattern of sub-parallel consequent rivers that developed on an uplifted and tilted marine plane or were superimposed from overlying cover strata across a fundamental unconformity. The identification and interpretation of discordant drainage therefore became a highly contentious issue and led to great debates about the extent and significance of former marine planation surfaces (see Jones 1981).

It is often assumed that classical denudation chronology evolved from W.M. Davis's exposition of the concept of the CYCLE OF EROSION, but this is incorrect. Long before the Davisian model was first outlined in 1889 others had begun to develop simple, embryonic denudation histories. For example, in Britain there was the work of Ramsay in 1846 and 1864, Jukes in 1862 and Topley in 1875, while in America McGee in 1888 developed an erosional chronology for that very same part of the Appalachians that was to be made classic by W.M. Davis's own detailing of the cycle of erosion concept (Davis 1889).

The Davisian model, as subsequently refined and elaborated (see Davis 1909), with its notions of peneplanation and cyclic change due to variations in base level, clearly fitted in so well with notions of the sequential development of landscapes, that it is no surprise that the two approaches merged. Many high-level marine surfaces were reinterpreted as PENEPLAINs and every attempt was made to place identified 'flats' into discrete groupings on the basis of elevation (Figure 43) in order to recreate cycles and partial cycles which could then be correlated with evidence from adjacent regions on the basis of elevation alone. Thus the cycle of erosion concept invigorated denudation chronology and, in turn, the concept was to survive as a basic element of classical denudation chronology long after it had been discarded by the remainder of geomorphology, following advances in process-based understanding of landform development.

During the first half of the twentieth century, denudation chronology became a major preoccupation of geomorphological studies: in America, under the influence of D.W. Johnson (1931), in Britain, where S.W. Wooldridge was the dominant figure (see Wooldridge and Linton 1955) and in France, where the pioneering study of the Massif Central by H. Baulig (1928) established the blueprint for later studies in Europe. However, there were differences across the Atlantic regarding the emphasis placed on eustatic versus diastrophic mechanisms as the main cause of base level changes, with European studies following the lead of Baulig in favouring eustatic change. This is classically displayed in B.W. Sparks's (1949) interpretation of the morphology of the South Downs cuesta backslopes in southern England as consisting of eight marine levels between the two proven raised beaches below 40 m and the postulated Plio-Pleistocene marine plain at 170–200 m (Figure 43) – a geomorphological staircase encompassing the Pleistocene in no fewer than eleven treads, the majority of which lacked supporting sedimentological evidence.

Denudation chronologies were developed for other areas using modified models of landscape evolution, most dramatically in the case of South Africa where Lester King (1972) produced a classic sequential interpretation of the area between the Drakensberg Escarpment and the Natal Coast (Figure 44), including artistic representations of the landscapes developed at each stage. King's model of landscape evolution represents an amalgam of the views of Davis and Penck; episodic uplift resulting in both downwearing and backwearing, with the parallel retreat of slopes leading to the formation of PEDIMENTs which coalesced to form pediplains through the process of pediplanation. King himself went so far as to state that the morphological and sedimentological sequences identified in Natal provided the basis for a chronological scheme that had global application (King 1967), a grand design published at a time when workers elsewhere were experiencing increasing difficulty in making compatible the numerous 'local histories' that had been identified, as was well shown in David Linton's (1964) valiant attempt to provide a synthesis of Tertiary landscape evolution in Britain.

Since the 1960s there has been increasingly widespread and severe criticism of classical denudation chronology. Some of these criticisms focused on the inadequacies of the Davisian

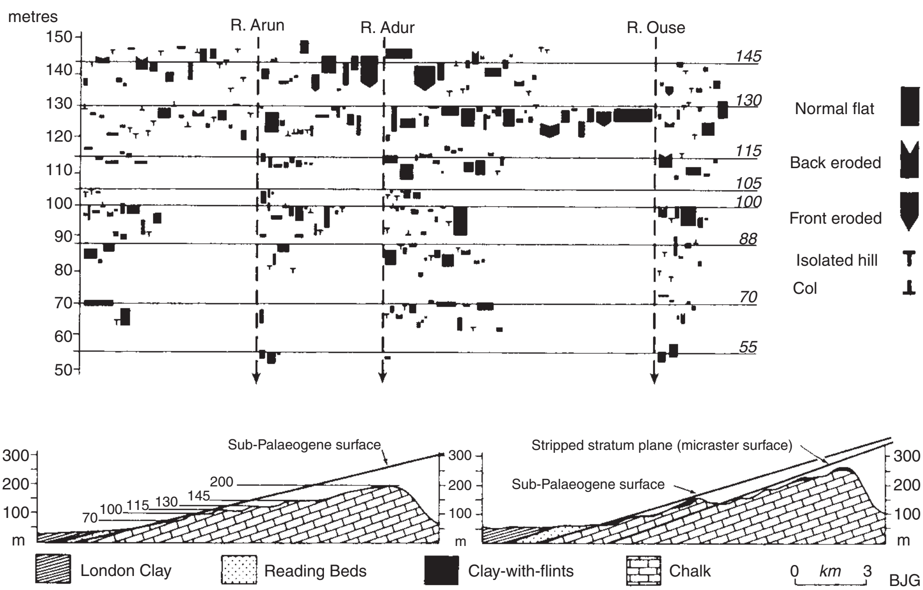

Figure 43 Height-range diagram for erosional levels identified on the South Downs backslopes by Sparks (1949), together with his interpretation of several horizontal marine benches. The two sections compare Sparks's interpretation with that of later workers (see Jones 1981)

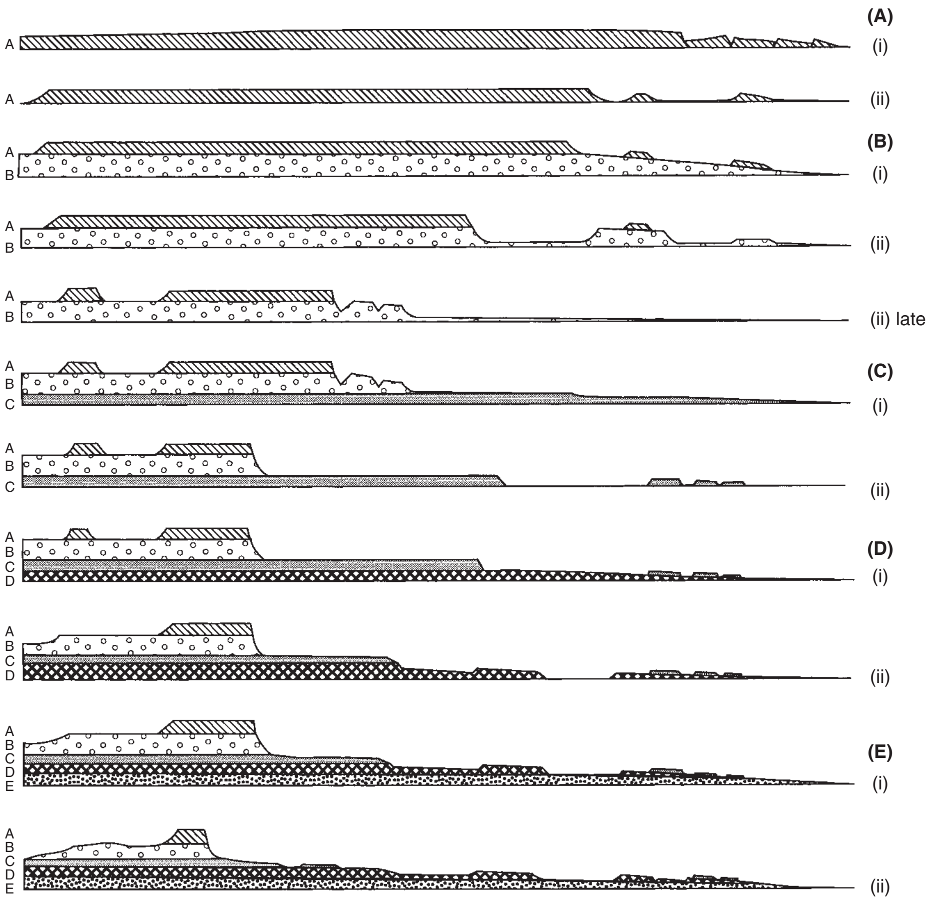

Figure 44 The topographic evolution of the area between the Drakensberg Escarpment and the Natal Coast as detailed by Lester King in 1972. Continental rupture in Jurassic–Cretaceous (A) was followed by pulses of differential uplift in the mid-Cretaceous (up to 1,250 m – B), Miocene (C), end Miocene (up to 800 m – D) and Pliocene (up to 625 m – E), each of which generated incision, backwearing and pediplanation. The result is a stepped and warped landscape in which remnants of the Mesozoic Gondwana Surface, originally at 600 m (A), survive at *c*.3,500 m (E(ii))

model of landscape evolution (Chorley 1965) compared with the approaches of Gilbert, Hack and Penck. Others pointed to the oversimplified theoretical concepts employed and the dangerous over reliance on morphological evidence compared with the far more rigorous, scientific approach adopted by the relatively new but rapidly expanding field of Quaternary Science, with its focus on the analysis of sediments. Yet others pointed to the difficulties of separating base level controls from structural controls and the fundamental problems of correlating surfaces over significant distances simply on the basis of elevation, especially when the origin and age of such 'surfaces' were often largely based on speculation. It came to be recognized that 'flats' or erosional levels could be produced by a wide variety of processes including pediplanation, etchplanation (see ETCHING, ETCHPLAIN AND ETCHPLANATION) and CRYOPLANATION and that contemporary landscapes could contain EXHUMED LANDFORMS, including planes of unconformity that may have been warped.

But, most importantly, the 1960s witnessed the onset of radical changes to prevailing views of the past arising from growing knowledge about global tectonics and Quaternary climate change. The new paradigms indicated 'ceaseless motion' as a characteristic of the lithosphere, together with oscillating climatic conditions and sea levels, and thereby seriously undermined notions of both the formation and survival of surfaces from the distant past, except under special conditions or where LANDSCAPE SENSITIVITY is very low.

As a result of these criticisms, denudation chronology fell into disrepute and almost became a term of abuse. Many of the detailed chronologies came to be dismissed as pure speculation or were demolished after detailed reinvestigation (as in the case of Sparks's (1949) sequence – see Figure 43). However, alternative explanations of landscape development as due to waves of aggression acting on rock sequences offering variable resistance to erosion have proved to be neither edifying nor satisfying. Most landscapes contain conspicuous morphological features (low relief surfaces, bevels, benches) that require explanation and the evolution of topographic landscapes remains a fascination. As a consequence, studies of landscape development have continued, albeit in different form and under new names. Long-term landform/landscape development or evolutionary geomorphology has many of the same aims as denudation chronology but within a much more complex conceptual framework, utilizing analysis of surficial deposits and the ever-growing range of absolute dating techniques. One of the most significant developments has been the attempt to correlate offshore sedimentary sequences with onshore denudation episodes, as pioneered by L.C. King. Some idea of the range of approaches adopted can be gained from Smith *et al.* (1999).

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SEE ALSO: clay-with-flint; dating methods; drainage pattern; duricrust; dynamic equilibrium; erosion; geomorphic evolution; global geomorphology; inselberg; sea level; slope, evolution; tectonic geomorphology

DAVID K.C. JONES

DESERT GEOMORPHOLOGY

The scientific study of deserts and desert landforms had its origins in the latter half of the nineteenth century driven by many external and internal forces including imperialism, colonialism, military adventurism, romance as well as the desire to explore and exploit mineral resources and claim land for agriculture, ranching and grazing.

Some of the earliest descriptive observations on deserts can be found in the works of Greek and Roman geographers (e.g. Herodotus, Aristotle, Seneca, Strabo, Pliny, Ptolemy, among others). Arab and Muslim geographers also wrote about deserts during their various journeys within the Dar-el-Islam which stretched from Morocco to Indonesia. Deserts have also been mentioned in the writings of European travellers that went along the great Silk Road to China (via the deserts of the Middle East and Central Asia) beginning in the thirteenth century (e.g. Marco Polo and others).

The imperial ventures of Spain and Portugal in the New World during the sixteenth century led to some of the earliest descriptive observations of deserts. The Spanish, in the shape of the *conquistadores* and missionaries, were probably the first to venture in the Sonoran and Chihuahuan deserts of North America (e.g. De Vaca, Coronado, De Soto, Father Kino, among others) and provided detailed descriptions of the landforms.

Historical framework I: 1850 to 1950

With colonial aspirations and military adventurism in full swing during the middle part of the nineteenth century, England, France, Germany and the United States sent forth the first wave of soldiers, surveyors and scientists to investigate deserts which lay within their purview (Tchakerian 1995: 2).

With British administration in Egypt, some of the earliest scientific works on desert landforms were conducted from the late 1890s to the late 1920s (see the bibliographies in Goudie 1999) by Cornish, Beadnell, King and Ball, with particular emphasis on general dune forms and processes. The book *Waves of Sand and Snow* by Vaughan Cornish (1914) is one of the earliest detailed studies on bedforms and wind regime. It was Ralph Alger Bagnold (1896–1990), who, during his travels to the Western Desert in Egypt in the late 1920s, became fascinated with dune processes and dynamics and, once discharged from the British Army in 1934, built a wind tunnel in Imperial College, London to study the mechanics of blown sand and dunes. His pioneering research culminated in the now classic book *The Physics of Blown Sand and Desert Dunes* (1941), one of the most significant canons in the geomorphic literature (Goudie 1999: 6). He laid the theoretical foundations of AEOLIAN GEOMORPHOLOGY with his detailed analysis of fluid flow, particle motion, sediment transport and dune forms (particularly barchans and siefs). Bagnold's autobiography, *Wind, War and Sand* (1990), written shortly before his death, is a wonderful journey into the mind and soul of this great scientist, and, along with his earlier monograph of desert travels, *Libyan Sands* (1935), should be read by all desert scholars.

With France gradually extending its colonial empire from Morocco eastwards to Algeria and Tunisia and ultimately to all of the Sahelian countries (Mauritania, Niger, Mali and Chad), the western and central Sahara Desert became the focus of study by French scientists and their African colleagues. The work of the French geoscientists has gone relatively unnoticed outside of western Europe (most have not been translated into English) and some very notable works remain to be read and cited. An excellent synopsis is provided by Goudie (1999). A significant body of works is devoted understandably to aeolian processes and landforms (particularly to dune processes and formation, sand sea dynamics and dune orientations and wind regimes) owing to the fact that some of the world's most extensive sand seas (ergs) (see SAND SEA AND DUNEFIELD) are found in the Western Sahara. Some of the most notable contributors include Rolland, Chudeau, Aufrère (the most prolific), Capot-Rey, Dubief and Clos-Arceduc. Many of their papers were published in the *Travaux Institute de Recherches Sahariennes*, in Algiers. Another significant contributor was Emile Gautier, whose seminal monograph *Le Sahara* (1935) is one of the classic works on the general geomorphology of the Sahara Desert (including the human impact on the environment). This rich tradition of French research in the Sahara has continued in the second half of the twentieth century including works by Birot, Cailleux, Dresch and Tricart.

German colonial interests in south-west Africa (including the Namib and the Kalahari deserts) resulted in a number of expeditions to evaluate the economic potentials for the deserts of southern Africa. The works of the German geographer Passarge are especially significant including his monograph *Die Kalahari* published in 1904. In the deserts of interior Asia, German scientists such as Richthofen were one of the first westerners to recognize the significance of desert dust and LOESS. The Swedish geographer and military opportunist Sven Hedin wrote about the Gobi Desert and its landforms. The Australian arid zone was investigated by a number of explorers and scientists from the mid-nineteenth into the early twentieth centuries, driven largely by ranching and mineral resource evaluation (Cooke *et al.* 1993: 13).

The scientific study of the North American deserts begins with the United States federal exploration of lands west of the Mississippi

River (largely as a result of the acquisition of large swaths of territory beginning with the Louisiana Purchase of 1803 and ending with lands gained from Mexico as a result of the US–Mexican war of 1848). After the end of the US Civil War in 1865, the western surveys (primarily to evaluate the mineral, settlement and railroad prospects) led by King, Hayden, Powell and Wheeler produced some of the first detailed descriptions of the American deserts. John Wesley Powell's 1878 classic monograph *Report on the Lands of the Arid Region of the United States* has extensive descriptions of deserts and desert landforms and was one of the first studies to evaluate the natural resources of the region as well as its suitability for extensive human occupation (something that Powell did not recommend). G.K. Gilbert's influential *Report on the Geology of Henry Mountains* (1877) for the first time showed how process geomorphology (with its foundation in detailed fieldwork and data gathering based on principles from physics, mathematics, statistics) can contribute to the understanding of deserts and desert landforms. Gilbert's contributions to geomorphology are too numerous to cite here (he is considered the 'founder of process geomorphology') but his works on desert fluvial processes, Quaternary lakes and WIND EROSION are particularly noteworthy. In the 1920s and 1930s, Kirk Bryan published numerous influential papers on wind erosion, differential weathering and erosion, pedestal rocks, pediments, and arroyos of the southwestern USA. During the early part of the twentieth century, AEOLIAN PROCESSES were seen as the dominant sculptor of desert landforms, as promulgated by Keyes in 1912 in his 'aeolianation' cycle (Tchakerian 1995: 2). William Morris Davis and his colleagues in a series of papers in the 1930s put to rest the dominating role of wind in the evolution of desert landscapes and accurately pointed out the more substantial role played by weathering, mass movement and fluvial processes. Also in the 1930s, the severe environmental consequences of the 'Dust Bowl' years in the Great Plains of the United States, led many scientists to focus their studies on wind and soil erosion, including those by W.S. Chepil and colleagues, resulting in many publications dealing with the mechanics of aeolian entrainment and transport under different land use activities (Tchakerian 1995: 3).

Historical framework II: 1950 to 2000

Global studies of deserts and desert landforms were rejuvenated in the second half of the twentieth century as a result of a number of technological and theoretical advances. In the following section, some of the major themes are briefly highlighted but detailed consideration and bibliographies in specific desert processes or landforms is left to other contributors to this volume.

Major developments were led by advances, refinement and availability of air photos (after the Second World War) and satellite imagery (during the 1970s), enabling for the first time a continental and global perspective in desert research, including the first comprehensive global dune classification using Landsat imagery (McKee 1979). The emergence of planetary geomorphology has led to more focused attention on desert landforms as terrestrial analogs for arid Mars, particularly in aeolian processes and landforms, canyon formation, sapping and other mass movement processes (e.g. Malin and Edgett 2000).

Increased studies in desert geomorphology in the latter half of the twentieth century have been driven for several reasons: (1) intrinsic fascination as distinct landforms, such as the study of desert PEDIMENTS or STONE PAVEMENTS, ALLUVIAL FANS, sand dunes. Studies in fluvial processes and landforms have been at the forefront of much desert geomorphology research during this period (e.g. Graf 1988); (2) mineral resource potential of desert landforms, such as evaporites (sulphates, nitrates, etc.) from playas, aggregates and groundwater resources in alluvial fans, uranium in DURICRUSTS; (3) desert landforms as indicators for past climatic and ecological change, such as Quaternary lakes and their sediments; (4) the archaeological value of certain desert landforms, such as the study of petroglyphs on rock coatings; and (5) the increased human occupancy and flood hazard risk on or near alluvial fans, ARROYOS, BADLANDS, as well as the environmental hazards associated with increased desert aeolian dust and mineral aerosols and their effects on human health and atmospheric visibility.

The refinement and extension of Bagnold's theoretical foundations have been the primary focus for most aeolian-related research during the past fifty years. This can be seen in the plethora of scientific papers, monographs and symposia proceedings devoted exclusively to aeolian processes and landforms during the past two decades alone

(Tchakerian 1995: 4). Associated with the above refinements were developments in mathematical modelling, computer simulations, and the use of complex, non-linear dynamical systems theory for understanding wind flow and bedforms (e.g. Werner and Kocurek 1999). The resurgence of aeolian geomorphology has also been stimulated by the fact that aeolian sedimentary environments are good analogues for studying hydrocarbon reservoirs in the geologic record. Advances in the understanding of single dune formation and dynamics has now led researchers to tackle the more daunting task of analysing draas (megadunes) and ergs (e.g. Pease and Tchakerian 2002). Other significant developments include the application of luminescence techniques for dating aeolian sands and Quaternary dune systems, and the wide availability of very sophisticated instruments (anemometers, electron microscope, sediment traps) and data loggers for the gathering and analysis of wind data and sediments. Technological advances were also instrumental in heralding numerous studies in aeolian desert dust and mineral aerosols culminating in a series of papers assessing the global distribution of major atmospheric dust sources (e.g. Prospero *et al*. 2002). Research in aeolian processes and landforms has also been driven by concerns for land degradation (DESERTIFICATION) and global environmental change.

The growing scientific interest in desert geomorphology of the past three decades has led to the establishment of many centres of desert research, new journals, international associations, and to UN-sponsored research and publications. This has led to vigorous exchanges of ideas and co-operation among Earth scientists unconstrained by traditional disciplinary boundaries.

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VATCHE P. TCHAKERIAN

DESERT VARNISH

Desert varnish, a paper-thin deposit, drastically darkens the appearance of desert rocks. Any rock type can host desert varnish, so long as its surface remains stable for the thousands of years it usually takes varnish to accrete. Rock varnish is the preferred term where this ROCK COATING occurs in non-desert settings, for example, alpine, Antarctic, Arctic, periglacial, stream, temperate and tropical environments. The term desert varnish is most often used in arid regions.

Like other rock coatings, desert varnish is deposited on rock surfaces and does not derive from the host rock itself. Arrows in the middle image in the middle row of Plate 34 exemplify this discrete contact. Like CASE HARDENING and other rock coatings, many varnishes seen at the surface today actually start in the subsurface in fissures. Varnishes are usually less than 100 μm thick, and even where micro-basins host deposits of a few hundred micrometres, median thicknesses are usually less than 10 μm thick.

Wind does not cause shiny varnish; wind abrades away this relatively soft coating. In fact, the presence or absence of desert varnish is an important clue that a particular desert pavement was not or was made by aeolian deflation. Usually dull in lustre, its occasional sheen comes from a smooth surface micromorphology in combination with manganese enrichment at the very surface of the varnish.

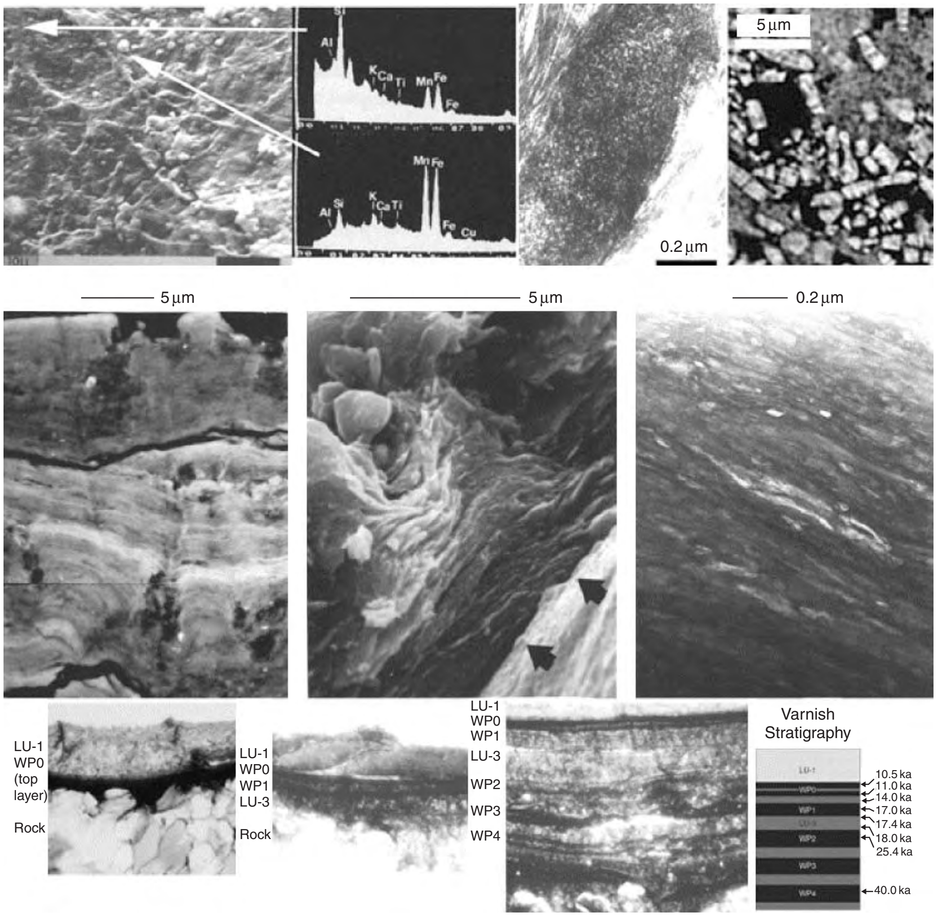

Plate 34 Microscopic views of desert varnish from arid environments. Top row: microscopic evidence for bacterial origin of rock varnish from left to right: secondary electron image of Negev Desert budding bacteria where the bacteria greatly enhanced manganese and iron; transmission electron image of manganese encrusting a bacterial form; and backscattered electron image of bacteria revealed by acid etching. Middle row: layering of desert varnish shown in backscattered (left), secondary (middle), and high resolution transmission electron microscope (right) images. Bottom row: calibration of microlaminations seen in optical microscopic views of ultra-thin sections, where black layers in the varnish represent wet periods that have been calibrated by Tanzhuo Liu's research (Liu *et al*. 2000). The thin sections from left to right show progressively older varnishes with progressively more complex layers from Death Valley, California

Clay minerals are the major ingredient of desert varnish, typically comprising more than half and sometimes as much as 90 per cent. The clay minerals impose the layered structure seen in the middle row of Plate 34. Clays are deposited as dust on rock surfaces, and are then cemented to the host rock by hydroxides and oxides (Potter and Rossman 1979) of manganese (birnessite) and iron (goethite and hematite). Manganese and iron make up about a third of varnish, with typically less than 5 per cent of varnish composed of other components.

The mystery of desert varnish surrounds how to explain the great abundance of manganese, the element that gives varnish its dark brown to black colour. The elemental abundance of manganese in varnish is 50 to 300 times the concentrations found in dust that falls on rock surfaces. Put another way, ratios of manganese to iron are about 1:40 to 1:60 in surrounding soils and dust, but are about 1:1 in varnish.

In the past century, there have been two competing models to explain manganese enrichment. The first was a chemical process favoured by geochemists, whereby naturally acidic rain dissolves manganese in the rock or dust (but not the iron). Then, manganese oxidizes upon exposure to a slightly higher pH. The competing model was a microbial process, whereby bacteria precipitate manganese (Drake *et al.* 1993).

Although bacteria have been cultured from varnish and have made 'artificial varnish' in the laboratory (Dorn 1998), the typically slow rate of varnish growth (on average, about a micrometre per thousand years) makes it very difficult to have confidence that bacteria cultured today in the laboratory make varnish. In fact, the type of gram-positive bacteria most easily cultured from desert varnish today have not yet been identified within varnish layers. To make the matter more difficult, 'biomolecular fossils', such as amino acids generated by these bacteria, exist in both desert varnish and unvarnished weathering materials. Thus, most convincing evidence for a bacterial mechanism is actually seeing manganese enhancement *in situ*. In the upper row of Plate 34, budding bacteria can be seen concentrating manganese and iron.

New high resolution transmission electron microscope evidence (Krinsley 1998) reveals that these chemical and biological models are not truly in competition, but work in tandem. Varnish formation can be explained by a four-step process. Step 1 is the enhancement of varnish (and to a lesser extent iron) by bacteria; the top row in Plate 34 shows manganese-rich sheaths of bacteria. Step 2 is the chemical dissolution of the bacterial sheaths, whereby manganese and iron are broken down into nanometre-sized granules. Step 3 is chemical transport of manganese and iron into clay minerals. Step 4 is the precipitation of unit cells of manganese and iron inside clay minerals. Potter and Rossman (1979) noted that the hexagonal arrangement of oxygens in clay mineral layers form a template for crystallization of the manganese mineral birnessite seen in desert varnish.

Krinsley (1998) shows high resolution imagery revealing all steps in this polygenetic process whereby clay minerals and oxides are co-dependent in varnish formation. Clay provides the overall structure and template for oxide precipitation, while bacteria simply provide a ready source of manganese (and iron) cement. Varnish formation all takes place within a few micrometres of the bacterial source, where the manganese and iron are redistributed with hygroscopic water – all inside layers like those seen in the middle row of Plate 34.

Environmental changes play an important role in the development of desert varnish. Where lichens start to grow, for example, biological acids destroy desert varnish by dissolving the manganese and iron oxides. Where rocks come to exist in a desert pavement, a ground-line band of very thin and shiny varnish forms a circle around a desert pavement clast. But where varnish grows on the tops of boulders less influenced by local environmental changes, regional climatic change plays an important role in varnish formation.

In these boulder-top varnishes wetter climates favour bacterial enhancement, yielding layers that are particularly rich in manganese. Drier climates with more alkaline dust produce layers that are not as rich in manganese. The bottom row in Plate 34 shows desert varnishes from Death Valley, California, where growth of these layers has been calibrated using a combination of numerical dating methods (Liu *et al.* 2000; Zhou *et al.* 2000). Progressively older varnishes show progressively more complex layers. Varnish microlaminations provide archaeologists and geomorphologists with a powerful tool, because they reveal both climatic change and a time signal.

Some of the most interesting aspects of desert varnish surround its minor and trace constituents. Lead, for example, is greatly enhanced in the uppermost micron of the varnish from twentieth-century air pollution. The carbon that is trapped within and underneath varnish shows some potential as a means of radiocarbon dating varnish, but the history of the carbon is usually too complicated to make this technique useful. Mobile trace elements decline progressively over time, as they are leached by hygroscopic and capillary water (Krinsley 1998). Varnish also traps foreign material crushed into rock engravings as a part of religious ceremonies (Whitley *et al*. 1999). New experimental studies of trace isotopes such as ^{7}Be, $\delta^{13}C$ and $\delta^{17}O$ show potential to reveal new insights into this ubiquitous weathering phenomenon. Some planetary scientists believe that desert varnish exists on Mars, and that Martian varnish might preserve active organisms or at least biological fossils such as those seen in the top row of Plate 34.

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SEE ALSO: rock coating

RONALD I. DORN

DESERTIFICATION

Desertification is a term for land degradation caused by adverse human impacts in arid, semi-arid and dry-subhumid areas, together called the 'susceptible drylands' (Middleton and Thomas 1997). Hyper arid areas are in general terms not regarded as sites of desertification, since these areas are extremely desert-like due to natural conditions. The term desertification has also been used more widely, for example to refer to land degradation in non-dryland contexts, and there are over a hundred published definitions. There is widespread consensus that the term should be restricted to susceptible drylands, a view embodied in the 1994 UN Convention to Combat Desertification (UNCCD), which by March 2002 had been ratified by the governments of 179 countries. Desertification occurs not just in the developing world, but also under the impact of the inappropriate or excessive use of agricultural technologies in the dryland areas of the developed world. Overall, desertification relates to the unsustainable use of the land in drylands.

The term desertification is widely regarded to have been coined by Aubreville (1949), who used it to describe the environmental impact of forest clearance in West Africa, leading in his opinion to the creation of an ecological desert. It gained renewed use in the 1960s and 1970s (and was sometimes used interchangeably with desertization) when the major Sahel drought at the time led to a decline in biomass, famine and livestock and human deaths in the Sahel zone along the southern margin of the Sahara Desert.

Desertification has both social and environmental components. Social dimensions relate to the pressures and processes that cause people to carry out activities that degrade the land. Today, international efforts to reduce desertification, such as the UNCCD, lay great emphasis, particularly in the developing world, on addressing the social and economic issues that lead people to carry out land use practices that lead to land degradation. Environmental dimensions of desertification are important too, however, and relate to the actual physical processes of land degradation and to distinguishing impacts of natural dryland environmental variability from anthropogenically aggravated changes to land systems. The latter is important since it is possible to confuse the impacts of dryland rainfall variability with changes in soils and vegetation caused by human actions (Thomas and Middleton 1994). These confusions have led, in the past, to overestimates of the physical dimensions and scale of desertification, including for example, inappropriate statements being made about the advance of deserts over productive land (see Hellden 1991; Thomas 1993). The causes and use of the term desertification are complex, however, since natural drought may itself place pressures on social and agricultural systems that in turn lead to land degradation.

Physical processes of desertification

The physical processes of desertification comprise depletion of and damage to the soil, ground water, and to some extent vegetation systems, that reduces their productive capacity or biological potential. The soil provides the geomorphological context of desertification, with loss of productive potential occurring either through the physical loss of soil by water or wind erosion, or internal physical and chemical changes such as compaction, salinization, alkalinization and nutrient depletion. Dryland soils may be thin or skeletal, with generally slow rates of formation due to the limited availability of water for the weathering of underlying bedrock and also to slow build up of organic material. Exceptions may occur in specific geomorphic locations such as valley floors where seasonal water supply may be greater. Natural recovery from soil erosion or internal changes is therefore likely to be slow.

There has been considerable debate regarding whether changes in dryland vegetation systems that are unaccompanied by soil system changes constitute desertification. These debates have arisen for a number of reasons: many savanna systems are 'non-equilibrium' ecosystems that do not achieve a spatial or temporal steady state because of the natural dynamism of dryland climates (Mistry 2000); dryland vegetation systems can be both resilient and adapted to disturbance and can exhibit rapid recoveries; the impacts of droughts and land degradation on vegetation systems can be hard to distinguish; observed vegetation changes caused by human pressures may not be accompanied by soil system changes (Dougill *et al*. 1999), facilitating recovery if land use pressures are reduced or removed. It should be noted however that vegetation depletion can set the scene for desertification processes to take effect via the processes of erosion.

Geomorphological dimensions of desertification

Desertification via erosion processes may be relatively easy to recognize and in susceptible drylands both wind and water erosion can be important. WIND EROSION potential is greatest in areas of low relief and unconsolidated sediments, for example the Canadian Prairies and the midwest of the USA, as witnessed by the occurrence of severe DUST STORMs during the 1930s, while areas of steeper topography are more susceptible to water erosion, for example in the Highlands of Ethiopia and Kenya and around the Mediterranean basin in Europe and North Africa.

Changes within the soil due to human activities are, with the possible exception of salinization, less visible and more insidious than those caused by erosion. Salinization and alkalinization associated with irrigation schemes are widely cited causes of productivity decline in drylands. Other internal changes relate to waterlogging, also associated with irrigation, and the crusting and compaction of soils, increasingly caused by the mechanized cultivation procedures used in agriculture. Nutrient depletion is a notably widespread but often underestimated form of desertification (Thomas and Middleton 1994). Nutrient loss can be caused by the actual physical removal of soil by erosion but is often a function of the clearance of natural vegetation to create fields for cultivation, the subsequent intensity of cultivation or the lack of application of fertilizers, especially in developing world dryland areas.

Assessing the extent and nature of desertification

There are no readily available means of gaining absolute data on the occurrence of desertification. Earth observation via remote sensing can be useful for detecting dimensions of vegetation change, and for distinguishing natural fluctuations due to rainfall variability from changes caused by human actions (e.g. Tucker *et al*. 1991), but even the highest resolution imagery can be too coarse to identify many elements of soil erosion and unsuitable for determining soil internal changes. Thus estimates of the global extent of desertification are likely to be crude, and sometimes highly erroneous (Thomas 1993). Field studies and modelling approaches are important for local and regional investigations of degradation (Mairota *et al*. 1998). Increasingly, as the social dimensions of land degradation are recognized as vital elements of any efforts to ameliorate the problem, land user understanding and knowledge, often untapped, especially in the developing world, are viewed as essential for an effective understanding of where, why and how desertification occurs (Reed and Dougill 2002).

The UN has been the instigator of the few attempts to assess the global scale of desertification. Estimates produced in the 1970s and 1980s have subsequently been heavily criticized in the scientific community for their lack of methodological rigour and for confusing natural cyclic changes in vegetation systems due to drought impacts with desertification caused by human actions.

The most recent global assessment of land degradation caused by soil degradation (GLASOD) was carried out by the International Soil Reference Centre in the Netherlands on behalf of UNEP in the late 1980s and early 1990s. A GIS was used to analyse data collected through a clearly defined, but somewhat qualitative, methodology (Middleton and Thomas 1997). Despite its flaws, GLASOD has provided a database through which assessments of susceptible dryland soil degradation can be analysed in terms of geographical coverage (see Table 10), contributory degradation processes, and relationships to land use.

GLASOD estimates that in the late 1980s and early 1990s approximately 1,030 million hectares, equivalent to 20 per cent of the susceptible drylands, had experienced soil degradation processes caused by human activities. Water erosion was identified as the major physical process of degradation in 48 per cent of this area and wind erosion in 39 per cent. Chemical degradation (mainly salinization, alkalinization and nutrient depletion) was dominant in only 10 per cent of the area, and physical changes such as compaction and crusting in just 4 per cent. These latter figures may well be underestimated given the problems of identification of these less visible processes. The severity of degradation was described by GLASOD as strong or extreme in 4 per cent of the susceptible drylands – meaning that land where the original biotic functions of the soil have been destroyed, and which are irreclaimable without major restorative measures does not occur widely.

What human actions lead to desertification?

Almost any land use in the susceptible drylands has the potential to lead to desertification, if it is conducted to excess or in locations to which it is not well suited. The literature on desertification widely cites four forms of land use as major contributors to the problem: cultivation, irrigation schemes, livestock production and deforestation.

Overcultivation has sometimes been viewed as the main cause of desertification, especially in areas of the developing world where increasing populations have necessitated attempts to increase yields without the resources available for additional fertilizers. Nutrient depletion is therefore a potentially serious issue, with the World Bank attributing declining yields of staple crops in Sahel nations and in parts of South America to this problem. Attempts to increase crop yields through mechanization can lead to soil compaction, increasing runoff and erosion under intensive dryland rainfall events. Aeolian deflation can also be exacerbated by the removal of shelter belts to allow large machinery to be used.

Irrigation systems, whether by canals from storage dams or through centre pivot systems, can cause desertification through waterlogging, salinization and alkalinization, the excessive accumulation of sodium in the soil. High evapotranspiration rates in drylands mean that excessive irrigation can lead readily to the accumulation of salts in the soil. Waterlogging arises when irrigation rates are so high that the water table is raised excessively.

Table 10 Dryland soil degradation (million ha) by continent according to GLASOD

	Susceptible dryland area	Light and moderate desertification	Strong and extreme desertification	Total desertified
Africa	1,286.0	245.3	74.0	319.3
Asia	1,671.8	326.7	43.7	370.4
Australasia	663.3	86.0	1.6	87.6
Europe	299.7	94.6	4.9	99.5
North America	732.4	72.2	7.1	79.3
South America	516.0	72.8	6.3	79.1
Total	5,169.2	897.6	133.7	1,035.2

Source: Derived from data in Middleton and Thomas (1997)

Many dryland regions may be better suited to extensive livestock production than to cultivation. Pastoralism is seen as a contributor to desertification when it is over intensive, which may occur in developed world drylands and in the developing world when traditional practices are replaced by commercial systems. What may result is excessive grazing that can alter plant community composition and reduce overall plant cover, thereby increasing the susceptibility of the land to erosion processes. The role of pastoralism in desertification is however complex and somewhat controversial: as noted earlier vegetation changes may not always be accompanied by soil system changes and therefore may be reversible.

Deforestation in drylands is associated both with the clearance of lands for cultivation and, in developing world contexts, with the collection of wood for use as the dominant domestic fuel. The greatest threat deforestation offers for desertification is in areas of steep slopes. For example, during the twentieth century a tenfold reduction in woodland cover occurred in the Ethiopian Highlands through the effects of expanding cultivation and fuelwood collection: it is not surprising that soil erosion is a major problem in this area.

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DAVID S.G. THOMAS

DESICCATION CRACKS AND POLYGONS

As evaporation of water from a saturated, fine-grained, cohesive sediment occurs, volume reduction may be accompanied by sufficient tensional stress for rupture to take place. Cracks are thereby formed and may display polygonal patterns. The morphology of the rupture patterns depends both on material properties (structure, degree of packing, moisture content, etc.) and on environmental conditions (temperature, humidity, rate of drying, etc.) (Corte and Higashi 1964). Cracks and polygons of this type are common geomorphic phenomena of both periglacial and arid environments (Maizels 1981), of vertisols and of drained proglacial lakes (Huddart and Bennett 2000).

Cracks tend to be fairly straight or smoothly curved in plan. However, their patterns, lengths, depths, widths and number show great variation. The plan form of blocks between cracks, which is determined by the crack pattern, can be flat, convex, concave or irregular, but the size of blocks tends to increase with the thickness of the material of which they are composed. The number of cracks is generally inversely proportional to sediment size, being greatest in materials rich in silt and clay. Mean crack length is proportional to sediment moisture content. It tends to decrease with time, because new, short cracks continue to be formed during each new cycle of drying. The spacing of cracks may increase with the rate of desiccation and with the proportion of clay present and according to the nature of the clay type. Sediment that is montmorillonite rich, for example, is prone to greater contraction than one with a comparable proportion of kaolinite. The presence of stones in and on fine-grained sediments may also affect the nature of the cracking.

Lachenbruch (1962) identified two common crack patterns. One is an orthogonal pattern in which cracks meet at right angles. The other is a non-orthogonal pattern characterized by triradial intersections that form obtuse angles of around 120°. The former are probably characteristic of inhomogeneous or plastic media in which stress accumulates gradually. Cracks form first at loci of low strength or high stress concentration. Because the cracks do not form simultaneously, a new crack has a tendency to join a pre-existing one orthogonally. The latter systems form in more homogeneous, relatively non-plastic media which are dried uniformly.

Particularly large 'giant desiccation cracks' are especially common on salty, playa surfaces (Neal *et al*. 1968) and may result not only from desiccation, but also because of such factors as salt mobilization, subsidence caused by groundwater withdrawal and seismic activity. The sudden creation of giant fissures can damage engineering structures (Al-Harthi and Bankher 1999; Corwin *et al*. 1991).

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A.S. GOUDIE

DEW POND

Dew ponds are closed depressions, often filled with water, usually associated with chalk downland in the southern UK. The constant supply of water to these depressions on permeable substrates and frequently close to the top of slopes has caused considerable discussion over the past two centuries. As the name indicates, some writers have suggested dew (or mist or fog) as the source of replenishment but others believe that rainfall and surface runoff provide sufficient supply. Many dew ponds were dug on agricultural land, occasionally by people employed specifically for this purpose who moved from farm to farm, and the ponds were lined with clay, straw and more recently cement. Today some are dry because of lack of maintenance of the lining. There is considerable confusion in assigning origins to closed depressions in calcareous areas (see KARST) because some are natural DOLINEs or sink holes while others have anthropogenic origins as dew ponds or as pits dug for chalk, marl or clay. Some even have complex origins, forming initially as a doline but subsequently being lined and used as a dew pond.

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IAN LIVINGSTONE

DIAGENESIS

All the changes (physical, chemical and biological) undergone by a sediment after its initial deposition, exclusive of weathering and metamorphism (and incorporates processes such as reworking, authigenesis, replacement, leaching, hydration, bacterial action, and the formation of concretions). The term was coined by Gümbel in 1868 (*diagenese*), though no universal definition exists.

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SEE ALSO: lithification

STEVE WARD

DIAMICTITE

Diamictite (paraconglomerate, mixtite, diamixtite) is a non-genetic term for sedimentary rock consisting of sand and/or larger particles resting in a muddy matrix (Flint *et al*. 1960). Its unlithified counterpart is known as diamicton, and diamict is a general term comprising both consolidated and non-consolidated deposits.

Diamictites consist of a wide range of particles with matrix dominating, giving an overall appearance of clasts chaotically dispersed through structureless or laminated mud. Clasts are dropstones sporadically deposited on the soft substrate due to ice rafting or volcanic explosions. They are angular to subrounded, may be slightly imbricated (see IMBRICATION) and often come from remote sources hundreds of kilometres away. If lamination is present, drapes (see DRAPE, SILT AND MUD) around clasts caused by post-depositional compaction (see COMPACTION OF SOIL) occur frequently.

The origin may be by GLACIERs, DEBRIS FLOWs or turbidity currents. Most diamictites are interpreted as lithified tills (tillites) or GLACIMARINE deposits of pre-Quaternary glaciations (Schermerhorn 1974; Hambrey and Harland 1981) because of a range of features resembling tills known from modern glacial environments, such as lack of sorting, grain sizes from clays to blocks, clasts bearing GLACIAL EROSION features (e.g. STRIATIONs and polishing), glacidynamic structures (e.g. shear planes (see SHEAR AND SHEAR SURFACE) and folds), and mixed lithological components corresponding to ice flow paths. Glacial origin of most diamictites is also supported by intimate association with striated bedrock surfaces, varved clays and lithified GLACIFLUVIAL deposits, arctic fauna, periglacial (see PERIGLACIAL GEOMORPHOLOGY) structures and the correspondence to polar regions of the past.

Diamictites of glacial and glacimarine origin are known from many regions of the Earth (Miller 1996: 469–483). The Late Palaeozoic Dwyka Formation occupies extensive areas in southern Africa and consists of facies deposited mainly under disintegrating ice shelves, near the grounding line and in FJORDs by tidewater glaciers (Visser 1991). Also the Lower Proterozoic Gowganda Formation, which covers more than 12,000 km^2 on the Canadian Shield is possibly a glacimarine diamictite (Eyles *et al*. 1985). Terrestrial tillites are known from the Upper Ordovician of north-west Africa (6–8 million km^2), Upper Proterozoic of western Mauritania and north Norway, Permo-Carboniferous of the Transantarctic Mountains, and Upper Palaeozoic of Oman and Brazil.

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JAN A. PIOTROWSKI

DIAPIR

Vertical intrusions, bulbous or cylindrical in shape, resulting from the upward movement of mobile materials, such as salt (halite), magma, mud and ice, which lie beneath more competent strata (see MUD VOLCANO).

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A.S. GOUDIE

DIASTROPHISM

A general term for the various types of tectonic processes that change the level, position and altitude of the Earth's surface. It is derived from the Greek word *diastrophos*, which means 'turned', 'twisted' or 'distorted'. There are five classes of diastrophic movement (Chorley *et al.* 1984: 98): (1) orogenesis; (2) epeirogeny; (3) isostasy; (4) igneous (including volcanic); and (5) eustasy.

Reference

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A.S. GOUDIE

DIATREME

Vents and pipes which have been injected through sedimentary strata by the explosive release of magmatic gases and filled with the products of the eruption, and fragments torn from the side of the pipes. Kimberlite pipes are examples of diatremes. Some maars, such as those in Germany, are lakes that are the surface expression of diatremes. They can be rich in environmental information (Narcisi 1996).

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A.S. GOUDIE

DIGITAL ELEVATION MODEL

A numerical description of the ground surface is helpful in addressing many geomorphological problems. Continuous topography may be quantified by a digital elevation model (DEM), any spatial array of terrain heights but most commonly a square mesh, or grid. DEM point spacing, or horizontal resolution, varies from fine (1 m) to coarse (⩾5 km), depending on the application, required level of detail, and the limitations of computer storage and processing (Table 11).

DEM-related nomenclature can be inconsistent. Digital terrain model (DTM), a frequent synonym for DEM, also is applied loosely to any calculated result, such as a map of slope gradient, rather than to the input heights themselves. Digital terrain *modelling*, confusingly also DTM, widely denotes DEM processing or GEOMORPHOMETRY.

For efficient computation and display, terrain heights are arranged in a defined structure (Figure 45), usually a regular grid, a triangulated irregular network (TIN), or digitized height contours or slope lines normal to contours. DEM structures are compromises, each with advantages and drawbacks.

Square-grid, actually rectangular, DEMs (hexagons or equilateral triangles are rare) store height Z as an array of implicit longitude X and latitude Y co-ordinates (Figure 45A). While this regular and discretized (discontinuous) structure is ill matched to the varied intricacies of surface features, a grid optimizes algorithm development, data processing and registration with spacecraft images. Grid DEMs can adapt somewhat to complex terrain by recursively subdividing squares, but computational efficiency declines.

The irregularly distributed heights of a TIN (Figure 45B) are vertices of triangles that vary in shape and size. A TIN is interpolated directly from surveyed points or discrete features that are extracted manually from maps or by computer from a grid or contour DEM. The TIN is adaptive, or surface specific: it aligns with ridges and channels and has many heights in complex areas but few redundant heights in planar terrain. Offsetting these advantages is the storage and processing burden imposed by the explicit X,Y co-ordinates required for each value of Z.

Ground-surface form is neither rectilinear nor triangular. Although not all topography is fluvial, the DEM structure best reflecting processes of erosion and deposition mimics paths of steepest gradient (Moore 1991). Terrain heights at intersecting contours and interpolated slope lines (Figure 45C) comprise an adaptive DEM that defines quadrilateral units of varied size and shape – most significantly the hillside concavities followed by surface flow. Explicit X,Y co-ordinates are necessary.

Early DEMs were created by field survey, manual interpolation of topographical maps, or semi-automated tracing of contours coupled with computer interpolation. Photogrammetric profiling and later optical scanning and automated interpolation of contours replaced these techniques. Grid DEMs now are available for the Earth, its seafloor, and the planet Mars. GTOPO30, compiled from many contour maps from several sources (Gesch *et al.* 1999), covers Earth at 30′ (nominally 1 km) resolution; two older global DEMs, ETOPO5 and TerrainBase,

Table 11 Sources and applications of Digital Elevation Models (DEMs)

DEM spacing	Sources of height measurements	Some geomorphological applications
1–50 m	Contours and stream lines from airphotos and topographic maps at scales 1:5,000 to 1:50,000 Surface-specific heights and stream lines from ground survey by GPS Remotely sensed heights from airborne and spacebome photogrammetry, radar and laser altimetry	Detailed terrain parameterization and visualization Estimates of flood inundation, soil moisture and other distributed-parameter hydrological modelling Spatial analysis of terrain and soil properties Terrain-aspect corrections to remotely sensed data Effects of terrain aspect on patterns of solar radiation, evaporation and vegetation
50–200 m	Contours and stream lines from airphotos and topographic maps at scales 1:50,000 to 1:200,000 Surface-specific heights and stream lines digitized from topographic maps at 1:100,000 scale	Broader scale distributed hydrological modelling Geomorphometric regionalization and analysis Sub-catchment analysis for lumped-parameter modelling and assessment of biodiversity
0.2–5 km	Surface-specific heights and stream lines digitized from topographic maps at scales 1:100,000 to 1:250,000 N.B.: Coarse-scale DEMs often are compiled by local averaging of fine-scale data	Height-dependent representations of surface temperature and precipitation Effects of terrain aspect on precipitation and surface roughness on wind Mapping continental drainage divisions
5–500 km	Surface-specific heights digitized from topographic maps at scales 1:250,000 to 1:1,000,000 (see also, N.B., above) National archives of trigonometric points, bench marks and other ground-surveyed terrain heights	Modelling relation of erosion and sediment distribution to tectonism Orographic barriers for general circulation models Broad-scale height and shaded-relief base maps for non-topographic information

Source: Modified after Hutchinson and Gallant (2000)

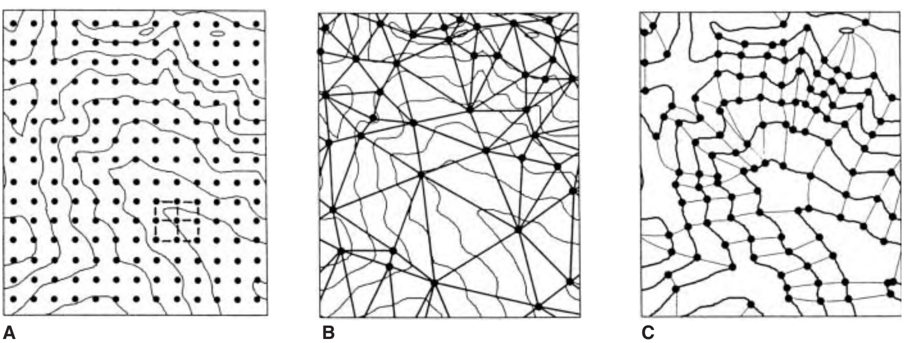

Figure 45 Three contrasting DEM structures for part of a small watershed in California. Dots are height locations. A: square grid, showing 3×3 subgrid used for neighbourhood operations; B: triangulated irregular network, TIN; C: intersections of 20-m contours (heavy lines) with slope lines (light). Each panel is 780 m across

are spaced at 10 km. The US National Elevation Dataset (NED) is a seamless 1″ (30 m) DEM (Alaska at 2″) assembled from all 55,000 1:24,000- and 1:63,360-scale topographical maps. Japan and the UK are gridded at 50 m, Italy nominally at 230 m, Australia 9″ (250 m), and Germany and other countries at various resolutions. Distribution of most military DEMs, e.g. DTED for the USA is restricted.

Bypassing contour maps, remote sensing (see REMOTE SENSING IN GEOMORPHOLOGY) can generate DEMs from direct measurements of terrain height (Maune 2001). Technologies include digital photogrammetry, the Global Positioning System (GPS), laser-ranging altimetry (LiDAR), synthetic-aperture radar interferometry (InSAR or IfSAR), thermal emission and reflection radiometry (ASTER), and, for bathymetry, deep-towed SONAR. The 3″ (90 m) DEM compiled from the 2000 Shuttle Radar Topography Mission (SRTM) includes 80 per cent of Earth's land surface (www.jpl.nasa.gov/srtm/). A variably spaced (1–12 km) depth grid, devised from radar altimetry inverted to sea-surface gravity and thence to bathymetry, covers Earth's entire seafloor.

Random and systematic flaws degrade DEM quality, defined by horizontal and vertical accuracy and precision of the constituent heights. Much of the error in DEMs derived from contours originates in the maps themselves, which never were intended to supply data of the high quality desirable for geomorphometry. Because contour maps only approximate terrain, just as DEMs approximate the maps, declared levels of DEM quality are merely statistical; locally, accuracy can be low. Heights expressed in integers can have insufficient vertical precision, or resolution, especially in level terrain. Contour-to-grid processing, always a compromise, is a second source of error. Some interpolation algorithms overrepresent contours; others add spurious terracing, closed depressions, and linear artefacts. Nor do advanced methods assure DEM quality. InSAR, LiDAR, and other remotely sensed data all contain errors, some of them severe, that are unique to their technologies. Where 1″ SRTM data reflect the dense tree canopy, they reproduce the ground surface no more faithfully than the 1″ NED.

Most DEMs must be refined for subsequent analysis (Figure 46). Computer processing can create a TIN or grid DEM from scattered heights and convert from one data structure or map projection to another. A DEM may be interpolated to finer – within limits, to avoid creating artefacts – or coarser resolutions for compatibility in merging DEMs and registering terrain height to other data. Bulk processing or point-by-point editing corrects errors or replaces parts of a DEM with data of higher quality. Removing long- or short-wave variation from a DEM by digital filtering can enhance detail or subdue erroneous artefacts of production. In creating a grid DEM intended for hydrological application, a drainage-enforcement algorithm reduces digitizing errors and preserves continuity of streams by incorporating channels, ridge lines, and other slope discontinuities (Hutchinson and Gallant 2000).

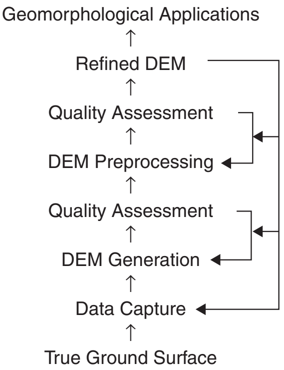

Figure 46 Sequence of activities in preparing a DEM to address geomorphological objectives; modified after Hutchinson and Gallant (2000)

Suitably preprocessed, grid DEMs are used to describe continuous topography through a spatial calculation adopted from digital image-processing, the neighbourhood operation, wherein a result – for example, relief shading – is obtained from adjacent input values. The input from a grid DEM is a compact matrix of heights, usually 3 × 3, moved through the data in regular increments; calculations for TINs or contour DEMs are on triangles or quadrilateral facets. Neighbourhood operations characterize terrain in three overlapping domains – RELIEF (Z), spatial (X,Y) and three-dimensional (X,Y,Z).

Most MORPHOMETRIC PROPERTIES derived from DEMs are moment statistics of height Z and its first two derivatives, slope gradient and profile curvature. Spatial parameters of terrain pattern and texture, unreferenced to an absolute datum, are more abstract; common X,Y measures are aspect, the compass direction faced by a slope, and contour curvature. Processing DEMs in the X,Y,Z domain captures the most complex properties – roughness, intervisibility, and variance of relief with azimuth.

Calculations on DEMs both visualize and parameterize the ground surface (Table 11). Digital maps in colour or monochrome portray topography, often in oblique perspective, by shaded relief, slope gradient, or aspect. Multispectral data, symbols for GEOMORPHOLOGICAL MAPPING, and other types of information are commonly displayed as overlays on a base of contoured height.

Among important parameters are the eight DEM derivatives calculated across each continent from the GTOPO30 data (Verdin and Greenlee 1998): a hydrologically integrated DEM, slope gradient and aspect, streamflow direction and accumulation, the topographical wetness index, stream networks, and drainage basins. DEM parameters are used to quantify hillside form (see HILLSLOPE, FORM), map landslide susceptibility, conduct TERRAIN EVALUATION, devise LAND SYSTEMs, assess cross-country trafficability, plan military operations, describe remote submarine and extraterrestrial surfaces, model slope evolution, simulate WATERSHED hydrographs, forecast the extent of flooding, and estimate sediment delivery.

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SEE ALSO: allometry; applied geomorphology; complexity in geomorphology; cross profile, valley; engineering geomorphology; equilibrium slope

RICHARD J. PIKE

DILUVIALISM

A form of CATASTROPHISM in which it is believed that the landscape was shaped by Noah's Flood, as reported in the book of Genesis. Before the true origin of glacial deposits was recognized, such materials, called 'drift' were ascribed by workers such as Buckland (Davies 1969) to a great deluge, when 'waves of translation' covered the Earth. By the 1830s the recognition of the complex stratigraphy of the drift and the discovery of the importance of the Ice Age greatly weakened the diluvial viewpoint.

The term diluvial is still sometimes used in the context, for example, of supposedly water-lain loess deposits.

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A.S. GOUDIE

DISSOLUTION

In the geomorphological context, dissolution is the process whereby a rock, or parts of a rock, combine with water to form a solution. As the rock dissolves the different minerals disintegrate into individual ions or molecules and these diffuse into the solution. Hence, study of dissolution must focus on specific minerals as opposed to the rocks that are made up from them. Dissolution of a mineral is congruent when all components dissolve together (i.e. no solid remains) and incongruent where only a part of the components dissolve (for example the alumino-silicate minerals where ions are released in reaction with water but retain most of their elements in re-ordered solids such as kaolinite). There is a very wide range of mineral solubility in water, from gibbsite which is virtually insoluble (0.001 mg l^{-1} at pH 7) through to halite (360 000 mg l^{-1} at pH 7). Rocks made up of minerals with a very low solubility are highly resistant to CHEMICAL WEATHERING, while rocks containing highly soluble minerals such as rock salt are only found at outcrop in the driest places. Between these two extremes are a group of rocks in which dissolution along groundwater flow paths leads to the development of concentrated underground drainage and a landform assemblage known as KARST. Karst develops on silicate and evaporite rocks but is most common on the carbonate rocks, limestone and dolomite. Hence the remainder of this entry discusses the carbonate dissolution process.

The solution chemistry of carbonates is relatively simple as only two major minerals, calcite ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$), are involved. Both are only slightly soluble in pure water (*c.*14 mg l^{-1}) and the solvent action of natural waters depends on their acid content. Organic and mineral acids may be important in some localities, particularly during the earliest (inception) phase of karstification (Lowe *et al.* 2000) but dissolution of calcite and dolomite is generally dominated by carbonic acid produced by hydration of dissolved carbon dioxide. It is frequently stated that the reaction between carbonic acid and 'insoluble' limestone produces calcium bicarbonate which is soluble. However, this is incorrect as there is no evidence for the existence of calcium bicarbonate molecules in solution. In fact there are three elementary chemical reactions in the dissolution of calcite which proceed in parallel:

$$CaCO_3 + H^+ \leftrightarrow Ca^{2+} + HCO_3^- \quad (1)$$

$$2CaCO_3 + H_2CO_3 \leftrightarrow 2Ca^{2+} + 2HCO_3^- \quad (2)$$

$$CaCO_3 + H_2O \leftrightarrow Ca^{2+} + HCO_3^- + OH^- \quad (3)$$

These can be summarized into:

$$CaCO_3 + H_2O + CO_2 \leftrightarrow Ca^{2+} + 2HCO_3^- \quad (4)$$

Similar processes take place in the dissolution of dolomite and are summarized as:

$$CaMg(CO_3)_2 + 2H_2O + 2CO_2 \leftrightarrow Ca^{2+} + Mg^{2+} + Mg^{2+} + 4HCO_3^- \quad (5)$$

The reactions continue until the forward and reverse rates become equal at which point the system is in equilibrium and the solution is said to be saturated with calcite. Addition of any acid to the system will increase the concentration of hydrogen ions and displace the equilibria in a forward direction. This reduces the concentration of ${CO_3}^{2-}$ and permits more $CaCO_3$ to dissolve so that when equilibrium is re-established the saturated solution has a higher calcium concentration.

In contrast to mechanical erosion processes, these reactions may occur in static water as well as through the range of water velocities. The speed of the reactions, and the amount of mineral dissolved, are controlled by the detailed solution kinetics (discussion of which is beyond the scope of this entry). However, the role of four important factors: carbon dioxide concentrations, temperature, equilibrium conditions and mixing corrosion will be considered briefly.

1. *Carbon dioxide concentrations* The atmospheric concentration of carbon dioxide is close to 0.035 per cent which would yield a saturation value of 70 mg l^{-1} at 10 °C under open system conditions. Observed concentrations are frequently higher and it is generally assumed that this is due to the biogenic carbon dioxide in the soil atmosphere. However, the fluctuations in soil carbon dioxide concentrations are frequently more pronounced than those of calcium concentrations at springs and it is possible that ground air carbon dioxide in the subcutaneous zone may provide a relatively stable source.
2. *Temperature* For any fixed carbon dioxide concentration in a gas mixture in contact with water and rock the calcite solubility decreases with increasing temperature at a rate of approximately 1.3 per cent per degree Celsius. However, this effect is usually less significant than carbon dioxide concentrations in the gas phase and reaction rates, both

of which broadly increase with temperature. In addition, regional runoff variations account for a greater proportion of the observed variability in solutional erosion rates than do solute concentration variations.

3 *Equilibrium conditions* The two principal equilibrium conditions under which limestone may be dissolved are the 'open' system in which gas, water and rock are all in contact together such that carbon dioxide is available to replace that used up in the reaction of limestone and carbonic acid, and the 'closed' system in which gas and water come into equilibrium but the gas supply is cut off before contact with rock. Since there is no replacement of carbon dioxide under closed system conditions, the amount of limestone which can be dissolved is less than under open system conditions.

4 *Mixing corrosion* The mixing of two saturated waters produces an unsaturated (aggressive) solution and the mixing of a saturated and an aggressive solution, or of two aggressive solutions, may result in increased aggressivity. In extreme cases the new solution may be capable of dissolving 20 per cent more calcite but 1–2 per cent is more usual in natural waters. Hence, the mixing effect is generally less effective than 'normal' solution and its importance lies in its ability to operate in conditions under which normal solution is impossible, such as narrow fissures and in the phreatic zone.

While the key role of carbonic acid in the dissolution of carbonates was understood by the end of the eighteenth century it was not until towards the end of the twentieth century that details of the equilibrium chemistry of carbonate waters and their importance for speleogenesis and landscape evolution were elucidated, most notably by Dreybrodt (1988, 2000), Palmer (1991) and White (1984).

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SEE ALSO: corrosion

JOHN GUNN

DIVERGENT EROSION

The term was derived from the evolution of INSELBERGS. When a rock outcrop is laid open in the tropics the rainwater runs off very fast. Thus weathering lacks the moisture for further decomposition. In the surrounding zone the water can percolate into the soil and guarantees continued weathering of the rock there. During the contemporaneous lowering of the surface, the rock outcrop is resistant. Often its sides are further exposed, whilst the neighbouring areas are eroded. Mostly this occurs at a similar rate to that at which the rock outcrop at the base of the regolith is decomposed. Thus the rock grows out relatively slowly from the weathering mantle. Therefore diverging erosion follows diverging weathering. Inselbergs rise up to 300 m and even more. They are developed, however, in all sizes, so that sequences are observed in the tropics. The above derivation developed from these observations. Isolated mountains outside of the tropics are as a rule palaeoforms, sometimes exhumed.

The initial stage of the exposure of the rock outcrops has different causes: thinning of the soil cover occurs in special geomorphic positions, quite often in an area where an ESCARPMENT is originating, i.e. where planational processes produced a slight increase in gradient. The first rock outcrops then evolve into inselbergs which lie in front of or on top of escarpments. Eventually rock outcrops are so numerous that an escarpment develops. Sometimes inselbergs occur near rivers or sea coasts. Rapid downwearing produces an initial small rock outcrop by chance, e.g. by tree fall. Other positions for inselbergs are watersheds, large and small. Examples are the prominent inselbergs in the south of Central Australia: Ayers Rock, the Olgas, and Mt. Connor. All these are developed in sedimentary rocks, which shows that divergent weathering is independent from rock hardness as the lithology is more or less the same in these rocks and the neighbouring plain, at least in areas larger than the inselbergs. Quite often a different spacing of fractures is postulated as a reason for special resistance, but tectonic lines should show repetitive patterns and thus a regular spacing of inselbergs, which is not the case. Divergent erosion is sometimes used for different processes controlled by rock hardness, too, especially in the case of inselbergs. But it is always hard to prove this as rock samples for comparison from the deeply weathered plain are difficult to retrieve.

Rock outcrops are generally resistant in the tropics due to the minor importance of physical weathering and the overwhelming power of chemical processes. The first needs water, if at all, only for a short time, while the second can only work with long wetting, preferably with water containing organic acids. Both are missing on bare slopes. Therefore inselbergs and escarpments in the tropics are often very old. After exposure these rock outcrops are only slightly weathered. Even if special forms like runnels, exfoliation sheets, or small caverns developed, the overall form is not changed. At Ayers Rock these weathering forms are nested. Thus they are of different ages, which proves the stability of the slope during a very long time even under changing climates.

Very steep slopes in Sri Lanka are surprisingly stable. This is not only due to rock outcrops but to the rapid movement of the subterraneous water. For this, soil analysis gives an explanation: in thin sections a very high volume of large pores is seen, as is the relatively stable soil texture due to iron and silica minerals in the matrix or even as cutans on pore walls. Thus the internal water movement has good pathways. Soil stability is maintained due to the low swelling capacity of the kaolinite minerals. The rapid water movement is similar to that on rock outcrops. Thus these processes were called internal divergence. Once the soil fabric is disturbed, e.g. by building a street on a steep slope, water movement is blocked and severe slides may occur.

Divergent erosion as a principle in tropical geomorphology shows a discontinuity of erosion in space and time. It is nearly independent of the forces of gravity but dependent on differences in friction. One can consider divergent erosion as a positive feedback mechanism. A threshold of resistance to weathering and erosion is not surpassed in the case of the exposed rock facets. Thus they possess an extremely low sensitivity to change. The ergodic principle is applicable in the humid tropics where planational lowering is still active to different heights on the periphery of inselbergs due to divergent erosion.

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H. BREMER

DOLINE

Dolines are natural enclosed depressions found in karst landscapes (Ford and Williams 1989). They are subcircular in plan, tens to hundreds of metres in diameter, and can range from a few metres to about a kilometre in width. They are typically a few metres to tens of metres in depth, but some are hundreds of metres deep. Their sides range from gently sloping to vertical, and their overall form can vary from saucer shaped to conical or even cylindrical. Dolines are especially common in terrains underlain by carbonate rocks, and are widespread on evaporites. Some are also found in siliceous rocks such as quartzite. Dolines have long been considered a diagnostic landform of KARST, but this is only partly true. Where there are dolines there is certainly karst, but karst can also be developed subsurface in the hydrogeological network even when no dolines are found on the surface. Dolines have a similar function in karst landscapes to the drainage basin in non-karstic lithologies, in that they drain rainwater from the surface, but in the case of the doline it is discharged underground via an outlet at the lowest point in the doline basin.

The term sinkhole is sometimes used (especially in North America) to refer both to dolines and to depressions where streams sink underground, which in Europe are described by separate terms (including ponor, swallow hole, and stream-sink). Thus the terms doline and sinkhole are not strictly synonymous. Table 12 lists the terms employed by different authors and Figure 47 illustrates six main doline types (both are from and are discussed in more detail in Williams 2003).

Enclosed depressions in karst can be formed by four main mechanisms: DISSOLUTION, collapse, SUFFOSION and regional SUBSIDENCE. In practice the complexity of natural processes often results in more than one mechanism being involved, in which case the doline is polygenetic in origin. A typical case is a depression formed initially by dissolution that later in its development is subject to collapse of its floor into an underlying cave. In such a case, the gentler upper slopes of the doline were formed by dissolution and the steeper lower slopes by collapse.

Solution dolines

The bowl-shaped form of a typical doline indicates that more material has been removed from its centre than from around its margins. Where the principal process responsible for this is dissolution of the bedrock, it follows that there is a mechanism that focuses chemical attack. The amount of limestone that can be removed in solution depends upon two variables: first, the concentration of the solute and, second, the volume of the solvent (in this case the amount of water draining through the doline). Variations in either or both of these variables could be responsible for the focusing of dissolution near the centre of the depression, but if local variation in solute concentration alone were sufficient to explain the occurrence of solution dolines, then they would be found on every type of limestone in a given climatic zone. This is not the case, as illustrated by comparison of landscapes formed on Devonian, Carboniferous, Jurassic and Cretaceous limestones in England, where dolines are most frequently found on Carboniferous limestones and tend to be less prevalent on Cretaceous and Jurassic limestones. It follows, therefore, that local spatial variations in water flow must be responsible for focusing corrosional attack.

The development of dolines of all kinds depends on the ability of water to sink into and flow through karst rocks to outlet springs. The exposure of limestones by erosion provides an input boundary for infiltration of water and a valley incised into the limestone provides an output boundary. Infiltrating rainwater is acidified in the atmosphere and further acidified in the soil. On percolating downwards this water accomplishes most of its dissolutional work within 10 m of the surface. Joints (see JOINTING), faults and bedding-planes vary spatially within the rock because of tectonic history and variations in lithology. Consequently the frequency and interconnectedness of fissures available to transmit flow also varies. Some fissures are more favourable for percolation than others, for example where several joints intersect, and as a result these develop as principal drainage paths. Water flows towards them and as a result they are subjected to still more dissolution by a positive feedback mechanism and so vertical permeability is enhanced. The local surface of water saturation in the EPIKARST is drawn down over the preferred leakage paths similar to cones of depression in the water table over pumped wells; streamlines adjust and resulting flow lines are centripetal and convergent on the preferred drainage zones. By this means solvent flow is focused and, as the surface lowers, the more intensely corroded zones begin to obtain topographic expression as solution dolines. Particularly large solution depressions often occur in the humid tropics where corrosion processes were uninterrupted by Pleistocene glaciations. In these places the term cockpit is sometimes applied to them after a particular style of landscape in Jamaica, where depressions are incised between intervening conical hills.

Although small solution dolines have formed in 15,000 years or so in some mid to high latitude areas that were glaciated in the late Pleistocene, several tens to hundreds of thousands of years are required to develop large solution dolines in limestone. Once formed they may persist in the landscape for several million years provided there is sufficient thickness of limestone for their continued incision. Individual dolines may merge to form compound closed depressions (known as uvalas) and large dolines may subdivide internally into smaller second generation basins. Where all the available space is occupied by depressions, rather like an egg box, the landscape is termed polygonal karst, because the

Table 12 Doline/sinkhole English language nomenclature as used by various authors

<table>
<tr><th colspan="2">Doline-forming processes</th><th>Ford and Williams 1989</th><th>White 1988</th><th>Jennings 1985</th><th>Bogli 1980</th><th>Sweeting 1972</th><th>Culshaw and Waltham 1987</th><th>Beck and Sinclair 1986</th><th>Other terms in use</th></tr>
<tr><td colspan="2">Dissolution</td><td>solution</td><td>solution</td><td>solution</td><td>solution</td><td>solution</td><td>solution</td><td>solution</td><td></td></tr>
<tr><td colspan="2">Collapse</td><td rowspan="3">collapse</td><td>collapse</td><td>collapse</td><td rowspan="2">collapse (fast) or subsidence (slow)</td><td>collapse</td><td>collapse</td><td rowspan="2">collapse</td><td></td></tr>
<tr><td colspan="2">Caprock collapse</td><td>—</td><td>subjacent collapse</td><td>solution subsidence</td><td>—</td><td>interstratal collapse</td></tr>
<tr><td>Dropout</td><td rowspan="2">subsidence</td><td>cover collapse</td><td rowspan="2">subsidence</td><td rowspan="2">alluvial</td><td rowspan="2">alluvial</td><td>subsidence</td><td>cover collapse</td><td></td></tr>
<tr><td>Suffosion</td><td>suffosion</td><td>cover subsidence</td><td>—</td><td>cover subsidence</td><td>ravelled, shakehole</td></tr>
<tr><td colspan="2">Burial</td><td>—</td><td>—</td><td>—</td><td>—</td><td>—</td><td></td><td>—</td><td>filled, palaeo-</td></tr>
</table>

Source: from Williams 2003 modified from Waltham and Fookes 2003

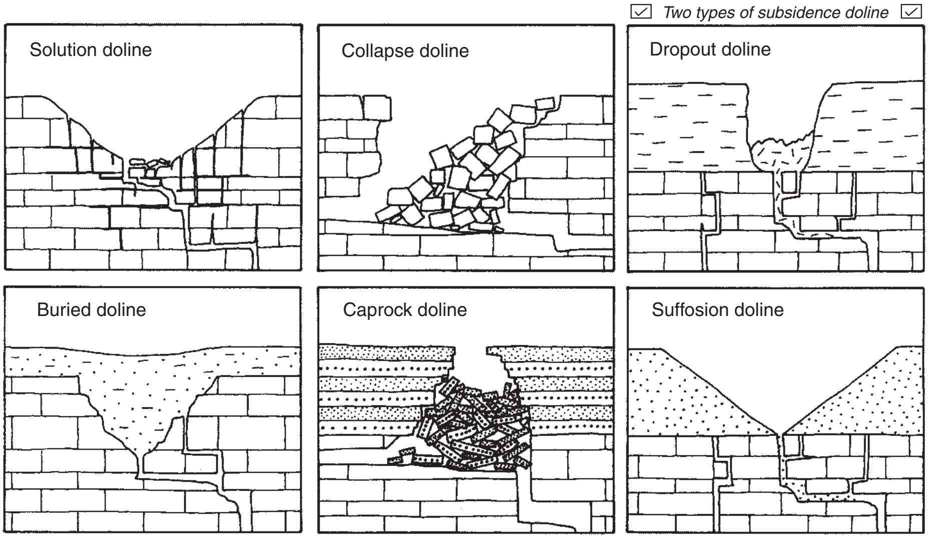

Figure 47 Classification of dolines (after Williams 2003)

topographic divides of the adjoining solution depressions have a polygonal pattern when viewed in plan.

Collapse dolines

Collapse dolines are formed mainly by mechanical processes. There is considerable variation in nomenclature concerning depressions formed mainly by mechanical processes (Table 12), largely because of the variety of materials and processes involved (Waltham 1989). Collapse refers to rapid downward movement of the ground, whereas subsidence refers to gradual movement sometimes without even ripping the surface. These processes can occur in karst bedrock, in caprock that may stratigraphically overly it, and in veneers of unconsolidated sediments. In all cases the collapse has to be preceded by dissolution of the karst rock to form a void into which material can fall. The kind of landforms produced depends upon which of the various materials and processes were involved.

Where collapse dolines form in karst bedrock then the void is commonly part of a cave system. Collapse may occur following undermining from below as the roof of a cavity stopes upwards, ultimately causing the surface above to collapse, or following dissolution from above that weakens the span of a cave roof, causing it to collapse. For example, solutional attack by drainage water near the bottom of a solution doline may combine with upwards stoping of an underlying cave roof to weaken a span from above and below, thereby causing the doline floor to collapse into a cave. Collapse dolines are on average smaller in diameter than solution dolines, although particularly large examples 700 m along their largest axis and up to 400 m deep are known in the Nakanai Mountains of New Britain, Papua New Guinea.

Sometimes a collapse extends from a cave below the modern water-table level, in which case the collapse doline will contain a lake. Such features are known as cenotes after the type-site in the Yucatan Peninsula of Mexico, although similar landforms are found elsewhere, such as in south-east Australia. The deepest known case of a collapse doline containing a lake is the Crveno Jezero (Red Lake) in Croatia, which is 528 m deep from its lowest rim, the bottom of the collapse extending 281 m below the modern level of the nearby Adriatic Sea.

Another process that increases the effective stress on rock arches and subsurface domes is removal of buoyant support by water-table lowering. This increases the effective weight on the span of the roof, resulting in its strength being exceeded and so in its failure and collapse. This occurs because in a fully saturated medium the buoyant force of water is 1 t m^{-3}, and if the water table is lowered by 30 m, the increase in the effective stress on the rocks is 30 t m^{-3}. A gradual lowering of the water table occurs with valley incision, because springs are lowered too, and with them the level of the saturated zone that feeds them. More rapid still is the lowering caused by sea-level fall, a process that occurred frequently in the Pleistocene because of repeated glacio-eustatic (vertical movement of sea level caused by glaciation and deglaciation) fluctuations. This particularly affected karsts well connected to the coast such as in Florida, southeastern Australia and Yucatan, where it probably was a significant influence in the development of cenotes.

If unconsolidated coverbeds are drained by water-table lowering, then consolidation and compression occurs, leading to subsiding of the surface and collapse where clastic sediments span de-watered unsupported arches. This is a common process in Florida where porous sandy formations overlie karstified limestones, and has been exacerbated by groundwater pumping for water supplies, which has still further reduced buoyant support. This process and the resulting incidence of collapse attains dangerous hazardous proportions in karstified areas extensively de-watered by mining activities (Beck and Pearson 1995). These dolines in unconsolidated coverbeds are sometimes referred to as cover collapse sinkholes (Table 12).

Subsidence (suffosion/dropout) dolines

When unconsolidated deposits such as alluvium, glacial moraine, loess or sand mantle karstified rock, the sediments are sometimes evacuated downwards through corrosionally enlarged pipes in the underlying karst, resulting in gradual or rapid SUBSIDENCE of the surface. Hence, the term subsidence doline is sometimes used for any closed depression in unconsolidated deposits, although the term is also used for depressions formed by much larger scale regional subsidence. Often a combination of processes is involved in the development of subsidence dolines including corrosion and collapse of the underlying bedrock, as well as suffosion, mudflow and void collapse in the mantling materials. However, the main process by which the sediment moves is known as suffosion and involves the gradual downwashing of fines by a combination of physical and chemical processes. The topographic consequence of this activity depends on whether the material is cohesive or non-cohesive. In cohesive sediments evacuation of material may proceed for some time without any surface expression. However, a void is formed that enlarges and stopes upwards resulting in a sudden, and sometimes catastrophic, failure of the ground surface. The depression thus formed is called a dropout doline or cover collapse doline. In Britain suffosion dolines formed in glacial boulder clay overlying limestone are widely referred to as shakeholes. Similar features but in more uniform finer grained materials are referred to as cover subsidence sinkholes in the USA.

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PAUL W. WILLIAMS

DONGA

Derived from the Nguni word *Udonga*, meaning a wall, it is a term used in southern Africa to describe a gully or BADLAND area caused by severe erosion (Plate 35). Widespread in Lesotho, Zimbabwe, the middleveld of Swaziland, in the Karoo, and Kwazulu-Natal, they are especially prevalent in COLLUVIUM and in deeply weathered bedrock in areas where the mean annual precipitation lies between *c.*600 and 800 mm. Where the materials in which they are developed have high ESP (Exchangeable Sodium Percentage) contents, they may have highly fluted ‘organ pipe’ sides (Watson *et al.* 1984). Repeated oscillations have taken place in colluvium deposition and palaeosol formation on the one hand, and incision on the other (Botha and Federoff 1995). Causes of incision may include climatic change, and land cover changes brought about by human activities, the latter including the spread of pastoralism and deforestation for iron smelting. Debates about their origin are similar to those that have been raised in connection with the formation of ARROYOs in the American west. Piping

Plate 35 A deep donga developed in highly erodible colluvial material in a valley bottom near St Michael's Mission in central Zimbabwe

(see PIPE AND PIPING) is probably an important process in their development (Rienks *et al.* 2000).

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A.S. GOUDIE

DOWNSTREAM FINING

The characteristic decline in the average size of riverbed material with distance downstream is downstream fining. This may include a full sequence from boulder-sized material close to the river source, through gravel-, sand- and silt-sizes to clay-sized material where the river enters the sea. Many rivers do not have all these changes, and may have gravel- or sand-beds at termination. The downstream decline in grain size was first recognized to follow a negative exponential trend by Sternberg (1875) who explained this by the process of ABRASION. Abrasion is significant in many cases, but laboratory measurement of abrasion rates suggests that abrasion alone is insufficient to account for observed rates of fining. It has long been recognized that smaller sediment particles should move more frequently and further than larger ones during BEDLOAD transport. This selective transport mechanism was questioned during the 1980s when it was found that, in GRAVEL-BED RIVERs, there is only slight size selectivity in bedload transport. Further investigation has shown that even a small degree of size selectivity can cause significant downstream fining over long time periods. Downstream fining is thus best explained as a consequence of size sorting during bedload transport, with abrasion and particle breakdown generally acting as secondary effects that may accelerate the rate of fining.

Downstream fining is also one of the downstream adjustments that takes place in graded river systems (see GRADE, CONCEPT OF), along with changes in bed slope, channel width and depth, and flow velocity. The rate of downstream fining (the degree of concavity of a graph of particle size versus distance downstream) is inversely proportional to the length of the river, such that fining is rapid in short rivers and slow in long ones. Close inspection of bed material size data shows that downstream fining is rarely a smooth, continuous process. Abrupt changes in grain size occur where tributaries enter the river, or close to sediment sources (see TERRACE, RIVER). These perturbations are smoothed out at the scale of the whole river, but demonstrate how river networks route both water and sediment downstream, and cause changes in stream ecology. Particularly notable is the abrupt transition from gravel- (>2 mm) to sand-sized (<2 mm) bed material that occurs in many rivers. This transition can result from the supply of large amounts of sand-sized material to the river, or from complex interactions between sediment movement and flow hydraulics that occur as the percentage of sand in the river bed exceeds about 20 per cent.

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SEE ALSO: channels, alluvial; hydraulic geometry

TREVOR B. HOEY

DRAA (MEGADUNE)

Draa, the Arabic word for ‘arm’, may be used to denote the largest members of the aeolian bedform hierarchy. The term was first used in English by Wilson (1972).

Draas are also known as compound and complex dunes (Breed and Grow 1979), or megadunes (Warren and Allison 1998). They are typically large bedforms with a spacing exceeding 500 m and a height reaching 200 or 300 m and may occur as linear, crescentic, or star forms. Examples of linear draa occur in the Namib Sand Sea, Rub al Khali of Arabia and the Akchar erg of Mauritania; crescentic draa can be found in the Liwa area of the United Arab Emirates and Saudia Arabia, the Namib Sand Sea, and the Algodones dunefield of California. Draa of star form occur in the Grand Erg Occidental and Oriental of northern Africa, the Namib Sand Sea, and the Gran Desierto of Mexico.

Draas are characterized by superimposed bedforms of dune size, with heights up to 10 m and a spacing of up to 300 m. In some places, e.g. the northern Namib Sand Sea, the superimposed dunes appear to be features contemporary with the main bedform (Bristow *et al.* 2000); elsewhere, e.g. in Wahiba Sands of Oman and in Mauritania, the superimposed dunes represent different generations of dunes, in some cases formed in a wind regime different from that which formed the main draa (Warren and Allison 1998; Lancaster *et al.* in press). Thus crescentic dunes may be superimposed on linear draa, and two or more smaller sets of linear dunes are superimposed on older linear draa.

The large size of draa has been thought to be the product of strong winds (e.g. Wilson 1972), but others have suggested that their large size is a product of long continued development in a wind regime that promotes deposition on the dune (e.g. Lancaster 1988). Their large size indicates persistence over long periods of time and reconstitution times in the order of 1 to 100 ka.

Recent stratigraphic and dating studies suggest that some draa (especially linear draa, which tend to conserve their form over long periods) may be composite landforms constructed by multiple generations of aeolian deposition, stability and reworking. In several areas (e.g. UAE, Oman, Mauritania), the cores of large linear draa are at least 15–22 ka old (Glennie and Singhvi 2002; Lancaster *et al.* in press).

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SEE ALSO: dune, aeolian

NICK LANCASTER

DRAINAGE BASIN

A drainage basin is an area of land that contributes water and sediment to a specific outlet point on a stream. It is separated from other drainage basins by its drainage divide, a boundary that encircles a basin along its highest, outermost ridge tops. The drainage basin is recognized as a fundamental geomorphological unit (Horton 1932: 350) and is frequently used as the primary landscape unit for hydrological, water supply and ecological investigations and for land management activities.

Drainage basins are EROSION created landforms sculpted predominantly by the actions of flowing water. They may be conceptualized as consisting of two geomorphological components: a set of hillslopes dominated by unconfined OVERLAND

FLOW and a branching network of stream channels conveying concentrated flows. The transition from hillslope to channel has been characterized as both indistinct and distinct. Davis (1899: 495) wrote: 'Although the river and hillside waste-sheet do not resemble each other at first sight, they are only the extreme members of a continuous series; and when this generalization is appreciated, one may fairly extend the "river" all over its basin and up to its very divides.' From this perspective, every point within a drainage basin is located along a flow pathway, and a basin is composed of a branching, space-filling drainage network of flow pathways extending from outlet to basin divide. The alternative viewpoint is that the transition from hillslope to channel is determined by a geomorphological threshold (see THRESHOLD, GEOMORPHIC) of channelization that sets a finite scale for dividing a landscape into valleys and hillslopes. Because a drainage basin may be defined upstream of any point on the land's surface, the delineation of a landscape into specific drainage basins is done for some designated purpose.

Several other terms are used synonymously with drainage basin. In Great Britain, catchment is commonly used, whereas in the United States, watershed is a preferred term. Unfortunately, watershed is an ambiguous term that has historically been used as a synonym for drainage divide, and this usage is retained in Great Britain. For large basins drained by a major river, the term river basin is often used (e.g. Amazon River basin). Drainage basin, catchment and watershed do not inherently imply a particular size of drainage area. However, some government agencies in the United States and others are using these terms in size-based classification systems, such as catchment being smaller than a watershed and watershed being smaller than a basin.

The form and structure characteristics of drainage basins and their associated drainage networks are described by their MORPHOMETRIC PROPERTIES, which can be classified into the categories of size, surface, shape, relief and texture. Drainage area, a variable specifying the amount of land area contained within a drainage divide, is an important basin descriptor and is frequently used as a surrogate for the amount of water and sediment yielded by a drainage basin. Because a basin's drainage network is its most prominent feature, network morphometric properties are also used for drainage basin description. A qualitative indication of drainage basin size is indicated by the stream order of its outlet stream (see STREAM ORDERING). The delineation of drainage basins and determination of their morphometric properties has traditionally been done using topographic maps and manual methods. With DIGITAL ELEVATION MODELS (DEM) and geographic information systems (GIS), watershed delineation, drainage network extraction and the automatic calculation of morphometric properties is possible.

Although drainage basins are fundamental geomorphological units, they may not always prove the best choice for organizing landscapes for research or land management purposes. In landscapes dominated by non-fluvial features such as kettle holes (see KETTLE AND KETTLE HOLE) or aeolian dunes (see DUNE, AEOLIAN) drainage networks and drainage basins are often poorly defined and may be difficult to delineate using either manual or GIS methods.

Drainage basin organization

A drainage basin may be organized into two subsidiary landform units: a set of hillslopes and a drainage network. Although hillslopes may occupy 95 per cent or greater of a basin's area, it is the drainage network that noticeably provides the organization of the hillslopes within a basin. The drainage network is the tree-like structure of flow pathways along which water and sediment are concentrated and delivered to the basin outlet. Drainage networks are comprised of exterior and interior links between successive nodes, where nodes are sources, junctions or the basin outlet. Exterior, or first-order links, connect an upstream source node to a downstream junction. Interior links connect two junctions or a junction to the basin outlet.

Using GIS and DEMs, a space-filling network of flow paths can be delineated within a drainage basin, with external links terminating at the basin's exterior divide or internal divides. There are several subsidiary networks contained within the space-filling drainage network. Many investigators discuss the drainage network in terms of streams and stream system, referring to the blue-line streams on topographical maps. Some have considered the stream network to be synonymous with the channel network, but others view channels as geomorphological features identifiable only from field investigation. A drainage basin will also contain a network of VALLEYs, which may or may not contain streams or channels.

Much of the geomorphological analysis devoted to drainage basins has been with respect to the organization and development of the branching link drainage network structure. In a seminal geomorphology paper, Horton (1945) provided many of the concepts supporting modern geomorphological analysis of drainage basins. He provided the basis for the hierarchical method of stream ordering and laws of drainage composition that with later modifications due to Strahler (1957) and others provide a means to organize the understanding about the topologic and geometric properties of drainage networks. In the Horton/Strahler ordering system, source streams (exterior links) are designated as first order. When two first-order streams join, the stream that continues is designated as second order and, in general, at the junction of two streams of equal order, the order of the downstream segment is increased by one. Low-order tributaries may flow into high-order streams without the order being incremented, and the entire section of stream of same order is referred to as one stream segment for the purposes of quantifying the number of streams, stream length, stream slope and contributing area. HORTON'S LAWS of drainage composition refer to the empirical straight-line relationships between these quantities and stream order on semi-log plots.

Horton (1945: 283) also devised the concept of DRAINAGE DENSITY, which indicates the degree of dissection of a drainage basin into subsidiary hillslopes by its channel network. Horton's concepts of network analysis can be applied to any of the drainage networks including channel, valley and GIS-derived networks.

Probabilistic-topologic approaches to network analysis have been devised that examine both the regularity and randomness of drainage networks (Shreve 1966; Smart 1968). The random topology models can readily explain many of Horton's laws. Horton's laws are not actually laws in the strictest sense, but merely expressions of the most probable states of network composition.

Horton's laws characterize the self-similarity in the organization and structure of river networks. This self-similarity has stimulated the use of fractals to characterize river networks (see FRACTAL). Fractals are objects with self-similar geometry, retaining similar organization and complexity over a range of scales. The planform river network when characterized as a fractal has a fractal dimension between one (linear features) and two (filling a two-dimensional space), that can be related to Horton's bifurcation and length ratios (Tarboton *et al.* 1988; La Barbera and Rosso 1989). Hack (1957) first noted an apparent dimensional inconsistency between the lengths of the mainstream and drainage area of a river basin. This can imply elongation with increasing basin size, an idea inconsistent with self-similarity but that has been advanced by some, or that individual streams are themselves fractal with dimension between 1.1 and 1.2. There have been suggestions that space filling is a constraint on the organization of river networks, because in general they should drain an entire two-dimensional area. This leads to a constraint that implies relationships between Horton's length, bifurcation and area ratios and the fractal dimension of individual streams.

The uniting of drainage network configuration and flow characteristics has been proposed in the theoretical framework of the optimal channel network. An optimal channel network is one in which there is energy minimization in the whole and parts of a drainage network. Three principles of optimal channel networks are that energy expenditure in every link is minimized for the transportation of a given discharge, equal energy is expended per unit area everywhere within the network, and energy expenditure is minimized for the network as a whole. A combination of these principles is sufficient to explain the tree-like structure of drainage networks and the empirical relationships describing network organization.

FIRST-ORDER STREAMS and drainage basins are substantial components of river basins. At the upper end of a first-order channel is an unchannelled HILLSLOPE HOLLOW or zero-order drainage basin. Nearly one-half the length of a river basin's drainage network may be contained in its first-order links, and first-order basins can contain 50 per cent of a river basin's area. It is within such small watersheds that runoff produced on hillslopes concentrates into streamflows that initiate channel formation. Low-order basins exhibit the tightest HILLSLOPE-CHANNEL COUPLING and competition between hillslope processes (see HILLSLOPE, PROCESS) and channel processes.

Basin development and evolution

The expression of drainage basin organization is a spatial characterization of drainage basin condition at a point in time. Drainage basins and networks, however, are not static and change over time due to external influences and internal COMPLEX RESPONSES. To explore drainage basin development, geomorphologists have used three different methods: space-for-time substitution, experimental studies and computer simulation modelling. Early studies of drainage basin evolution were based upon space–time substitution, i.e. the ERGODIC HYPOTHESIS. Maps of different drainage basins in progressive stages of development were ordered to depict a temporally evolving basin undergoing advancing stages of evolution. EXPERIMENTAL GEOMORPHOLOGY has been employed through the use of rainfall simulators raining over 'sandboxes' with drainage system development documented through detailed mapping and time-lapse photography. With advances in computer technology, empirical and theoretical concepts have been implemented into computer models that simulate long-term drainage basin evolution.

Although modes of basin evolution depend upon specific boundary conditions and driving factors, several major steps in basin development can be specified. With the assumption that a basin originates on a smooth surface, its channel network grows through the processes of initiation, elongation and elaboration. After initial development of a skeletal network, a few streams elongate to extend in parallel fashion across much of the length of the surface to form a low drainage density network. Over time, downcutting of these elongated stream channels causes the elaboration of the network through the addition of tributary streams, with the concomitant increase in drainage density. During these initial phases of basin development, drainage density increases rapidly and SEDIMENT LOAD AND YIELD from the basin are high.

Eventually, the channel network reaches a period of maximum extension in which stream elongation through HEADWARD EROSION and infilling by hillslope processes reach an equilibrium condition. Smaller drainage basins become integrated into larger ones through the mechanism of RIVER CAPTURE. As erosion continues and the entire basin drops closer in elevation to BASE LEVEL, the process of network abstraction occurs. Lowered stream slopes reduces STREAM POWER and stream EROSIVITY thereby allowing hillslope processes (see SOIL CREEP and SLOPEWASH) to infill low-order streams and abstract them from the drainage network.

Except in experimental models, rarely will a drainage basin originate on a flat, sloping surface of uniform material, so evolution of real drainage basins will be much more complex than the above model depicts. Also, there is no timescale associated with the evolutionary model described above, but longevity of a drainage basin and its drainage system can be correlated with increasing basin size. Large river basins may persist for tens of millions of years. During such long time frames, changing climatic conditions and tectonic events can so alter conditions that completed evolutionary stages are never fully realized. Channel networks can expand and contract through upstream and downstream migration of channel heads as climate changes over periods of decades to thousands of years. Tectonic events can drop or raise the base level for a drainage basin and initiate a new cycle of headward erosion or halt an existing erosional stage (see TECTONIC GEOMORPHOLOGY).

Also, the channel network evolution model presented above does not account for the causal processes of network development and evolution. Different processes are responsible for channel initiation and evolution in different terrain and environments. In steep terrain, LANDSLIDES may result in the initiation of channels, but in low-gradient basins, headward erosion of the channel head because of changed CONTRIBUTING AREA hydrological conditions may be the primary method of network growth. Though simple descriptive models can specify the overall pattern of evolution, detailed circumstances of basin evolution will be quite variable from one basin to another and may require complex computer simulation models to fully understand.

Geologic and climatic influences

Seeking to explain the regularity exhibited by drainage networks has guided much of the geomorphological interest in these features. Although such explanations provide theoretical bases for network pattern regularity when boundary conditions are uniform in space and invariant over time, actual drainage networks evolve under spatially and temporally varying conditions. Therefore, variability in DRAINAGE PATTERNs can arise because processes defined by geomorphological laws (see LAWS, GEOMORPHOLOGICAL) are operating within non-uniform environmental conditions. In particular, geology and climate have profound effects upon the processes and characteristics of drainage basins and drainage networks. Some have suggested that it would be more beneficial to seek relationships between geology and drainage network characteristics than refine sophisticated theories that disregard such an obvious control (Blöschl and Sivapalan 1995: 282).

The effect of geology upon drainage basin characteristics is difficult to quantify, but geology is nonetheless a controlling factor on drainage basin form and development at multiple scales. Empirical studies have identified relationships between drainage density and bedrock lithology, and rock type can be responsible for many drainage pattern details. Drainage basins underlain by shale or siltstone tend to have higher drainage densities than other lithologies due to low infiltration rates and high production of overland flow. Areas dominated by lithologies with high infiltration rates, such as dune sands, often have poorly defined drainage systems and very low drainage densities. Dendritic stream patterns are common on shales and siltstones, as they are weak rocks that provide limited lithologic resistance to erosion. In geologic formations comprised of stronger rock types, such as sandstones and granites, joint patterns frequently control network pattern because fractured rock along joints has greater ERODIBILITY.

Geologic structures also influence drainage basin shape and drainage network form. At a large scale, river basins may be coincident with geologic or structural basins, with the basin's drainage divide corresponding to the ridgetops of surrounding, uplifted mountain chains. For smaller drainage basins, folded geologic strata can control drainage basin shape and drainage pattern where weaker strata are more readily eroded. Trellis and annular drainage patterns are common in these circumstances. Multiple, low-order streams flow from divides to strike valleys, medium-order streams occupy longitudinal or strike valleys, and master high-order streams run across the strike of more resistant folds in superimposed traversal valleys. Drainage basins may be irregularly shaped with drainage divides following the ridgelines of HOGBACKs or CUESTAs formed by more resistant rock strata. Faults, similar to joints, are areas of weaker, fractured rock and are frequently occupied by streams in superimposed valleys.

Climatic effects upon drainage basin development are promulgated through their controls upon erosion processes. The most critical effect of climate upon drainage basin form and development is through the influence of precipitation and temperature upon vegetation cover. Density of vegetation cover is a predominant control upon erosion and sediment delivery to the drainage network by decreasing soil erodibility. Channel head location, and thereby drainage density, may be dependent upon vegetation because of increased shear stress required for channel formation where vegetation cover is dense. Drainage densities typically are low in arid regions where there is little runoff, are highest in semi-arid regions where sparse vegetation cover does little to prevent channel initiation, are low in moderate precipitation environments where vegetation cover restricts channel development, and can be high even with heavy vegetation cover in areas with high annual rainfall and runoff.

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SEE ALSO: GIS

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DRAINAGE DENSITY

Drainage density is defined as the cumulative length of all stream channels in a drainage basin, divided by the drainage basin area. The dimension of drainage density is the inverse of length. In principle, drainage density is a fundamental measure of landscape dissection and half its reciprocal is an average measure of length of overland flow (or distance from divide to the nearest channel). Drainage density has been shown to vary as a function of climate, past climate conditions, biomass, parent material, lithology, relief, time and land use. Unfortunately, no consistent relations have been demonstrated from region to region and Schumm (1997) comes to some discouraging conclusions from a number of detailed basin studies. Nevertheless, drainage density is an important geometric parameter for channel networks as it determines the spacing of channels, the length of hill slopes, the maximum length scale of slope failures and reflects the processes governing landscape dissection. The hydrological response of a channel network is strongly influenced by drainage density and sediment erosion rates have been linked to channel spacing. There are relatively few studies of how drainage density varies with time. Flume table experiments (Schumm *et al*. 1987), studies of land fills (Schumm 1956), glacial tills (Ruhe 1952), coastal terraces (Kashiwaya 1987) and drainage networks on an anticlinal fold which has been progressively uplifted during the past 250,000 years (Talling and Sowter 1999) are representative examples of such studies.

It should be pointed out that the definition of drainage density begs at least two questions:

1 *How is a stream channel defined?* This is a deceptively simple question which does not have a simple answer. Montgomery and Dietrich (1992) suggest that there is an empirically defined topographic threshold for channel head locations which defines the boundary between essentially smooth and undissected slopes and the valley bottoms to which they drain. They derived an experimental relation of drainage area versus local slope for channel heads, unchannelled valleys and low-order channel networks from different study areas. Local slope was measured in the field and drainage area was determined from topographic maps. High slopes generate channels from smaller basin areas and lower slopes require larger basin areas to produce a channel. At the same time, spatial heterogeneity, reflecting the controlling factors listed above, introduces variability into these relations. An empirical, field-based definition of a channel uses the presence of fluvial incision and one or two stream banks, but finger tip tributaries are often indeterminate in the field.

2 *What is the relation between stream channels drawn on maps or stream channels detectable on air photographs and actual stream channels on the ground?* There is a basic stream channel network, which is composed of perennial streams and which expands and contracts with runoff, or there is the active channel network, which is composed of ephemeral, intermittent and perennial streams. In addition, the use of contour crenulations as evidence of the presence of channels will result in the inclusion of parts of valleys that do not contain active channels. Clearly, the largest problems concern the uppermost finger tip tributaries of a drainage basin. When air photos are used, there are further problems of visibility below tree canopies and as always, the scale of the photograph or the map will be a constraint on the resolution achievable. In sum, what is measured by one investigator may not be the same phenomenon as that which is measured by another.

If we assume that the identification and measurement problem can be resolved, a variety of theoretical issues relating to drainage basin characterization and evolution can be broached. Strahler (1956) in developing his view of the drainage basin as an open system tending to achieve a steady state of operation asked how to predict erosional or aggradational response by drainage basins when land use or climate changed. Central to his theoretical discussion was the role of drainage density. He argued that because drainage density is the most valuable scale index with respect to degree of dissection of a basin that it should be possible to express drainage density as a function of several variables that control the evolution of the basin. These variables he deduced in part from Horton (1945) as runoff intensity, an erosion proportionality factor, slope gradient, relief, kinematic viscosity of runoff and acceleration of gravity. Through application of the Buckingham Pi Theorem, he reduced the equation to a function containing four dimensionless groups:

1 the product of drainage density and relief (the ruggedness number)
2 the product of runoff intensity, erosion proportionality factor and slope gradient (the Horton number)
3 the product of runoff intensity, kinematic viscosity and relief (a basin Reynolds number)
4 the square of runoff intensity divided by relief times acceleration of gravity (a basin Froude number).

By solving for drainage density, drainage density is shown to be inversely proportional to relief times a function of the Horton, Reynolds and Froude numbers. The challenge of solving this function has still not been met, though the topic of drainage basin transformation has been put onto a more rational basis.

Melton followed up Strahler's analysis with one paper on drainage basin growth models (1958a) and another on the theory of variable systems (1958b), both of which relied heavily on the assumption of the importance of drainage density. Melton (1958a) demonstrated a close relation between F (stream frequency = number of streams per unit area) and D (drainage density) for mature basins with a wide range of orders, valley side slope angles, climates and rock types. The relation, subsequently known as Melton's Law, is of the form $F = 0.694\ D^2$. Shreve (1967) revisited this relation using links instead of streams and found a related term $K = 0.667$ for topologically random networks. The dimensionless ratio F/D^2 varies inversely with valley side slope and basin relief (where area and channel length are held constant) and is interpreted as a measure of the completeness with which a channel system fills a basin outline. For an ideal basin of $1\,\mathrm{mi}^2$, Melton's Law is postulated to be a growth model. This argument is predicated on the assumption that many basins measured at one point in time can be considered equivalent to the behaviour of a single basin over time. The approach taken in Melton (1958b) is different. He arranges fifteen variables of geomorphic, surficial and climatic elements into two related variable systems on the basis of correlation coefficients for every possible pair in the study of 59 drainage basins. 'Melton's ambitious field program of data collection, coupled with his analysis of the interrelations of the components of a drainage basin and the variables that influence morphology, is a model for future geomorphic studies' (Schumm 1977: 180). The high correlation of drainage density with per cent bare area and a precipitation effectiveness index fits well with the Horton theory of drainage density as a function of the resistivity of the surface to erosive forces, determined in part by vegetation which in turn determines the mean length of overland flow. Keylock (personal communication) has shown that the most frequently cited of Melton's contributions are Melton's Law (1958a) and his correlation structure approach to geomorphology (Melton 1958b), both of which emphasize the importance of drainage density.

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SEE ALSO: dynamic geomorphology; stream ordering

OLAV SLAYMAKER

DRAINAGE PATTERN

Because river channels concentrate surface flow and erode into landscape more efficiently than other processes, new channels tend to persist from the pattern initially developed and are subsequently hard to alter. A collection of river channels joined together is called a drainage network, how it is laid out on the ground in plan view is called the drainage pattern, and the channels together with all the land surface that drains to the channel is called the DRAINAGE BASIN. Channels when they join normally do so in an accordant manner, the channels join without a sudden break in elevation (sometimes called Playfair's Law), unless they occupy unmodified glacial terrain, in which case a discordant junction is called a HANGING VALLEY. Subsequent adjustments to networks and patterns may occur when rivers are close together, and the divides between them may be broached by erosion or overflow, or underground drainage may divert water from one system to another prior to there being a surface connection of the rivers. Exploitation of geological weakness by surface erosion eventually causes the overall drainage pattern to reflect the patterns of weakness in the underlying rocks. Major joints and fracture zones may influence subsurface as well as surface drainage and tend to localize major channels. Adjustments by divide erosion and breaching (river piracy, RIVER CAPTURE, diversion) will be most common early in the history of a landscape when relative RELIEF is least. Adjustments by underground diversion may take longer to become active features because large subterranean networks, usually developed in soluble bedrock such as carbonates (KARST terrains), are needed to divert substantial drainage (abstraction). Subterranean diversion is favoured by increasing local relief in the drainage which may permit steeper hydraulic gradients between adjacent channels. Drainage patterns which derive their water entirely from regions external to the locality in question – such as the Nile River in Egypt – are called exoreic, and systems which drain to a central closed depression such as the Jordan River to the Dead Sea, and the basin draining to the Great Salt Lake in Utah – are endoreic.

The nineteenth and early twentieth-century geomorphologist W.M. Davis (1889, 1899) developed an elaborate scheme to describe the components of a river drainage network as they related to stages in its physiographic development. Of those terms, those which remain in common use are *consequent* and *subsequent*. *Consequent* streams are those that develop on the initial land surface in response to regional slope and any random surface declivities. Because they must eventually follow regional slope they usually reflect the tectonic framework of uplift, rather than details of the underlying geology. The term has normally been applied to large streams, but can also describe initial drainage on any new surface – such as recently glaciated terrain. *Subsequent* streams describe streams which, through geologic time, have been able to exploit differences in the relative erodibility of the underlying geology as the drainage system incises slowly into the uplifted block of land. Typically they develop along the geological strike exploiting, for example, weak shales or clays exposed between stronger formations (e.g. sandstones or limestones) in a sequence of sedimentary rocks so that long continued weathering and subaerial erosion over CYCLIC TIME etches out a skeleton of the underlying geology – thus the ridges and valleys of the Appalachian Ranges along the eastern side of North America reveal the folded structures; less dramatically the valleys and escarpments of southern England and northern France also reveal the geological structures. The effect is even more dramatic in dry climates with no masking vegetation. In igneous and metamorphic terrain master joints and shear zones may provide weakness to exploit (Figure 48c). Faults and fault zones, with heavily fractured rocks allowing access for weathering agents, are often weak zones in any geological terrain.

Because consequent drainage flows down the regional slope regardless of local variations in geology, such streams are often used to reconstruct the initial stages of a landscape. However,

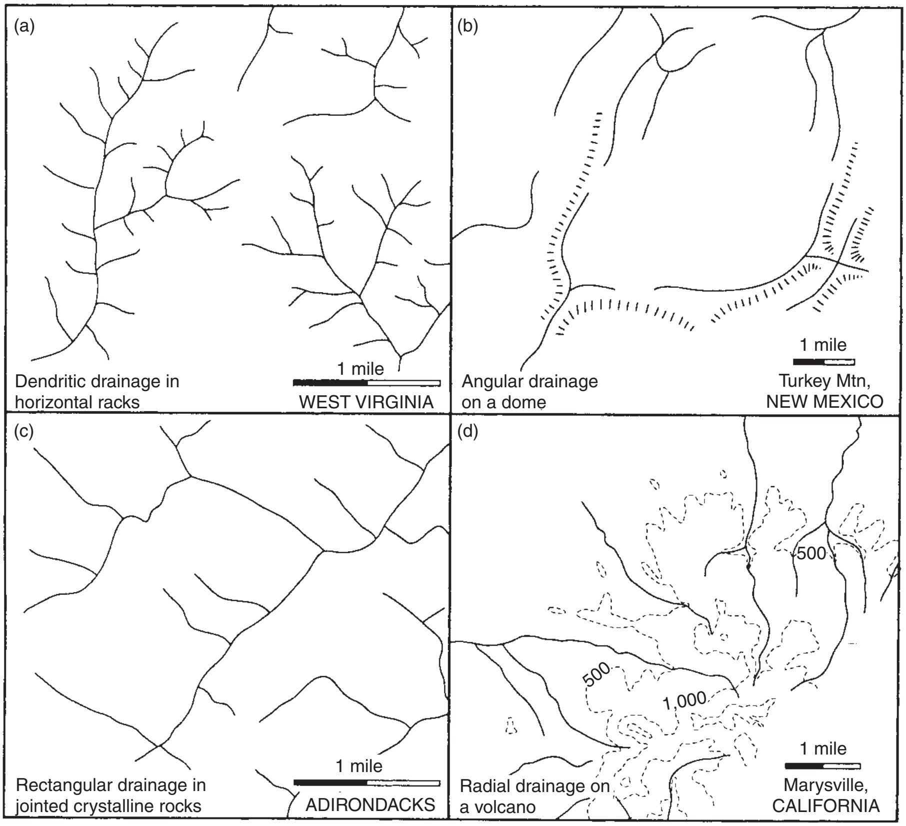

Figure 48 Drainage patterns in relation to topography and geological structures

even consequent streams can be disrupted by continued uplift, with geological structures growing upwards into the overlying streams. If the river channel can erode its bed fast enough to maintain a continuous downslope against the rising land, the river is called antecedent. As a result a river may be seen to have cut a channel, often seen as a deep gorge (see GORGE AND RAVINE), through a prominent topographic ridge around which it might have otherwise been forced to flow. The north to south segments of the Ganges and the Brahmaputra in the Himalaya have been cut in response to, and across the rising folds of, the mountain system.

On occasion though, the rising structure blocks the channel and causes upstream ponding, whose new outlet may provide an entirely new pattern. Complete reversal is possible too. The Amazon originally drained to the Pacific, but its course was reversed by the rising Andean ranges. A related condition, however, is when a regional river system, developed for example on gently tilted sedimentary strata, slowly erodes away that sediment and then erodes into a very different geological underlay. If the sedimentary cover rocks are lying unconformably upon the rocks below, the drainage pattern is said to be superimposed or superposed (Tarr 1890). It is doubtful in practice that either antecedence or superimposition are ever pure conditions because rarely can the full tectonic history of the region be known (Smith *et al.* 1999).

Also, large-scale topographic patterns characterizing the initial topography of the area may be reflected, as for example with radial drainage, such as in the English Lake District where original drainage lines have been greatly accentuated by glacial deepening. Part of a miniature example of radial drainage developed on a volcano is shown in Figure 48d.

Davis developed many terms for other parts of the drainage system as they related to a supposed sequence of drainage and landscape development, and with respect to the original regional slope. These other terms are: insequent, resquent, obsequent; but they have fallen into disuse. Full definitions are available in Lobeck (1939: 171). Of these, insequent streams describe the myriad of streams for which no discernible control can be detected, and which give rise to dendritic patterns (Figure 48a).

Despite the variations in apparent patterns (Figure 48) the patterns that matter most to the operation of the system are the internal structure of connections, and the plan of the DRAINAGE BASIN on the ground. Circular basins concentrate flow more rapidly, and generate larger peak flows than elongate basins. The structural arrangement of channels tends to reflect that of the ground plan – dendritic or vein-like structures being found usually in oval and round basins with homogeneous bedrock (Figure 48a). The Kentucky region with nearly level sedimentary rocks, and lying beyond the glacial limit, has often been used as a basis for comparison with random or randomly generated drainage networks (Mark 1983).

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DRAPE, SILT AND MUD

A thin deposit of waterlain silt or mud coating a pre-existing morphological feature. Drapes generally grade upwards, sometimes exhibiting internal laminations. They are typically several centimetres thick, though their form and composition vary spatially and temporally, in response to factors including sediment supply, and the hydrological (fluvial or current) regime.

Drapes are indicative of tidal/subtidal settings and are considered one of the most distinctive features of such an environment. Typical tidal regimes exhibit ebb-flood cycles in which one current is more dominant than the other. During periods of tidal dominance various bedforms are produced (e.g. sand bars, ripples and dunes) that are characteristic of the tidal regime. However, there is a period of time during high tide and low tide where no dominant direction of flow exists (termed the slackwater period). During this short period, the suspended sediment of the water may fall and settle on the pre-existing features formed during the dominant tidal period. The subsequent current stage may partly rework the mud or clay drape producing an erosive reactivation surface, though the cohesive nature of the fine clay-rich drapes commonly protects against tidal erosion and preserves the drape. Over time, continued preservation of alternating tidal (sand deposited) and slackwater deposits (mud/clay drapes) produces sand/mud couplets, also called tidal rhythmites. This systematic deposition of tidal rhythmites has allowed detailed reconstruction of past tidal regimes (e.g. Visser 1980), and are particularly distinctive of inshore tidal environments.

Drapes may also form in fluvial environments, particularly within rivers that exhibit seasonal flow and flooding. As flooding wanes, the clay/mud settles on river levees, etc. thus signifying slackwater periods and forming drapes.

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STEVE WARD

DRUMLIN

Drumlins are roughly ovoid-shaped hills dominantly composed of glacial debris. Typically, they occur within groups or fields of several thousands, exhibit strong, *en echelon*, long-axis preferred orientation paralleling the main direction of ice flow. The classical shaped drumlin usually has a steeper stoss end and a tapered lee-side, however variants on this shape are perhaps more common than the classical shape itself. Drumlins may vary from 5 to 200 m in height, 10 to 100 m in width, and overall from 100 m to several kilometres in length (see e.g. Mills 1987). Interestingly, few modern drumlins appear from beneath modern-day ice masses other than on James Ross Island, Antarctica, in the proglacial areas of some Icelandic outlet glaciers, and at the Bifertensgletscher, Switzerland. Vast drumlin fields – numbering in thousands – exist, for example, in Canada, Estonia, Finland, Ireland, Germany, Poland, Russia and the USA. The topographic locations within which drumlins are found are many and varied. Drumlins occur in both lowland and highland terrains, beneath ice sheets and valley glaciers, close to terminal MORAINEs and may, in places, appear contiguous with these moraines, while elsewhere they occur on the edge of ice sheet centres. Occasionally, a radiating pattern can be observed within a drumlin field that has been interpreted as evidence of basal crevasse infilling owing to divergent ice flow close to an ice margin. It has also been suggested that drumlins, in association with Rogen and fluted moraines, may be related to deformable beds beneath ice sheets, linked to fast basal ice ($>500\,\mathrm{m\,a^{-1}}$mm) and a preferential location within ice streams. Limited relationships appear to occur between drumlins and topography.

Drumlins are composed of a vast range of sediment types of varied provenance, containing an array of sediment structures and forms (Figure 49). In the past, drumlins were mistakenly perceived as being composed almost exclusively of subglacial tills; although many drumlins contain stratified sediment. Drumlins composed of stratified sediments are known, for example, near

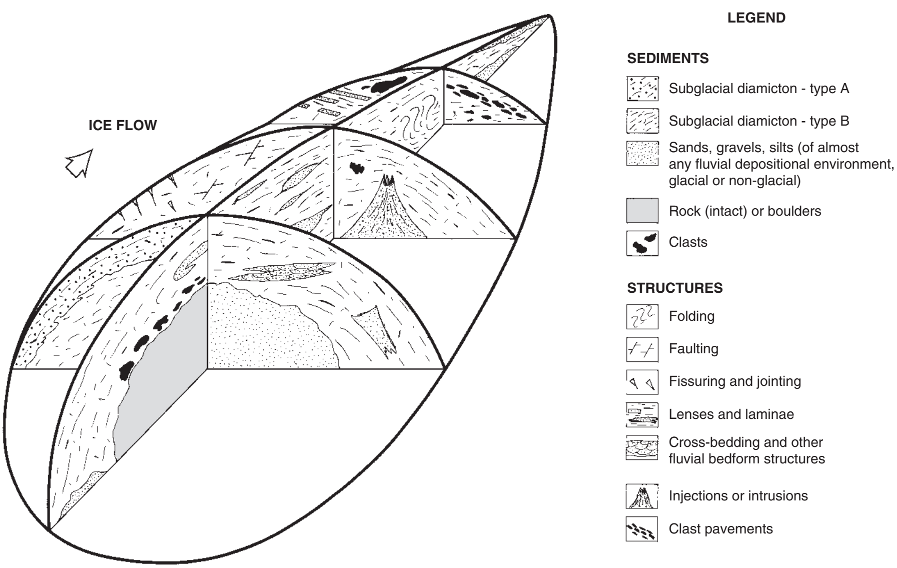

Figure 49 A general model of internal sediments and structures found within drumlins

Velva, North Dakota and Livingstone Lake, Saskatchewan. Also, stratified sediments can occur in individual drumlins that often 'sit' adjacent to till drumlins as in Peterborough, Ontario. Many drumlins have observable cores of bedrock, boulder dykes and other non-glacial nuclei around which subglacial debris has accreted or been emplaced. In some cases, drumlin or drumlinoidal forms can be observed 'carved' from bedrock in the form of roc-drumlins. However, most drumlins do not appear to have obvious cores around which they have been 'built' and these forms remain enigmatic in origin.

Clast fabrics within drumlins appear, in some cases, to follow the outer morphology of the drumlin, while others exhibit a transverse, 'herringbone' style pattern. In many cases the complexity of internal sedimentological structures provides a random fabric orientation. Drumlins exhibit such a wide complexity of form and internal composition that it is impossible to characterize an 'ideal' drumlin. Many drumlins are found lying on top or obliquely across other larger drumlin forms (megadrumlins). Drumlin shapes may vary enormously and may reflect formative processes or simply post-depositional subaerial mass movement. Many drumlin fields progressively change as part of a continuum of bedforms, thus drumlin genesis would appear tied to subglacial environments conducive to Rogen and fluted moraine formation. The question of drumlin formation has attracted an array of research. In terms of drumlin formation, it is germane to consider the 'conditions' that must be met by any formative hypotheses, assuming that a single explanation does exist for such a diverse landform/bedform type.

Any explanation of drumlin formation must address the following issues: (1) the diverse location of drumlins and their propensity to occur in 'fields'; (2) the differing shape and morphology of drumlins; (3) the range of sediment types and structures within drumlins; (4) the existence of rock-cored and non-rock-cored drumlins, often in proximity to each other; (5) the presence of drumlins in bedform continua in some, but not all, cases; (6) the relationship of drumlins to subglacial glacidynamics and hydraulics; (7) the chronology of drumlin formation whether drumlins form simultaneously as a single field or develop into a field by repeated 'overprinting' in a single glacial phase or repetition over several glacial phases; (8) stages of drumlin development whether formed *en masse* or by gradual accretion in a single continuous event or interrupted accretionary events; and, finally, (9) a 'trigger' mechanism(s) that is operative in certain specific conditions yet not under others.

At present, three main groups of drumlin-forming hypotheses can be identified:

1. By moulding of previously deposited material within a subglacial environment in which a limited amount of subglacial meltwater activity occurs (possibly where a frozen bed transforms to a melted bed; Menzies 2002). Meltwater may influence moulding and deformational processes by acting either as a lubricating basal film at the upper ice–bed interface, or as porewater reducing subglacial sediment effective stresses. Debris is moulded by direct deformation of previously deposited sediment (both glacial and non-glacial) into drumlinoidal shapes by basal ice contact following smearing-on or sculpting process(es).
2. By anisotropic differences in the subglacial debris under melted-bed conditions owing to: (a) dilatancy (Smalley and Unwin 1968); (b) porewater dissipation; (c) localized freezing; (d) helicoidal basal ice flow patterns (Aario 1977); or (e) localized subglacial debris deformation (Boulton 1987; Menzies 1989). Within this specific group, meltwater activity is of limited impact, whereas porewater is considered critical in local bed debris deformation. Changing stress field and/or stress/strain histories owing to transient basal glacidynamics locally affecting subglacial debris rheology are the important parameters in determining whether drumlins begin to form or not.
3. By the influence of active basal meltwater (under frozen bed conditions) carving cavities beneath an ice mass and later infilling with assorted but predominantly stratified sediment or by the subglacial meltwater erosion of already deposited sediment at the upper ice–bed interface (Dardis and McCabe 1987; Sharpe 1988), or through the entire drumlin, or the sculpting by fluvial processes of previously deposited sediment (Shaw *et al.* 2000). This hypothesis demands meltwater flow of catastrophic discharges from beneath certain areas of an ice mass across the upper ice–bed interface yet permitting the overall ice mass to remain glacidynamically stable. This form of

drumlin development, as with the hypothesis in (1), requires a two-stage process of initiation, beginning with either a pre-formed cavity or pre-existing sediment at the upper ice–bed interface. The latter stage need not be linked directly to the former stage, therefore in some cases although conditions may be suitable for initiation for the first stage, the second stage may not continue toward the critical point (trigger) of drumlin development.

In all these hypotheses the conditions at the subglacial interface(s) are the key to subsequent drumlin formation and, in the long-term, to drumlin 'survival'. A complex relationship must exist between basal glacidynamics, subglacial sediment rheology, and hydraulics for any particular area of ice bed. Fluctuations in state or stress levels or meltwater production and pathways will affect all other parameters to some extent. Certain fluctuations may cross critical thresholds that cannot be reversed, while others may exhibit varying degrees of hysteresis. The likelihood or otherwise of subglacial conditions occurring in any or all these hypotheses remains a fundamental, ongoing, research problem.

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JOHN MENZIES

DRY VALLEY

A valley which is seldom, if ever at the present time, occupied by an active stream channel. Such valleys occur in a wide range of climatic and lithological environments, including extensive areas of Britain and Europe, where they have often been regarded as a product of intense incision under former periglacial conditions (Büdel 1982). There have been many different hypotheses put forward to explain why such valleys are dry (see Table 13; Goudie 1993).

The uniformitarian hypotheses require no major changes of climate or base level, merely the operation of normal processes through time; the marine hypotheses are related to base-level changes; and the palaeoclimatic hypotheses are associated primarily with the major climatic changes of the Pleistocene. British dry valleys (Plate 36) show a considerable range of shapes and sizes, from mere indentations in escarpments, to great winding chasms like Cheddar Gorge in the Mendips. Many, but not all, are formed in carbonate rocks.

Other dry valleys include those that occur in the world's warm deserts and which are relicts of former pluvial conditions and of extensive groundwater sapping (e.g. the MEKGACHA of the Kalahari (Nash 1996). At the other extreme, there are the famous dry valleys of Antarctica, which were cut by former outlet glaciers draining from the Polar Plateau (Summerfield *et al.* 1999).

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Table 13 Hypotheses of dry valley formation

Uniformitarian	
1	Superimposition from a cover of impermeable rocks or sediments
2	Joint enlargement by solution through time
3	Cutting down of major through-flowing streams
4	Reduction in catchment area and groundwater lowering through scarp retreat
5	Cavern collapse
6	River capture
7	Rare events of extreme magnitude
Marine	
1	Non-adjustment of streams to a falling Pleistocene sea level and associated fall of groundwater levels
2	Tidal scour in association with former estuarine conditions
Palaeoclimatic	
1	Overflow from proglacial lakes
2	Glacial scour
3	Erosion by glacial meltwater
4	Reduced evaporation caused by lower temperatures
5	Spring snowmelt under periglacial conditions
6	Runoff from impermeable permafrost

Plate 36 A dry valley, the Manger, developed in the Vale of the White Horse near Wantage, southern England. The valley is developed in Cretaceous chalk and on the left side shows a series of parallel flutes or dells, which some investigators have proposed to be avalanche chutes

Summerfield, M.A., Stuart, F.M., Cockburn, H.A.P., Sugden, D.E., Denton, G.H., Dunai, T. and Marchant, D.R. (1999) Long-term rates of denudation in the Dry Valleys, Transantarctic Mountains, Southern Victoria Land, Antarctica based on in-situ-produced cosmogenic nuclides, *Geomorphology* 27, 113–129.

A.S. GOUDIE

DUNE, AEOLIAN

Aeolian dunes form part of a hierarchical system of bedforms developed in wind-transported sand which comprises: (1) wind ripples (spacing 0.1–1 m); (2) individual simple dunes or superimposed dunes on draa or compound and complex dunes (spacing 50–500 m); and (3) draa or compound and complex dunes (spacing >500 m). Most dunes occur in contiguous areas of aeolian deposits called ergs or sand seas (with an area of >100 km^2). Smaller areas of dunes are called dunefields (see SAND SEA AND DUNEFIELD). The majority of dunes are composed of quartz and feldspar grains of sand size, although dunes composed of gypsum, carbonate and volcaniclastic sand as well as clay pellets also occur.

The formation of areas of dunes is determined by the production of sediment of a range of suitable particle sizes, the availability of this sediment for transport by wind and the transport capacity of the wind (Kocurek and Lancaster 1999). Most dunes are derived from material that has been transported by fluvial or littoral processes. Important sources include marine and lacustrine beaches, dry lake basins, river floodplains and deltas. The availability of sediment for transport by wind is determined by its moisture content, vegetation cover, crusting and cohesion. The transport capacity of the wind is a cubic function of wind speed or surface shear stress above the transport threshold (see AEOLIAN PROCESSES). These conditions are satisfied in two main environments: (1) coastal areas with sandy beaches and onshore winds (e.g. the Atlantic coasts of north-west Europe, the Pacific north-west of North America, south-east and northeastern Australia and southern South Africa); and (2) subtropical and temperate desert areas.

Dune types

Aeolian dunes develop as a result of interactions between a granular bed (sand) and turbulent shearing flow (the atmospheric boundary layer). The resulting landforms are bedforms that are dynamically similar to those developed in subaqueous shearing flows (e.g. rivers, tidal currents). The morphology of aeolian dunes therefore reflects the characteristics of the sediment (primarily its grain size) and the wind (both the local shear stress, which determines local sand transport rates, and the long-term directional variability of the wind regime). Vegetation is a significant factor influencing the morphology of dunes in coastal dunefields as well as those in semi-arid and subhumid regions. Interactions with topographic obstacles may also result in dune formation.

Dunes occur in self-organized patterns that develop over time as the response of sand surfaces to the wind regime (especially its directional variability) and the supply of sand (Werner 1995). Development of these patterns is modulated by the effects of changes in climate and sea level on sediment supply, dune mobility and wind regime characteristics, often resulting in the formation of a series of different generations of dunes. The dune types described below represent the steady-state attractors of the aeolian transport system and can evolve from a wide range of initial conditions. The orientation of dunes with respect to the wind regime is another aspect of the self-organizing nature of the system, in which dunes are oriented to maximize the gross sand transport normal to the crest. Characteristic features of dune patterns include close correlations between the height and spacing of dunes and systematic spatial variations in dune type, orientation and sediment volume.

Despite the variety of different dune types and the multiplicity of local names that have been used to refer to them, satellite images show that dunes of essentially similar form occur in widely separated areas, and occur in five main morphologic types (Figures 50 and 51). The only dune form restricted to coastal areas is the foredune because it is an integral part of the complex of near shore processes forming the beach–dune system (Bauer and Sherman 1999). Three varieties of each dune type can occur: simple (the basic form), compound (superimposition of small dunes of the same type on larger dunes), and complex (superimposition of different dune types on the primary form, (e.g. crescentic dunes on linear dunes).

Relations between the occurrence of different dune morphological types and their wind regime environment indicate that the main control of dune type is the directional variability of the wind regime (Figure 52), which can be characterized by the ratio between the resultant (vector sum) of potential sand transport (RDP or resultant drift potential) and total potential sand transport (DP or drift potential). Sand grain size, vegetation cover, topography and sediment supply play subordinate roles in the majority of cases.

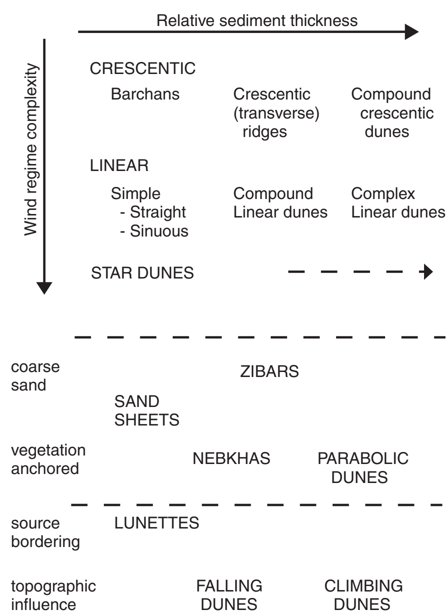

Figure 50 The main dune morphological types

The simplest dune types and patterns form in areas characterized by a narrow range of wind directions (unidirectional wind regime, RDP/DP > 0.8). In the absence of vegetation, the dominant form will be crescentic or transverse dunes with crest lines aligned approximately perpendicular to the dominant wind. Good examples are to be found in coastal areas of Namibia and the United Arab Emirates. Isolated crescentic dunes or barchans occur in areas of limited sand supply, and coalesce laterally to form crescentic or barchanoid ridges that consist of a series of connected crescents in plan view as sand availability increases. Larger forms with superimposed dunes are termed compound crescent dunes (e.g. Algodones Dunes, California; coastal areas of the Namib Sand Sea). In areas of partial vegetation cover and similar wind regimes, parabolic dunes will occur. These dunes are characterized by a U or V shape with a 'nose' of active sand and two partly vegetated arms that trail up wind. They are common in many coastal dunefields and semi-arid inland areas and often develop from localized blowouts in vegetated sand surfaces (Wolfe and David 1997). Both crescentic and parabolic dunes tend to migrate downwind, at a rate that is inversely proportional to their height.

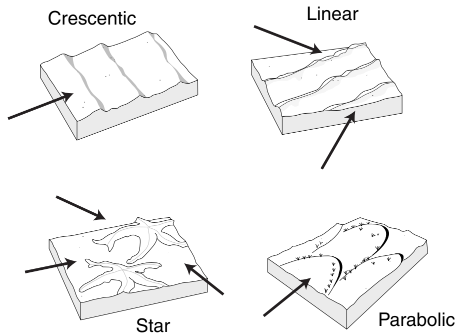

Figure 51 Schematic morphology of major dune morphological types and wind regime environments (modified from Lancaster 1995)

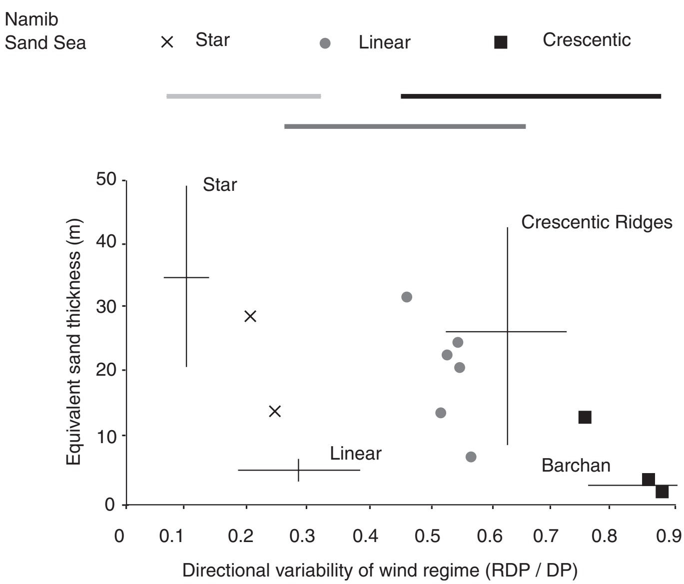

Figure 52 Relations between dune types and wind regimes. Redrawn from Wasson and Hyde (1983) with Namib Sand Sea data superimposed (symbols) and range of wind regimes for three major dune types from Fryberger (1979). Equivalent sand thickness is a measure of the sand available for dune building and represents the thickness of sand if the dunes were levelled. The directional variability of the wind regime is characterized by the ratio between resultant (RDP) and total sand drift potential (DP)

Linear dunes are characterized by their straightness, length (often more than 20 km), sinuous crestline, parallelism, and regular spacing, and high ratio of dune to interdune areas. Many linear dunes consist of a lower gently sloping plinth, often partly vegetated, and an upper crestal area where sand movement is more active. Slip faces develop on the upper part of the dune, their orientation depending on the winds of the season. The average form of the dune may be symmetrical with an approximately triangular profile, but in each season its profile tends to an asymmetric form with a concave stoss slope and a well-developed lee face. Linear dunes occur in areas of bimodal or wide unimodal wind regimes (RDP/DP $> 0.4 < 0.8$), and appear to be the most widespread dune type worldwide. Simple linear dunes occur in two forms: the straight partially vegetated dunes of the southwestern Kalahari and Simpson–Strezlecki deserts and the more sinuous 'seif'-type dunes of the Sinai and eastern Sahara. Complex linear dunes are best represented by the large (50–200-m high), widely spaced (1–2 km) linear dunes of the Namib Sand Sea.

The origins of linear dunes and their relationship to formative wind directions have been the subject of considerable controversy. A widely held view was that linear dunes form parallel to the prevailing or dominant wind direction. Their parallelism and straightness was believed to result from the existence of boundary layer roller vortices in which helicoidal flow sweeps sand from interdune areas to dunes. However, there is little empirical evidence to support such a model. Field studies of airflow and sediment transport over linear dunes (Bristow *et al.* 2000; Tsoar 1983) suggest that the fundamental mechanism for linear dune formation is the deflection of winds that approach at an oblique angle to the crest to flow parallel to the dune along its lee side and transport sand along the dune. Thus any winds from a 180° sector centred on the dune will be diverted in this manner. Linear dunes tend to extend downwind, as sinuosities in the crest migrate along its length. Evidence for lateral migration, is not conclusive.

Star dunes have a pyramidal shape, with three or four sinuous sharp-crested arms radiating from a central peak and multiple avalanche faces and are the largest dunes in many sand seas, reaching heights of more than 300 m in the eastern Namib Sand Sea and the Grand Erg Oriental in Algeria. The upper parts of many star dunes are very steep with slopes at angles of 15–30°; the lower parts consist of a broad, gently sloping (5–10°) plinth or apron. Small crescentic or reversing dunes may be superimposed on the lower flank and upper plinth areas of star dunes. Comparisons between the distribution of star dunes and wind regimes suggest that they form in multidirectional or complex wind regimes (RDP/DP < 0.3). A strong association between the occurrence of star dunes and topographic barriers has also been noted. Topography may modify regional wind regimes to increase their directional variability, as in the Erg Fachi Bilma or create traps for sand transport, as at Kelso Dunes and Great Sand Dunes.

The development of star dunes is strongly influenced by the high degree of form–flow interaction that occurs as a result of seasonal changes in wind direction, and the existence of a major lee-side secondary circulation. Most of the erosion and deposition involves the reworking of deposits deposited in the previous wind season. Sand, once transported to the dune, tends to stay there and add to its bulk, resulting in dunes that do not change position over time (Lancaster 1989).

Other important dune types include nebkhas or hummock dunes (common in many coastal dunefields) anchored by vegetation, lunettes (often comprised of sand-sized clay pellets) that form downwind of small playas; and a variety of topographically controlled dunes (climbing and falling dunes, echo dunes). Low relief sand surfaces such as sand sheets are common in many ergs and occupy from as little as 5 per cent of the area of the Namib Sand Sea to as much as 70 per cent of the area of Gran Desierto. Sand sheets occur where sediment availability is limited as a result of coarse sand, high water table or vegetation cover. Zibar, or low rolling dunes without slip faces composed of coarse sand, are transitional between sand sheets and crescentic dunes in some dunefields (e.g. Algodones, Skeleton Coast, Namibia).

Dune processes and dynamics

The initiation, development and equilibrium morphology of all aeolian dunes are determined by a complex series of interactions between dune morphology, airflow, vegetation cover and sediment transport rates. In turn, the developing bedforms exert a strong control on local transport rates through form–flow interactions and secondary flow circulations, leading to a dynamic equilibrium between dune morphology and local airflow. In multidirectional wind regimes, the nature of interactions between dune form and airflow change as winds vary direction seasonally, and lee-side secondary flow patterns become important in determining dune morphology and dynamics.

As dunes grow, they project into the atmospheric boundary layer so that they affect the airflow around and over them in a manner similar to isolated hills. Winds approaching the upwind toe of a dune stagnate slightly and are reduced in velocity, but likely not turbulence intensity. On the stoss, or windward slope of the dune, streamlines are compressed and winds accelerate up the slope. The degree of flow acceleration (the speed-up factor) is determined by the aspect ratio and the height of the dune. Wind speed at the crest of the dune is typically 1.1 to 2.5 times that measured immediately upwind of the dune (Figure 53a). Flow acceleration, coupled with effects of stream line curvature, on the windward slopes of dunes give rise to an exponential increase in sediment transport rates (Figure 53b) towards the dune crest (Lancaster *et al.* 1996; McKenna Neuman *et al.* 1997), resulting in erosion of the stoss slope, and a high level of erosion and deposition in crestal areas of linear and star dunes (e.g. Lancaster 1989; Livingstone 1989). Numerical models suggest that the non-linear increase in sediment transport with height on a dune limits dune size and results in an equilibrium dune configuration.

In the lee of the crest of dunes, wind velocities and transport rates decrease rapidly as a result of flow expansion between the crest and brink of the lee or avalanche face and flow separation on the avalanche face itself. There is a complex pattern of flow separation, diversion and re-attachment on the lee slopes of dunes, which is determined by the angle between the wind and the dune crest (angle of attack) and the dune aspect ratio (Walker and Nickling 2002). Secondary flows, including lee-side flow diversion, are especially important where winds approach the dune obliquely, and are an important process on linear and many star dunes.

High angles of attack on high aspect ratio (steep) dunes result in flow separation in the lee, while lower angles of attack result in flow diversion along the lee slope, whereas low aspect ratio dunes are characterized by flow expansion. Flow separation results in the development of an eddy in the lee of the dune, which may have the form of a roller vortex if flow is truly transverse, with the separation cell extending downwind for 4 to 10 times the height of crescentic dunes. When flow is oblique to the dune crest a helical vortex develops. The oblique flow is deflected along the lee slope parallel to the dune crest, with the degree of deflection being inversely proportional to the incidence angle between the crestline and the primary wind. Field studies indicate that the lee-side helical eddy affects the whole of the lee side on simple (5–10-m high) linear dunes, but extends for only 10–20 per cent of the height of large (50–150-m high) complex linear dunes and 40 m high star dunes. Changes in the local incidence angle between primary winds and a sinuous dune crest result in a spatially varying pattern of deposition and along-dune transport on the lee face. Deposition dominates where winds cross the crest line at angles approaching 90°, and erosion or along-dune transport occurs where incidence angles are $<40°$. Studies on flow-transverse dunes indicate that, downwind of and above the flow separation cell, there is a series of wakes that gradually expand, diffuse and mix downwind for a distance of as much as 25 to 30 times the height of the dune (Walker and Nickling 2002).

Flow separation also causes fallout of previously saltating sand grains. Field experiments show that 95 per cent of the sand transported over the crest is deposited within a metre of the crest, with the rate of deposition decreasing exponentially downwind (Nicking *et al.* 2002). High rates of deposition in the immediate lee of the crest result in oversteepening of the slope and avalanching of grains to form slip faces. All sand transported over the crest of crescentic dunes is deposited, so that they are typically 'sand trapping' bedforms. As a result, their movement can be described by: $c = Q/yh$ where c is the migration rate, Q is the bulk volumetric sand transport rate, y is the bulk density of sand and h is dune height.

Challenges and opportunities in dune studies

The past two decades have seen a dramatic change in the level of understanding of dune dynamics and morphology through intensive field studies of processes and synoptic views of sand seas provided by satellite images. As a result, the fundamentals of dune dynamics and the formative factors of major dune types are known in

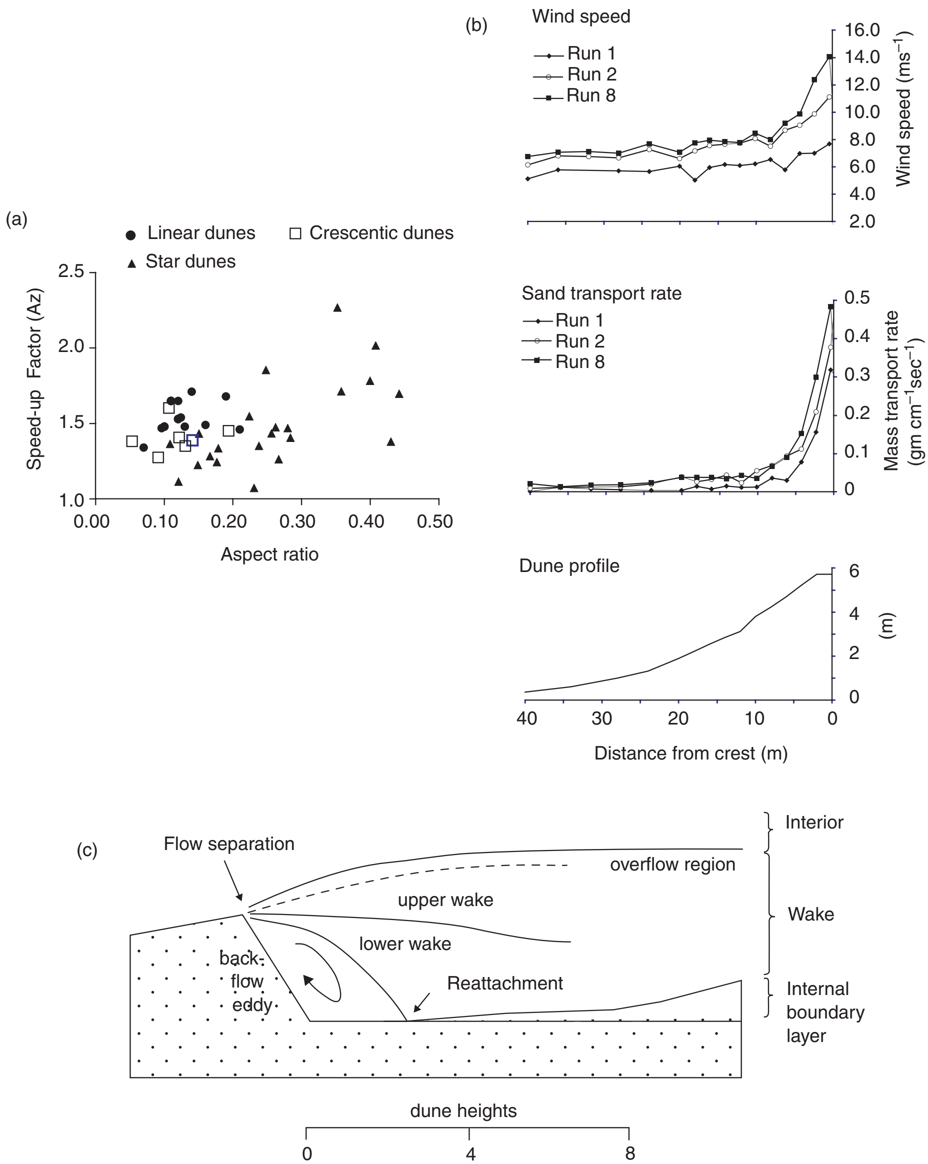

Figure 53 Elements of dune dynamics: (a): velocity speed-up (from Lancaster 1995); (b) winds and sediment transport rates on the stoss slope (from McKenna Neuman *et al.* (1997) (c) lee-side flow separation and wake mixing (from Walker and Nickling 2002)

some detail. Not well known are processes leading to dune initiation, the dynamics of lee-side processes (including avalanching), and the controls of dune size and spacing. It has also proved very difficult to extrapolate the results of short-term studies of dune processes to understanding of long-term or even annual dune dynamics. One promising approach is to develop numerical models of dune and dunefield evolution (Werner 1995). The other is to use ground-penetrating radar to image dune sedimentary structures, which provide a record of the results of dune-forming processes on a variety of timescales and allow empirical models of dune evolution to be developed. A good example of this approach is the five-stage model of linear dune development put forward by Bristow *et al.* (2000).

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SEE ALSO: aeolian processes; barchan; draa (megadune); dune mobility

NICK LANCASTER

DUNE, COASTAL

Foredune

Foredunes are shore-parallel dune ridges formed on the top of the backshore by aeolian sand deposition within vegetation. They may range from scattered hummocks or nebkha, relatively flat terraces, to markedly convex ridges. Actively forming foredunes occupy a foremost seaward position, but not all foremost dunes are foredunes. Other dune types may occupy a foremost position on eroding coasts or coasts where foredunes are unable to form. Foredunes generally fall into two main types, incipient and established foredunes.

INCIPIENT FOREDUNE

Incipient foredunes are new, or developing foredunes forming within pioneer plant communities. They may be formed by sand deposition within discrete or relatively discrete clumps of vegetation or individual plants forming shadow dunes, hummocks or nebkha.

Incipient foredunes may also form on the backshore by relatively laterally continuous alongshore growth of pioneer plant seedlings and/or rhizomes in the wrack line or spring high tide region (Hesp 1989). Morphological development principally depends on plant density, height and cover, wind velocity and rates of sand transport (Davies 1980).

Shadow dunes, hummocks, embyro dunes and nebkha all form due to high localized drag within and behind individual plants and clumps of plants. Wind velocities experience rapid deceleration on reaching the plants, local acceleration around the plants, and flow separation behind the plants (Hesp 1999).

Relatively continuous plant canopies variously impact the wind/sand flow depending on plant density, distribution and height. High, dense canopies act to reduce flow velocities very rapidly, and sand transport (saltation and traction) is markedly reduced from the leading edge. In canopies which vary alongshore in density or distribution, foredune morphology also varies (Nickling and Davidson-Arnott 1990). Plant density is increased as wind velocities increase as the vegetation bends and streamlines to the wind.

Incipient foredunes generally display one of three morphological types: ramps, terraces and ridges. Swales (lee dune depressions) are generally created by seaward accretion of a foredune. They develop as low to limited aeolian deposition zones (Hesp 2002).

ESTABLISHED FOREDUNE

Established foredunes develop from incipient foredunes and are commonly distinguished by the growth of intermediate or 'secondary' plant species, and/or by their greater morphological complexity, height, width, age and geographical position.

Foredunes range from very low, and commonly scattered, dunes less than a metre or so in height on some barrier islands dominated by overwash and in areas of limited sediment supply to over 30 m in height in some instances. The morphological development and evolution of established foredunes depends on a number of factors including: sand supply, beach width and fetch, surfzone-beach type, the degree of vegetation cover, plant species present (a function of climate and biogeographical region), the rate of aeolian sand accretion and erosion, the frequency and magnitude of wave and wind forces, the occurrence and magnitude of storm erosion, dune scarping and overwash processes, the medium to long-term beach or barrier state (stable, accreting or eroding), and increasingly, the extent of human interference and use (Davidson-Arnott and Law 1996; Short and Hesp 1982; Hesp 1999).

The wind flow is topographically accelerated over foredunes, particularly up stoss slopes and over crests. However, the variable vegetation cover of foredunes and their topographic variability leads to local decelerations and variations in roughness length (Arens 1997), and these become more pronounced as foredune morphological complexity and vegetation cover increases.

Foredune Plain

Foredunes may gradually, or rapidly, become isolated from accretion and erosion processes by the seaward development of a new incipient foredune which itself may evolve into an established foredune. The original foredune then becomes a relict foredune as it is largely or wholly removed from a foremost beach position. Systematic beach progradation over time frames of tens to thousands of years has led to the development of wide foredune plains.

Blowout

A blowout is a saucer-, cup-, bowl or trough-shaped depression or hollow formed by wind erosion on a pre-existing sand deposit. The adjoining accumulation of sand, the depositional lobe, derived from the depression and possibly other sources, is normally considered part of the blowout (Nordstrom *et al*. 1990).

Blowout morphology may be highly variable, ranging from cigar-shaped, V-shaped, scooped hollow, and cauldron and corridor types, from pits to elongated notches, troughs or broad basins, and saucer and trough blowouts (Cooper 1967). Saucer blowouts are semicircular or saucer-shaped and often appear as shallow dishes. Deeper cup- or bowl-shaped blowouts may evolve from these. Trough blowouts are generally more elongate, with deeper deflation floors and basins, and with steeper, longer erosional lateral walls or slopes.

INITIATION

Blowouts may be initiated in a variety of ways including:

1. wave erosion of dunes followed by wind erosion of dune scarps;
2. die-back of vegetation following dune wave erosion and subsequent wind erosion;
3. wind erosion of overwash hollows and fans;
4. topographic acceleration of airflow over (or through) dunes, dune cols, scarps and cliffs;

5 where the vegetation cover is naturally low, or is weakened, reduced or dies due to a prolonged dry or arid period;
6 vegetation die-back due to soil nutrient depletion;
7 localized aridity (e.g. on dune crests) reducing plant cover;
8 the activities of animals and humans;
9 water erosion;
10 high velocity wind erosion leading to either erosion, or sand inundation and burial.

Once initiated, the subsequent morphologic development may depend on the size of the initial constriction, the height and width of the dune in which the blowout is developing, the degree and type of vegetation cover, the magnitude of regional winds, and the degree of exposure to winds from various directions (Hesp 2002; Jennings 1957).

FLOW DYNAMICS

Flow in saucer blowouts is complex with flow separation occurring around much of the erosion walls. Sand erosion and deposition are also complex as a result of varying wind speeds and directions, although, in general, deflation basins deepen in most blowouts studied. Saucer blowouts commonly grow in length upwind against the prevailing wind.

The flow up trough blowouts is commonly topographically accelerated, and displays marked single and double jets up the deflation basin, corkscrew vortices over the lateral erosional wall crests, and rapid flow deceleration, lateral expansion and flow separation over the depositional lobe. Topographic steering can be significant (Hesp and Hyde 1996).

Parabolic dune

Parabolic dunes (also termed U-dunes, upsiloidal dunes, hairpin dunes) are typically U-and V-shaped dunes characterized by short to elongate, trailing ridges which terminate downwind in U- or V-shaped depositional lobes. The depositional lobes may be simple, relatively featureless sandsheets, or textured with a variety of dune forms (e.g. transverse dunes, barchanoidal dunes, etc). Deflation basins and plains, slacks, seasonal wetlands, ponds, lagoons and gegenwalle ridges occupy the area between the trailing ridges.

INITIATION

Parabolic dunes typically evolve in a number of ways including:

1 from blowouts (Pye 1983). In many cases, the blowout depositional lobe continues to advance downwind forming trailing ridges;
2 evolution from the landward and downwind margins of transgressive sandsheets and dunefield.

Blowouts and parabolic dunes may be formed on both stable (sediment supply balanced) and accreting/prograding (positive sediment supply) coasts which experience occasional or regular high energy wind events (Hesp 2002). They are commonly formed on eroding coasts where the foredune stability is reduced by wave erosion, and subsequent wind erosion (e.g. Ruz and Allard 1994).

MORPHOLOGY

Two principle sub-types of parabolic dune are common: long-walled types and squat, elliptical types. The multiple development of these leads to there being two principle sub-types of parabolic dunefields: long-walled types and imbricate types (Trenhaile 1997).

Long-walled parabolic dunes display long trailing ridges and extensive deflation basins. They are particularly well developed on relatively flat terrain, in regions of low heath or shrubland, high sand supply and strong, more unidirectional winds. Some parabolic dunes display a squat, shorter form, often with more semicircular or elliptical deflation basins. Multiple development results in the dunes overlapping each other in an imbricate fashion. They commonly develop in wetter areas, on flat terrain where deflation depths are limited and/or wind speeds are relatively low, on steep terrain, in less unidirectional or multidirectional wind regimes and/or in dense, tall vegetation where the rate of advance is low and/or migration is impeded.

EVOLUTION

Deflation basins tend to continue to erode until a base level is reached such as the seasonally lowest water-table level, a calcrete (or other cemented/indurated) layer, an armoured surface such as a pebble, shell, pumice or artifact surface. Trailing ridges develop due to trapping of the outside, marginal lateral edge of the depositional lobe as it migrates downwind. This sediment is trapped while the inside (deflation plain) portion of the ridge is eroded. Depositional lobes are arcuate, hairpin, V-shaped, radial or parabolic-shaped depending on wind direction, lobe height and volume, vegetation cover and species type, and speed of migration.

RATES OF MIGRATION

Rates of parabolic dune advance or migration vary considerably depending on the morphology, slope and type (e.g. sandy vs rocky) of terrain the dunes are moving across, the vegetation cover and type (e.g. woodland vs grassland), wind velocities and directional variability of the wind. Dune migration rates range from 0.05 to 25 $\mathrm{m\,yr^{-1}}$.

Transgressive dunefield and sheet

Transgressive dunefields and sheets are aeolian sand deposits formed by the downwind or alongshore movement of sand over vegetated to semi-vegetated terrain. Such sheets and dunefields may range from quite small (hundreds of metres in alongshore and landward extent) to draa or megadune size fields. They may be completely unvegetated, partially vegetated or fully vegetated (post-formation) (Nordstrom *et al.* 1990). Sheets display little or no surface dunes; dunefields are covered with a variety of superimposed dune forms. They have also been termed mobile dunes, migratory dunes, mendano and machair.

Transgressive dunefields are particularly well developed on high wind and wave energy (west and south) coasts with significant sediment supply, and in virtually all climatic regions (tropics to the arctic).

TRANSGRESSIVE DUNEFIELD TYPES

At the gross scale, transgressive dunefields may describe tabular forms (including headland bypass dunefields), buttress forms, or climbing, clifftop and falling dunefields.

INITIATION AND DEVELOPMENT

Transgressive dunefields develop for a variety of reasons. They may form, or have formed:

1 as a response to rising sea level and/or climatic change, particularly in the period, 10,000 to 7,000 years BP.
2 in regions of high alongshore and onshore sediment supply, often in high wind and wave energy environments;
3 on coasts experiencing erosion;
4 as continental shelves were exposed and/or climate changed during the Last Glacial;
5 as a response to periods of regional sea-level fall; and
6 on coasts experiencing climatic extremes such as in arid and arctic and subarctic environments, and where vegetation growth may also be limited.

TRANSGRESSIVE DUNEFIELD LANDFORMS

Active transgressive dunefields may extend (and/or migrate) directly alongshore, obliquely onshore or directly onshore. Dunefields migrating alongshore are typically characterized by transverse (and other) dunes extending inland from the backshore. Interdunes are dominated by nabkha, deflation flats, sandsheets and overwash plains and fans. Dunefields migrating obliquely or normally onshore are usually characterized by a small to extensive deflation basin (or plain) or a series of slacks on the upwind or seaward side, a small to extensive, mobile to partially vegetated sandsheet or dunefield, and a long-walled, commonly sinuous 'main' slipface or precipitation ridge on the landward side.

The surfaces of active transgressive dunefields are commonly covered with a variety of dune types (including 'desert' dune types) ranging from simple domes, transverse dunes and BARCHANs, to barchanoidal and sinuous transverse and oblique dunes, to complex aklé or network and star dune forms (e.g. Hunter *et al.* 1983).

Deflation plains and basins typically lie parallel to the shore and are eroded down to a base level such as a CALCRETE pavement, the seasonally lowest water table, older dune surfaces and PALAEOSOLs, carbonate or bedrock.

Active transgressive dunefields may display a variety of generally smaller scale dune forms and environments, including remnant knobs, hummocks, bush pockets, nabkha, shadow dunes and 'rim dune' (around the margins of washover fans) (Nordstrom *et al.* 1990).

Precipitation ridges (long-walled or main slipfaces) commonly occur along the downwind and surrounding margins of transgressive dunefields. Where the dunefields are migrating in one primary direction, they generally have one precipitation ridge. Where they are expanding/migrating landwards and alongshore, they may have two to many precipitation and trailing ridges.

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PATRICK HESP

DUNE, FLUVIAL

Dunes are the commonest sedimentary feature in sand- or silt-bedded streams. They are roughly triangular in profile with a gentle upstream slope and a steeper downstream slope (see DUNE, AEOLIAN). Dune height and wavelength are directly related to water depth. Reaching heights of up to one-third of flow depth they are commonly 0.1 m to 10 m high with a wavelength 4 to 8 times flow depth (Knighton 1998). They frequently form in streams with higher intensity flows than those with RIPPLE bedforms but, like ripples, they migrate downstream through the processes of erosion on the upstream slope followed by deposition on the downstream slope. Separation of flow from the crest of dunes generates large-scale turbulence in rivers and the downstream migration and change of dune form is an important mechanism of bedform adjustment to changing river discharge. Cross-bedding structures resulting from dune migration are often preserved in alluvial deposits and can be used to interpret palaeohydraulic processes.

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GILES F.S. WIGGS

DUNE MOBILITY

Approximately 20 per cent of the world's drylands are covered by aeolian sands, within which desert dunes (see DUNE, AEOLIAN) occur, while coastal dunes occur in a range of climatic settings. In dryland and coastal settings dunes may be mobile, but many dunes are not mobile in the sense of the dune body migrating across the landscape. Dune mobility therefore needs to be considered in terms of dune setting, dune type, and the nature of aeolian activity upon the dune surface.

The mobility of sand dunes is in broad terms a function of the relationship between the forces of erodibility, affecting the potential of the dune surface to be eroded by the wind, and erosivity, which is the potential of the agencies effecting AEOLIAN PROCESSES to move sediment. Dune mobility may be assessed from climatological data using a dune mobility index (e.g. that of Lancaster 1988) that integrates the forces that affect erodibility, such as P/PET, and those that affect the erosivity, which relate to the wind field. Mobility can also be assessed in the field, by monitoring dune surface change and airflow (e.g. Wiggs *et al*. 1995) or dune movement in the landscape (e.g. Hastenrath 1987)

Since dunes, except those that have become lithified, are mainly comprised of unconsolidated sand-sized sediment, their potential to be moved by mobile air via processes of sediment entrainment may appear to be considerable. Reality is more complex, however, since winds have to exceed a threshold velocity for entrainment to take place, and dune bodies can, once formed, store moisture and act as a host to plants, which can markedly reduce erodibility. Dune mobility can be a discontinuous process too. Winds capable of entraining and moving sediment do not occur continuously, but vary seasonally and daily. For example, sand transport on dunes in the Namib Sand Sea generally occurs in response to fairly persistent but moderate south westerly winds during summer months, while in winter short-lived but high magnitude wind events transport sediment from an easterly direction (Livingstone 1989).

Vegetation and dune mobility

Isolated or widely spaced plants on a dune can lead to localized zones of higher wind velocities as airflow is streamlined around the obstacle (Thomas and Tsoar 1991). But in general, dunes that possess some form of surface vegetation or crusting have, all other things being equal, a lower erodibility than those that do not. Crusts and plants can play several roles in affecting the potential mobility of surface sediments (Wolfe and Nickling 1993): protection of the sediment immediately below the plant or crust, increasing surface roughness and thereby reducing wind velocity, and trapping any moving sediment grains. A partial or discontinuous vegetation cover does not totally exclude sand movement but it may anchor a dune plinth and inhibit dune migration or lateral movement. On partially vegetated dunes, different studies have identified various threshold vegetation covers, ranging from *c*.6 per cent (Marshall 1970) to 30 per cent (Ash and Wasson 1983) above which any aeolian activity ceases. However, the impact of a given cover will vary not only according to plant shape and porosity but to both ambient wind velocities and position on the dune body (Wiggs *et al*. 1995).

Dune size and mobility

All other things being equal, smaller dunes move or experience surface change more quickly than large dunes. This is because, for a given sediment transport event, the volume of sand that can be moved represents a smaller component of the total volume of sand of a large dune than of a small dune. The ability of a dune to retain its form and position as environmental conditions change has been called 'dune memory' by Warren and Kay (1987), with small dunes of low volume having little memory, and therefore adjusting relatively rapidly to wind events, while large dunes with large volumes have 'mega memories' that may record histories spanning millennia.

Mobility of different dune types

Different basic dune forms develop in different wind directional regimes (Fryberger 1979; Thomas 1997). Generally, barchans and transverse dunes form in unimodal sand transporting wind regimes, linear dunes in bimodal or wide unimodal regimes, and star dunes in multimodal regimes, where regime refers to the overall annual directional pattern of sand transporting winds.

These different regimes determine the general types of mobility or, more appropriately activity, of these dune types. Transverse dunes are mobile in the true sense of the word, since with transport for a single direction the dunes are able to migrate. Migration rates differ between and within dunefields according to the available transport energy, but given the principle of dune memory (see above), in any location larger dunes will move more slowly than small dunes, as expressed by

$$c_r = (q_c - q_t)/h\gamma_\rho$$

where c_r is the migration rate, q_c is the mass transport rate at the dune crest, q_t is the mass transport rate at the dune base, h is dune height and γ_ρ is the bulk density of the sediment. A number of studies of migration rates have been conducted in different deserts, and are summarized in Thomas (1992) with examples of rates given in Table 14.

Linear dunes are extending forms. Net sediment transport is along the dune in the resultant direction of transport generated by the combined effect of bimodal winds. This can lead to elongation of the dune at the downwind end, but also to some lateral movement if one direction has greater transport potential than the other and if the dune plinth is not anchored by vegetation. Lateral migration can be extremely slow, for example at a rate of 50–100 m over the past 10,000 years as suggested by Rubin (1990) from evidence in the Strezlecki Desert in Australia. Other studies from Namibia and the Sinai Desert

Table 14 Examples of barchan and transverse dune migration rates

Location	Dune height (m)	Migration rate (m yr $^{-1}$)
Barchan dunes, southern Peru	1	32
	7	9
Barchan dunes, Salton Sand Sea, California	3.1	27
	8.2	14
Transverse dunes, Erg Oriental, Saudi Arabia	35	0.3
	240	0.16

Source: Data from various authors

suggest elongation rates may range from less than 2 m to over 14 m per annum.

Star dunes can be regarded as sand accumulating forms that, under the interactive effect of sand transport from at least three directions, gain in volume and height over time. The individual arms of the dune may, on a seasonal basis, behave as if they are transverse or linear forms and display displacements of up to 20 m (Lancaster 1989). If any of the contributory transport directions has a net advantage over the others, some migration of the dune body may occur over time.

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DAVID S.G. THOMAS

DUNE, SNOW

Aeolian bedform are common in snow, and include ripples, drifts, barchans, and the like (Cornish 1914).

In recent years the size and importance of various megadunes have become appreciated, particularly in eastern Antarctica. These are transverse features that are oriented perpendicular to the regional katabatic wind direction. Their amplitudes are small (*c*.4 m), but their wavelengths range from 2 to over 4 km, and megadune crests are nearly parallel and 10–100 km in length (Frezzotti *et al*. 2002).

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A.S. GOUDIE

DURICRUST

The word was introduced by Woolnough (1927) who subsequently defined the term thus (Woolnough 1930: 124–125): ‘The widespread chemically formed capping in Australia, resting on a thoroughly leached sub-stratum . . . The nature of the deposit varies from a mere infiltration of pre-existing surface rock, to a thick mass of relatively pure chemical precipitate’.

The mineral matter deposited from solution falls into three main groups: (1) aluminous and ferruginous; (2) siliceous; and (3) calcareous and magnesian. Woolnough believed that bedrock was an important influence on the distribution of these three types, which in effect are broadly equivalent to (1) laterites, bauxites, FERRICRETES (see Tardy 1997; Bardossy and Aleva 1990); (2) SILCRETES; (3) CALCRETES, dolocretes.

Because of subsequent work on the individual duricrust types, the *crete*-based terminology of which had been laid down by Lamplugh (1907), Goudie (1973: 5) proposed a modified definition which resulted from a synthesis of various definitions that had already been developed for the individual types, and stressed their essentially subaerial and near-surface origin and nature:

> A product of terrestrial processes within the zone of weathering in which either iron and aluminium sesquioxides (in the case of ferricretes and alcretes) or silica (in the case of silcrete) or calcium carbonate (in the case of calcrete) or other compounds in the case of magnesicrete and the like have dominantly accumulated in and/or replaced a pre-existing soil, rock, or weathered material, to give a substance which may ultimately develop into an indurated mass.

Sometimes duricrusts may incorporate characteristics of more than one type, as with the widespread calsilcretes of the Kalahari.

To understand the origin and development of these geomorphologically important materials some general considerations need to be borne in mind. First, there is the question of the sources of the materials which contribute to the make-up of duricrusts. The primary elements can be derived from at least four main sources: the weathering of bedrock and sediment, inputs from dust and precipitation, plant residues and the dissolved solids in ground water. Then these sources have to be translocated and concentrated either by lateral transfers, or by vertical movements, whether upwards (*per ascensum*) or downward (*per descensum*). Third, the transferred materials need to be precipitated, and here a very wide range of processes come into play. Among the most important of these are changes in chemical equilibria caused by evaporation, by temperature changes, by pressure changes in the soil, air and water systems, by the action of organisms and by miscellaneous changes caused by interactions of different solution types.

Models for the origin of duricrusts normally fall into one of two categories: those involving relative accumulation and those involving absolute accumulation. *Relative accumulations* owe their concentrations to the removal of more mobile components, while *absolute accumulations* owe their concentration to the addition of materials to a profile. However, as McFarlane (1983: 20) has pointed out, the utility of this subdivision depends on important scale considerations. At one extreme the accumulation is entirely relative since laterites would not exist at all were not Fe and Al less readily mobilized during rigorous chemical weathering. At the other extreme, in hand specimens even the residual laterites on interfluves show much addition of Fe, since samples are enriched absolutely in materials which originated above them in the formerly existing column of rock, consumed to provide the residuum.

Furthermore, laterites and silcretes differ in that, while ferricretes can result from either relative or absolute accumulation of iron, silcrete can only form by absolute accumulation. Weathering provides the silica and in some cases the material (a weathering profile, for example) in which the silica is deposited.

Many of the early models of duricrust formation involved vertical processes, and especially the role of capillary rise of solutions from ground water. Vertical process models of this *per ascensum* type were complemented by *per descensum* models, in which it was believed that material leached from the upper part of a profile would accumulate lower down. Some of the material to be leached downward might be added to the top of the profile in the form of inputs of dust, etc.

However, more recently appreciation of the importance of CATENAs and toposequences, and of lateral soil-water movements, has resulted in an increasing concern with lateral transfer models. For example, Stephens (1971) argued that the silcretes of inland Australia formed from silica that was leached during lateritization in the humid upland areas of the east and then transported by rivers to low relief areas lying to the west. Similarly the detrital model of calcrete formation (Goudie 1983: 115) involves the lateral transportation and redeposition of weathered fragments of calcrete, moving from plateaux surfaces to footslopes.

One slightly unusual explanation for duricrust formation is that proposed for the silcretes of parts of Australia, where, it has been suggested, overlying or adjacent basalt sheets have played a role. Even amongst those who have proposed this association there is little agreement as to whether the supposed basaltic effect has been hydrothermal alteration, contact metamorphism, a release of silica from weathered basalt, or a reduction in the migration of pore waters caused by the presence of a basaltic caprock. Some doubt, however, whether such a special mechanism is justified (e.g. Ollier 1991) for what is such a widespread phenomenon.

Another general feature of models of duricrust formation has been the appreciation of the importance of organic processes. For example, in the case of calcrete, laboratory simulations with micro-flora (Krumbein 1968), and studies of petrography which have revealed calcified organic filaments of soil fungi, algae, actinomycetes and root hairs of vascular land plants, have caused the role of organisms to be given the attention they deserve (see Goudie 1996, for a review). In the case of laterite, various organic agencies have also been mooted. Micro-organisms could contribute to both mobilization and precipitation of materials. The transition from geothite to haematite in laterite profiles could be the result of iron bacteria activity, and desilicifying bacteria could be used to remove combined silica (kaolin) from bauxite.

Several factors contribute to the geomorphological importance of duricrusts: the thickness of the profiles, the properties of the different components of the profiles (e.g. their occasional ability to harden on exposure) and the topographic situation in which duricrusts develop. Ferricrete profiles may be as much as 60 m thick. Calcrete profiles in parts of southern Africa, western Australia and the Texan High Plains may exceed 40 m, while in Zaire and Namibia maximum depths of silicification may also be of the order of 50 m.

The hard upper crusts of duricrust profiles form only a limited proportion of the total profile thickness. Typical values for alcrete and ferricrete hardpans are 1–10 m, for calcrete 0.1–10 m (with around 0.3–0.5 m being the most common), while for silcrete values of between 1 and 5 m appear normal.

Beneath the hardpan layer duricrusts display a variety of material types. Ferricretes, for example, often have rather erodible pallid and mottled horizons grading down into more or less coherent bedrock, while calcretes may be underlain by friable nodule horizons, and silcretes by kaolinitic clays. Related to the important geomorphological role of the differences between the properties of hardpans, sub-hardpan zones, weathered bedrock and bedrock in relatively simple profiles, is the role of alternations of different layers in complex profiles.

Another general aspect of duricrusts, which is relevant to their geomorphological impact, is the speed at which they form, and the rapidity with which they may harden on exposure. Rapid formation helps to preserve otherwise relatively ephemeral landforms (e.g. dunes or alluvial terraces). Quick rates of formation tend to be associated with duricrusts that originate through absolute accumulation rather than relative accumulation.

In spite of examples of rapid formation it is nonetheless apparent that for some of the great thicknesses of profiles to have developed, a considerable span of time (10^5–10^7 years) is required, together with a degree of land-surface stability. The Pleistocene was too short and too variable in climate for many of the great duricrust surfaces to have formed, and it may be for these reasons that so many of the world's duricrusts are of Tertiary age, or even earlier.

It is also important to realize that the geomorphological influence of duricrusts will depend to a considerable degree on the stage of evolution which the feature has reached. This affects both the overall thickness, the nature of the constituents, and the degree of induration.

Duricrusts may play a role in relief inversion (Plate 37). In the case of laterite, laterite-covered valleys may become ridges or strings of mesas flanking lower, younger valleys, and pediments may become mesas (McFarlane 1976). The relief of laterite surfaces may be modified by pseudokarstic processes so that the central areas of laterite-capped mesas may become gradually lowered. Thus the periphery stands relatively higher, giving a soup-plate form. Likewise, the tendency for some calcretes and dolocretes to form preferentially in valleys and depressions sometimes leads to inversion of relief in times of greater erosion, whether by water or wind. Examples of such inverted calcrete relief are provided by McLeod (1966).

Summerfield (1978) has also indicated that silcrete can cause relief inversion. In stage 1 of his model silcrete forms in areas subject to inundation and possibly reaches its thickest development in proximity to rivers. In stage 2 rejuvenation of drainage occurs leading to erosion and drainage inversion. Subsequent back-wearing (stage 3) creates silcrete-capped residuals. These may be highly resistant to further destruction by weathering since on a world basis silcretes have a mean silica content of around 96 per cent and may on occasion exceed 99 per cent. Silcrete residuals of Tertiary age are widespread in Europe and Britain, where they are known as sarsen stones.

The presence of duricrust profiles with marked differences in properties between hardpans and some of the more friable and fine-grained materials beneath, creates conditions that favour the formation of PSEUDOKARST produced by subsurface flushing, and in the case of calcrete, solutional effects. Cave formation and roof collapse produce karst-like forms in laterites. Calcretes, because of their high carbonate content and relative solubility, frequently show sinkhole development and pipe formation.

Many workers have used duricrusts as indicators of palaeoclimates, and in broad terms this may be acceptable. Calcretes, for example, are for the most part, though not exclusively, currently forming in semi-arid areas where annual rainfall is around 200–500 mm, so that their presence in various Tertiary sediments in western and central Europe may be used with a fair degree of certainty to infer formerly more arid conditions with an annual water deficit.

Plate 37 A laterite-capped plateau at Panchgani in the Deccan Plateau, India. The laterite acts as a caprock and has resulted from severe tropical weathering acting on Tertiary basalts

Much more controversy surrounds silcrete, however, as indicated by Summerfield (1983), with a range of inferred climatic conditions ranging from extreme arid to humid tropical. Summerfield maintains that silcrete may form under two distinct climatic regimes. He draws a distinction between 'the non-weathering profile' silcretes, which results from localized silica mobility and concentration in high pH environments under a predominantly arid and semi-arid climate, and 'the weathering profile' silcretes whose geochemical and petrographic characteristics are indicative of silicification under a much more humid climate in highly acidic, poorly drained weathering environments.

It is normally accepted that ferricretes and alcretes form under relatively humid conditions. Alternating wet and dry seasons were widely considered to be favourable if not essential to laterite genesis. In particular it was believed that seasonally alternating conditions were necessary for sesquioxide precipitation. However, as McFarlane (1976: 45) has pointed out, there is some evidence for its formation under permanently moist atmospheric conditions.

Duricrusts are widespread features, especially in low latitudes, though relict forms occur in more temperate ones. They have many geomorphological effects. Controversial is the question of their palaeoclimatic significance. They result from a complex interplay of different source materials, transfer processes and precipitation mechanisms in the surface and near-surface environment. In the past the roles of lateral translocations and organic processes have tended to be neglected.

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A.S. GOUDIE

DUST STORM

A large volume of predominantly silt-sized sediment blown into the atmosphere by a strong wind. The definition most widely used for this type of WIND EROSION OF SOIL event is that devised by meteorologists: a dust-raising event that reduces horizontal visibility to 1,000 m or less.

The entrainment of dust from the ground surface is controlled by the nature of the soil or sediment itself, the nature of the wind, and the presence of any surface obstacles to wind flow. Dust storms can occur in any environment given appropriate conditions of bare, unconsolidated sediment and a strong turbulent wind, but they are most common in deserts and on their margins. Most geomorphologists define dust particles according to the silt/sand boundary (i.e. less than 62.5 μm). The particles that make up desert dust storms are dominated by SiO_2, probably reflecting the importance of quartz in source areas. Grain size, mineralogy and chemical composition can be used to distinguish soil dust from other types of particles in the atmosphere such as those derived from sea salt, volcanoes and smoke particles from fire.

Different terrain types vary greatly in their susceptibility to dust storm occurrence. Important determining factors include the ratio of clay-, silt- and sand-sized particles, the soil moisture content, the compaction of sediments, the presence of particle cements such as salts or organic breakdown products, and the presence of crusts or armoured surfaces (Middleton 1990). The most favourable dust-producing surfaces are areas of bare, loose and mobile sediments containing substantial amounts of sand and silt but little clay. Terrains that satisfy these conditions are most commonly found in geomorphologically active landscapes where tectonic movements, climatic changes and/or human disturbance are responsible for rapid exposure, incision and reworking of sediment formations containing dust. Important sources of dust storms are generally located in specific, relatively small desert environments (Coudé-Gaussen 1984) such as floodplains, alluvial fans, salt pans and devegetated fossil dunes.

The most important meteorological systems capable of generating dust storms are synoptic in scale, dominated by the passage of low pressure fronts with intense baroclinal gradients that are accompanied by very high-velocity winds. Such frontal passage is the dominant dust-generating mechanism in many of the world's dusty regions, including Australia, northern China and Mongolia, Central Asia, the Levant, the Mediterranean coast of north Africa, the Sahelian latitudes of west Africa, the High Plains of the USA and the plains of the Argentine Pampas. Surface cyclones themselves may sweep out gyres of dust when circulation around the low pressure becomes very intense. Other synoptic-scale dust-raising systems include winds generated in areas with steep pressure gradients, such as in the Thar Desert of India and Pakistan and the northwesterly Shamal wind that blows down the Arabian Gulf from Iraq and Kuwait. More localized dust storms occur when katabatic winds deflate mountain foot sediments, as on the northern slopes of Kopet Dag on the Iran-Turkmenistan border, or in California (the Santa Ana wind). The high Andean Altiplano of Chile, north-west Argentina and southern Bolivia experiences strong dust-raising from the upper westerlies and similar upper airflow deflates sediments from the arid Tibetan Plateau. The cold downburst wind of a dry thunderstorm, the classic Haboob of southern Sudan, is perhaps the most common meso-scale dust-raising system, which raises dust at the gust front some kilometres in advance of the towering convective clouds.

Dust transport and deposition

Globally, the amount of material mobilized in dust storms is thought to be around a billion tons a year and up to half of this comes from the Sahara, indicating the geomorphological importance of AEOLIAN PROCESSES in moulding parts of its landscape. The world's two most active dust storm source areas are both in the Sahara: the Bodélé Depression to the south of the Tibesti Mountains and an area covering eastern Mauritania, western Mali and southern Algeria (Goudie and Middleton 2001). Their importance relative to other major global dust sources is indicated in Table 15 which shows maximum mean values of an Aerosol Index (AI) that indicates the intensity of atmospheric dust content. The AI is derived from the satellite-borne Total Ozone Mapping Spectrometer (TOMS) that detects UV-absorbing aerosols in the atmosphere.

Sediments from these and other world dust storm areas are regularly transported over great distances. Saharan dust is transported along three main trajectories: westward over the North Atlantic to North and South America; northward across the Mediterranean to southern Europe and sometimes as far north as Scandinavia, and along easterly trajectories across the eastern Mediterranean to the Middle East. Dust storm material from other major deserts also follows common trajectories, many of which are highly seasonal (Plate 38). They include flows from north-east Asia across the Pacific Ocean and from Mesopotamia down the Arabian Gulf. Dense dust loadings following such trajectories can be discerned on imagery from remote sensing platforms and many techniques have been applied to deposited material in order to detect its source. These include dust mineralogy and elemental composition, scanning electron microscopy of individual grain features and the presence of pollen and foraminifera.

While in the troposphere, dust can have effects on climate through a range of possible influences. Dust outbreaks may affect air temperatures through the absorption and scattering of solar radiation and may cause ocean cooling. Dust-induced changes in atmospheric temperatures and changes in concentrations of potential condensation nuclei may also affect convectional activity and cloud formation, thereby altering rainfall amounts.

Dust aerosols influence the nutrient dynamics and biogeochemical cycling of both marine and terrestrial ecosystems. Much of the material transported over long distances is deposited over the oceans (Prospero 1996) where dust storm sediments provide a major nutrient input. Where deposited on land, dust may affect soil formation. Dust that has a high carbonate content may be a factor in the formation of calcretes and dust

Table 15 Maximum mean AI values for major global dust sources determined from TOMS

Location	
Bodélé Depression of central Sahara	>30
West Sahara in Mali and Mauritania	>24
Arabia (southern Oman/Saudi border)	>21
Eastern Sahara (Libya)	>15
South-west Asia (Makran coast)	>12
Taklamakan/Tarim basin	>11
Etosha Pan (Namibia)	>11
Lake Eyre basin	>11
Mkgadikgadi basin (Botswana)	>8
Salar de Uyuni (Bolivia)	>7
Great Basin of the USA	>5

Source: Goudie and Middleton (2001)

Plate 38 A dust raising event at Disi, south-east Jordan, with dust being raised from a dry playa surface

contributes to the formation of other desert surface coverings such as desert varnish and case hardening of rocks. Salts carried in wind-blown dusts can act as weathering agents and increase the salinity of soils and water bodies.

LOESS is by definition a wind-deposited dust with a median grain size range of 20–30 μm (Tsoar and Pye 1987) and has been estimated to cover up to 10 per cent of the world's land area. Interestingly, however, the occurrence of loess in Africa is very limited, a fact that is surprising given the Sahara's prominence as the world's largest area of contemporary dust storm activity and evidence that suggests it produced more dust during cold phases of the Pleistocene (see below).

Changing frequencies of dust storms

There is considerable evidence that dust storm frequencies can change substantially in response to climatic changes both in the long term (e.g. during the Last Glacial Maximum) and in the short term (e.g. in response to the North Atlantic Oscillation and to drought phases). Analysis of dust in cores taken from deep-sea sediments, ice caps and loess deposits has enabled the reconstruction of long-term changes in dust storm activity. Dust in North Atlantic sediments has been dated back to the early Cretaceous, although aeolian activity in the Sahara appears to have become more active in the late Tertiary and high dust loadings were a particular feature of the Pleistocene in many parts of the world.

Intensification of dust storm activity during glacial periods, such as the Last Glacial Maximum, was probably due in part to lower precipitation, although changes in wind regimes may also have contributed. It has also been suggested that increased atmospheric dust during the Last Glacial Maximum was not only a response to climate change but also a contributory factor to the change, and regional dust loadings are being built into models of climate (Mahowald *et al.* 1999).

Drought is commonly associated with an increase in dust-raising activity as vegetation cover dies off and soils dry out (Brooks and Legrand 2000). In the more recent past, the effects of drought on dust storm activity have sometimes been exacerbated by human influences on the wind erosion system. Human activity has been shown to affect dust storm activity by destabilizing soil surfaces and altering vegetation cover. The most common human impact in this respect is agriculture. Type examples of large areas in semi-arid climates converted from grasslands to cultivation subsequently becoming enhanced dust-producing regions include the Great Plains of the USA in the 1930s, the so-called Dust Bowl, and the Virgin Lands scheme of the former Soviet Union in the 1950s. Other human activities that affect changes in dust storm frequencies through the breakup of wind-resistant surfaces and/or the removal of protective vegetation cover from soils include drainage, construction, vehicle use and military movements. The relationship between dust-raising, environmental change and human impacts has meant that changes in dust storm frequency have been studied as an indicator of DESERTIFICATION.

Dust storm hazards

Airborne dust presents a variety of problems to inhabitants of desert areas. In areas where dust is raised from agricultural fields it represents a serious form of wind erosion while blowing dust and sand can cause considerable damage to crops and natural vegetation by abrasion, which is particularly critical for young shoots when fields are poorly protected by vegetation cover. The reduction in visibility caused by dust storms is a serious hazard to aviation and road transport.

Dust storms are a form of atmospheric pollution and may transmit diseases that affect plants, animals and humans. Fungus carried in Saharan dust has been implicated in disease outbreaks in coral reefs throughout the Caribbean (Smith *et al.* 1996). Micro-organisms blown in dust may settle on the skin, be swallowed or inhaled into respiratory passages. In Arizona, Valley Fever is caused by *Coccidioides immitis*, a common airborne fungus blown by dust storms. Inhalation of fine particles can also aggravate diseases such as asthma, bronchitis and emphysema. These risks to human and ecosystem health have been noted both in dust source areas and in areas of deposition after long-range transport (Griffin *et al.* 2001).

Applied geomorphologists have played a useful role in identifying dust sources and methods for preventing dust entrainment in arid zones. Jones *et al.* (1986) have suggested a general procedure for the assessment of dust hazards in urban areas after their work in the Middle East.

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NICHOLAS MIDDLETON

DYE TRACING

A large variety of dyes and related compounds have been used in water tracing and these are generally classified into non-fluorescent and fluorescent dyes, although strictly some of the substances commonly included in the latter category are not dyes but dye intermediaries or optical brightening agents (OBAs). The earliest scientific water-tracing experiments used simple colouring agents which were detected visually. Later it was discovered that some dyes, such as Rhodamine, may be detected on treated cotton hanks. These visual dyes have no advantages and several disadvantages. Visual detection requires high concentrations and observers and of the two dyes which can be absorbed onto cotton hanks, Malachite Green and Rhodamine B, the latter has been shown to be toxic. Hence, they cannot be recommended and are no longer used.

Fluorescent dyes differ from simple colouring agents in that when irradiated at a particular wavelength they emit light at a different, longer, wavelength. Hence, they can be detected in water samples in concentrations invisible to the naked eye; theoretically down to $\mathrm{ng\,l^{-1}}$. They have become the most widely used water-tracing substances in limestone areas and have also been used successfully to trace water flow through other fractured rocks, through soils and through peat pipes. There are many fluorescent dyes and they are generally divided into three groups on the basis of their fluorescence spectra: Blue (e.g. Amino G Acid and Optical Brighteners such as Leucophor BS and Tinopal CBS-X and ABP); Green (e.g. Fluorescein (and its disodium salt Uranine), Pyranine and Lissamine) and Orange (e.g. Rhodamine WT, Sulpho-Rhodamine B and Eosine).

Field determination of the green and orange dyes is possible using a portable fluorimeter and in the laboratory a modern scanning spectrofluorimeter can distinguish between different dyes allowing multiple tracing experiments to be undertaken. A further advantage is that several of the green and orange dyes are absorbed by charcoal grains and may be released in the laboratory by an alkaline-alcohol elutant. This is particularly useful if it is necessary to monitor several sites as charcoal bags (variously known as fluocapteurs or receptors) can be deployed at each site and left for up to a week to scavenge dye. Unfortunately the charcoal also scavenges other organic substances which can make interpretation difficult. Treated cotton detectors can be used as fluocapteurs for optical brightening agents.

There are also disadvantages associated with each individual dye. Blue dyes, and especially OBAs suffer from high and very variable background and break down in sunlight; the green dyes fluoresce in the same area as certain organisms and organic substances; certain reds are potentially carcinogenic; and green, and to a lesser extent red, dyes may be lost on sediment.

Further reading

http://www.dyetracing.com. This website provides useful information on dye tracing.
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JOHN GUNN

DYKE (DIKE) SWARM

A dyke (dike) is a tabular igneous body that was sub-vertical at the time of emplacement.[1] A dyke swarm is a set of coeval dykes which typically display a linear, radiating or arcuate geometry. Dyke swarm compositions range from ultramafic to felsic. The largest swarms are of basaltic composition (with diabasic texture) and are most prominent in basement terranes. Individual diabase dykes can range in width from centimetres to hundreds of metres and in length from metres to 2,000 km or more.

Radiating swarms with radii of about 100 km are associated with individual volcanic centres. Radiating swarms with radii >300 km (referred to as giant radiating dyke swarms) form the feeder systems for large igneous provinces, and are thought to focus above the centres of mantle plumes (Ernst and Buchan 2001: chapters 12 and 19). Magma can be transported both vertically and laterally in dykes. In particular, giant radiating swarms can transport magma laterally more than 2,000 km from the plume centre and can feed sills and volcanic rocks at any distance along their extent. A classic example of a giant radiating swarm is the 1267 Ma Mackenzie swarm of the northern Canadian Shield which fans over an arc of 90° and extends 2,300 km from the focal point (Fahrig 1987). There are also numerous giant radiating swarms on Venus and Mars (Grosfils and Head 1994; Ernst *et al*. 2001).

Many giant linear swarms on Earth may be fragments of giant radiating swarms, which have been dismembered during episodes of continental breakup. However, linear swarms can also be associated with spreading ridges, ophiolite complexes and rift zones. Arcuate portions of otherwise linear or radiating swarms may reflect primary geometry (i.e. changes in the regional stress field) or later deformation. In addition, some arcuate swams occur as a set of ring dykes generated above an intrusion.

In addition to their magmatic significance, dykes can also act as a barrier to groundwater flow, and may localize hydrothermal fluids along their margins. They often weather positively as linear ridges or negatively as troughs due to differential erosion.

Note

1 This differs from the traditional definition of dyke which would allow an originally horizontal sheet to be termed a dyke if it is discordant to the bedding or foliation of its host rocks. Because vertical and horizontal sheets imply fundamentally different stress conditions, originally sub-vertical sheets should be termed dykes and originally sub-horizontal sheets should be referred to as sills, regardless of the degree of discordance. Tabular bodies for which the original orientation cannot be determined or where the dip is intermediate would be termed 'sheets'.

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DYNAMIC EQUILIBRIUM

This is a concept which describes a situation of relatively restricted fluctuations about a mean value, together with non-stationarity of that mean. It has been propounded, especially by the American geomorphologist S.A. Schumm (1977, 1991) as a device which permits the short-term variability of 'GRADED TIME' to be viewed alongside the longer trajectory of 'CYCLIC TIME' (most especially for changes in the relief of valleys). In this sense, Schumm, in particular, has sought to bring together the apparently conflicting timescales and explanatory modes which late twentieth-century geomorphologists have often considered to be favoured by G.K. Gilbert and W.M. Davis respectively. The need to reconcile such apparent conflict may be linked to the extremely influential paper by R.J. Chorley (1960) that advocated the shift from an historical emphasis on the long-term and largely predictable progressive change in landforms represented by the Davisian cycle, towards a focus upon the dynamic and process-oriented approach associated with Gilbert and considered to be strongly encouraged by the adoption of systems thinking (see SYSTEMS IN GEOMORPHOLOGY).

At the root of the use of the concept of dynamic equilibrium in late twentieth-century geomorphology is, undoubtedly, Gilbert's 1877 development of the concept of grade (see GRADE, CONCEPT OF), which was then – interestingly – taken up by Davis who incorporated it within his cyclic models (see Chorley *et al.* 1964; Chorley *et al.* 1973). Grade itself has proved one of the thorniest and most elusive geomorphological concepts (cf. Kesseli 1941) and one which, by the year 2000, had more or less vanished from the literature.

Although ideas of equilibria at a variety of timescales and with a variety of theoretical underpinnings characterized much of Anglo-American geomorphology in the latter half of the twentieth century, it became increasingly clear that the concepts work best in very closely defined circumstances, where the physics or chemistry of the situation is relatively unobscured by historically contingent variability which is poorly susceptible to mathematical treatment (cf. Selby's 'strength EQUILIBRIUM SLOPES'). The whole notion of geomorphological equilibria was extensively reviewed by Thorn and Welford (1994; and discussion) and it must remain a matter of opinion whether one accepts their conclusion that the concepts are of central, significant and continuing explanatory value to geomorphology as a whole.

Certainly Schumm's belief in the utility of the notion of dynamic equilibrium has been persistent and pervasive. However, in the light of his equal emphasis on the role of geomorphic thresholds (see THRESHOLD, GEOMORPHIC), it might be argued that a better concept to link the emphases on process and on evolution is his 'Model 2' (Schumm 1977: 12) in which the dynamic equilibrium of cyclic time is replaced, in the same time frame, by dynamic *metastable* equilibrium. Whereas Schumm's dynamic equilibrium (*sensu stricto*) couples a general reduction in relief (that is, a non-stationary mean elevation) with oscillating episodes of cut-and-fill; his dynamic metastable equilibrium assumes long periods of effective stationarity of mean elevation, with variable erosion and aggradation interrupted by abrupt, episodic erosion. This condition would certainly better describe the kind of situation observed in Piceance Creek, Colorado (Schumm 1977: 78–81) where gullying was shown to be episodic. It would also accord more neatly with developing ideas of complex, non-linear models.

It remains the case, however, that both forms of dynamic equilibrium are difficult to identify unambiguously and, further, that neither may be said to add true clarity to our understanding of geomorphic process and form. One central problem is the uncertainty of the temporal and spatial scales at which the 'graded' gives way to the 'cyclic'. Such basic questions of the identification of the crucial timescales and spatial scales of landscape development and the inherent problems of piecing together events on different scales (cf. Schumm and Lichty 1965) remain central (see Church 1996). But whether the concepts of dynamic or dynamic metastable equilibrium are what is needed as the framework to reconcile the process study with the evolutionary one, is far from certain.

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SEE ALSO: complex response; complexity in geomorphology; cycle of erosion; denudation chronology; equilibrium shoreline; geomorphic evolution; punctuated aggradation

BARBARA A. KENNEDY

DYNAMIC GEOMORPHOLOGY

Dynamic geomorphology is defined as an emphasis in geomorphology which treats geomorphic processes as 'gravitational or molecular shear stresses, acting on elastic, plastic or fluid earth materials to produce the characteristic varieties of strain or failure which we recognize as the processes of weathering, erosion, transportation and deposition' (Strahler 1952). This emphasis was first thoroughly exemplified in the work of G.K. Gilbert (1877, 1909, 1914, 1917) and emulated by Brigadier Bagnold (1941) but was largely overlooked by geomorphologists until Horton (1945), Strahler (1952) and Tricart's initiation of the *Revue de Géomorphologie Dynamique* (1950). It is no exaggeration to say that this emphasis is today dominant in Anglo-American and Japanese geomorphology and is often equated with process geomorphology.

The work of G.K. Gilbert is the first seminal antecedent to the study of geomorphic process or dynamic geomorphology. His report on the geology of the Henry Mountains (1877) described the physical erosional processes and derived a system of laws governing progress from initial to adjusted forms; his discussion of the convexity of hill-tops introduced the role of soil creep as a dynamic process; his discussion of transportation of debris by running water was based on flume experimental data; and his paper on hydraulic mining debris in the Sierra Nevada was path-breaking in its recognition of the effects of the passage of a slug of sediment passing through the Sacramento River system over a period of sixty years, from the time of the commencement of gold mining to the time of his publication in 1917. Gilbert, apparently single-handedly, established the paradigm of dynamic or process geomorphology.

Brigadier Bagnold published his monumental *Physics of Blown Sand and Desert Dunes* in 1941. This book remains a sourcebook for students of AEOLIAN PROCESSES. But Bagnold also contributed to the fundamental understanding of beach formation processes and fluvial processes, much of this understanding deriving from his home made 2-m flume which he maintained in his home grounds. The third antecedent of contemporary dynamic geomorphology was the Uppsala school of physical geography in Sweden, where Hjulstrom (1935), Sundborg (1956) and Rapp (1960) transformed our understanding of fluvial process and drainage basin geomorphology. A fourth antecedent is surely the French collaboration of Tricart and Cailleux and the significant influence of Tricart in maintaining the momentum of a journal dedicated to dynamic geomorphology. Tricart's department of applied geomorphology at the University of Strasbourg was unique in continental Europe in its pioneering of this emphasis.

The essential vision of what constituted dynamic geomorphology was developed by Strahler (1952). Shear stresses affecting Earth materials were divided into two major categories: gravitational and molecular. Gravitational stresses activate all downslope movements of matter, hence include all mass movements, all fluvial and glacial processes. Indirect gravitational stresses activate tide- and wave-induced currents and winds. Phenomena of gravitational shear stresses were subdivided according to behaviour of rock, soil, ice, water and air as elastic or plastic solids and viscous fluids.

Molecular stresses are those induced by temperature changes, crystallization and melting, absorption and desiccation, or osmosis. These stresses act in random or unrelated directions with respect to gravity. Surficial creep results from a combination of gravitational and molecular stresses on a slope. Chemical processes of solution and acid reaction were considered separately. Strahler went on to say that a fully dynamic approach requires analysis of geomorphic processes in terms of open systems which tend to achieve steady states of operation and are self-regulatory to a large degree. Finally, he specified that formulation of mathematical models, both by rational deduction and empirical analysis of observational data, to relate energy, mass and time was the ultimate goal of the dynamic approach.

Strahler's motivation was to counteract the heavy emphasis on descriptive, deductive studies of landform development and regional geomorphologies that had come to dominate the subject in the early part of the twentieth century. This dynamic emphasis in geomorphology has also been characterized as functional geomorphology (Chorley 1978) and is contrasted with historical geomorphology.

Strahler (1992: 72–73) describes his encounter with open systems theory (Von Bertalanffy 1950) in the following way: 'It was as if a closed door had opened before me, revealing an entirely new and powerful epistemology of science – a paradigm capable of unifying all dynamic processes and forms that can be observed in the universe.' Strahler (1980) describes the five levels of systems organization which compose his mature reflections on dynamic geomorphology. Level 1, which corresponds closely with his 1952 discussion, concerns the collection of data which are considered potentially useful in understanding the geomorphic system. The data must be quantitative and must be in the fundamental dimensions of mass, length, time, temperature and their products. The system variables are grouped into (a) dynamic, (b) mass-flow, (c) geometry and (d) material property variables. The dynamic variables relate to energy, force and stress; the mass-flow variables express rates of flow of matter; the geometry variables describe size and form, and material property variables include environmental constants and regulators. The second level of analysis relates to morphological elements; the third level examines flow systems of interconnected pathways of transport of energy and matter; the fourth level describes process-form systems, characterized by self-regulation through physical feedback loops; and the fifth level, systems regulated by cybernetic feedback, links natural systems with those regulated and/or disturbed by human intervention. The agenda described is reminiscent of the Chorley and Kennedy (1971) agenda and underlines the close relation between dynamic geomorphology and the general systems framework.

Following Strahler's most important impetus to the development of dynamic geomorphology, the contributions of Schumm, Leopold, Wolman, Gregory and Walling can be seen as setting the seal on the paradigm, especially in the context of fluvial and watershed geomorphology. John Miller, the third of the triumvirate of Leopold and Miller, would surely have had a major influence, perhaps even the greatest influence, had he not died tragically at the age of 39. The reason for such a bold suggestion is that John Miller alone among this group of leaders was a geochemist as well as a geomorphologist and he was attempting to link weathering processes as well as mechanical erosional processes to drainage basin evolution. The themes of dynamic equilibrium, magnitude and frequency (see MAGNITUDE–FREQUENCY CONCEPT) of operation of geomorphic processes, and a strong bias towards fluvial process were to become the hallmark of dynamic geomorphology in the Anglo-American literature. The paradigm became hugely popular partly because of its quantitative rigour, partly because it seemed to provide specific answers in a field where thoughtful arm-waving had become a tradition and partly because it recognized the value of the combination of theory, experiment and practice.

Dynamic geomorphology has now become equated with 'process geomorphology'. A current textbook on process geomorphology notes that valid interpretations of geomorphic history must be based on a thorough understanding of the processes involved in landform development. Geomorphologists therefore must be cognizant of process mechanics prior to analysing how landform history manifests past climatic or tectonic phenomena (Ritter *et al*. 1995). Five basic principles of process geomorphology according to Ritter *et al*. are: (1) a delicate balance or equilibrium exists between landforms and processes. The character of this balance is revealed by considering both landforms and processes as systems or parts of systems; (2) the perceived balance between process and form is created by the interaction of energy, force and resistance; (3) changes in driving force and/or resistance may stress the system beyond the defined limits of stability. When these limits of equilibrium or thresholds are exceeded, the system is temporarily in disequilibrium and a major response may occur. The system will develop a different equilibrium condition adjusted to the new force or resistance controls, but it may establish the new balance in a complex manner; (4) various processes are linked in such a way that the effect of one process may initiate the action of another; and (5) geomorphic analyses can be made over a variety of time intervals. In process studies the time framework utilized has a direct bearing on what conclusions can be made regarding the relation between process and form. Therefore the time framework should be determined by what type of geomorphic analysis is desired.

These principles are clearly related to the earlier dynamic geomorphology of Gilbert, Bagnold, Hjulstrom, Sundborg, Rapp, Cailleux, Tricart and Strahler through the ideas of 'a balanced condition', the centrality of energy, force and resistance, the language of systems theory, complex response and the importance of timescale of study. One implication of the reductionist functionalism of dynamic geomorphology was the emergence of semi-independent geomorphic process schools, such that coastal, slope, glacial, periglacial, karst, aeolian and fluvial geomorphology became more formally differentiated. The demands of learning the mechanics and dynamics of process led to an isolation of the new dynamic geomorphology from those who were more interested in the evolution of landscapes over geological time. The central conundrum was articulated by Church (1980) when he commented that contemporary records of geomorphological processes were not likely to represent long-term behaviour sufficiently well to provide any firm basis for understanding landscape evolution.

Schumm and Lichty (1965) made a significant contribution to the linking of short-term and long-term studies by explicitly recognizing the different status of process variables over cyclic, graded and steady-state timescales. Steady-state timescales were appropriate for process studies; cyclic timescales would be appropriate to geological evolutionary studies and the graded timescale would be appropriate to, perhaps, the Holocene timescale. They suggested a reconciliation between timeless and time-bound aspects of geomorphology by noting that the distinction between cause and effect among geomorphic variables varies with the size of the landform/landscape under consideration as well as with time. Theirs is an interesting and valuable insight, but the problem would seem to arise from an insistence on the idea of balance or equilibrium at the core of dynamic geomorphology. In spite of the power of the dynamic geomorphology paradigm, there remains a tension between the way in which time and space scales of variability are treated and the central assumption of balance and equilibrium. The question at issue is whether the landscape is fundamentally in equilibrium or whether it is fundamentally in a transient state between equilibria which are rarely if ever achieved.

Dynamic geomorphology has revived geomorphology from its pre-Second World War slumber, has connected geomorphology with the other natural sciences of physics, chemistry and biology, and has opened up opportunities for professional accreditation alongside engineers. At the same time, it can be suggested that process geomorphology has, at least for a few decades, lost touch with both traditional geology and geography. With geology in that the diastrophic framework supplied by global plate tectonics has been difficult to marry with local-scale process studies; with geography in that the challenges of global environmental change and the role of human society have been equally difficult to marry with site-scale process studies.

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SEE ALSO: cyclic time; dynamic equilibrium; force and resistance concept; graded time

OLAV SLAYMAKER

E

EFFECTIVE STRESS

The difference between total stress (σ) and pore pressure in a material (u), responsible for mobilizing internal friction. Effective stress (σ') is one of the two components of internal stress within a material, alongside pore pressure, and measures the distribution of load carried by the soil over a specific area. The principle of effective stress was developed by Karl Terzaghi between 1923–1936, and is a fundamental theory in soil mechanics. Changes in stress, such as distortion, compression and shearing resistance changes, are due to variations in effective stress. As effective stress values increase, the soil or rock becomes more consolidated, exhibiting a maximum value at complete consolidation and before shear failure. Thus, it is the effective stress that causes important changes in material strength, volume and shape. Long-term SLOPE STABILITY analysis often incorporates effective stress analysis (inclusive of internal stresses), rather than total stress analysis (short-term slope instability).

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STEVE WARD

EL NIÑO EFFECTS

Climate oscillations occur at many timescales. For example, in the tropics a sub-annual, intra-seasonal 40–60-day period Madden–Julian oscillation has been identified. At a slightly longer timescale there is a 2 to 2.5 year oscillation in the equatorial jet in the lower stratosphere, and this is called the Quasi-Biennial Oscillation (QBO). Every three to seven years a two or so year-long event occurs which is called the El Niño Southern Oscillation (ENSO). Decadal and interdecadal variability is evident in the North Atlantic Oscillation (NAO). At even longer timescales there are such important phenomena as the Dansgaard–Oeschger Cycles, Bond Cycles and Heinrich events, which occur at century to millennial scales. El Niño is the term used to describe an extensive warming of the upper ocean in the tropical eastern Pacific lasting up to a year or even more. The negative or cooling phase of El Niño is called *La Niña*. El Niño events are linked with a change in atmospheric pressure known as the Southern Oscillation (SO). Because the SO and El Niño are so closely linked, they are often known collectively as the El Niño/Southern Oscillation or ENSO. The system oscillates between warm to neutral (or cold) conditions every three to four years.

Precipitation and temperature anomalies appear to characterize all El Niño warm episodes. These include:

- The eastward shift of thunderstorm activity from Indonesia to the central Pacific usually results in abnormally dry conditions over north Australia, Indonesia and the Philippines.
- Drier-than-normal conditions are also usually observed over south-eastern Africa and north Brazil.
- During the northern summer season, the Indian monsoon rainfall tends to be less than normal, especially in the north-west of the subcontinent.

- Wetter-than-normal conditions are usual along the west coast of tropical south America, and at subtropical latitudes of North America (the Gulf Coast) and South America (south Brazil to central Argentina).
- El Niño conditions are thought to suppress the development of tropical storms and hurricanes in the Atlantic but to increase the numbers of tropical storms over the eastern and central Pacific Ocean.

In the twentieth century there were around twenty-five warm events of differing strengths, with that of 1997/8 being seen as especially strong. ENSO was relatively quiescent from the 1920s to 1940s.

Severe El Niños, like that of 1997/8, can have a dramatic effect on rainfall amounts. This was shown with particular clarity in the context of Peru (Bendix *et al.* 2000), where normally dry locations suffered huge storms. At Paita (mean annual rainfall 15 mm) there were 1,845 mm of rainfall while at Chulucanas (mean annual rainfall 310 mm) there were 3,803 mm. Major floods resulted (Magilligan and Goldstein 2001).

The Holocene history of El Niño has been a matter of some controversy (Wells and Noller 1999) but Grosjean *et al.* (1997) have discovered more than thirty debris flow events caused by heavy rainfall events between 6.1 and 3.10 Kyr BP in the northern Atacama Desert. The stratigraphy of debris flows has also been examined by Rodbell *et al.* (1999), who have been able to reconstruct their activity over the last 15 Kyr. Between 15 and 7 Kyr BP, the periodicity of deposition was equal to or greater than 15 years and then progressively increased to a periodicity of 2 to 8.5 years. The modern periodicity of El Niño may have been established about 5 Kyr BP, possibly in response to orbitally driven changes in solar radiation (Liu *et al.* 2000). Going back still further, studies of the geochemistry of dated *Porites* corals from the last interglacial of Indonesia have shown that at that time there was an ENSO signal with frequencies nearly identical to the instrumental record from 1856–1976 (Hughen *et al.* 1999).

El Niño events have considerable geomorphological significance. For example, the changes in temperatures of sea water between El Niño and La Niña years have a clear significance for coral reefs (Spencer *et al.* 2000). In 1998, sea surface temperatures in the tropical Indian Ocean were as much as 3–5 °C above normal, and this led to up to 90 per cent coral mortality in shallow areas (Reaser *et al.* 2000). Although other factors may be implicated in coral mortality (e.g. eutrophication, disease, heavy fishing, etc.) large changes in the health of reefs have been noted from remote islands and reefs with low levels of human influence. It would seem that warm conditions between 25 and 29 °C are good for coral growth, but that temperatures above 30 °C are deleterious (McClanahan 2000) and lead to such phenomena as coral bleaching. Bleaching tends to be greatest at shallow depths but a feature of the 1998 event was that it not only caused bleaching of rapidly growing species, but also affected massive species. It also reached to depths as great as 50 m in the Maldives. If global sea temperatures rise as a result of global warming and become closer to the thermal tolerance level of 30 °C, El Niño events of smaller magnitude will be sufficient to cause bleaching. Moreover, the closer the mean sea temperature is to this thermal limit, the longer will be the period for which the tolerances of corals will be exceeded during any El Niño, thereby increasing the likelihood of coral mortality (Souter *et al.* 2000).

Lake levels also respond to El Niño events. El Niño warming in 1997 led to increased rainfall over East Africa that caused Lake Victoria to rise by 1.7 m and Lake Turkana by *c.*2 m (Birkett *et al.* 1999). The abrupt rise in the level of the Caspian Sea (2.5 m between 1978 and 1995) has also been attributed to ENSO phenomena (Arpe *et al.* 2000). Similarly, the 3.7-m rise in the level of the Great Salt Lake (Utah, USA) between 1982 and 1986 was at least partly related to the record rainfall and snowfall in its catchment during the 1982/3 El Niño (Arnow and Stephens 1990). The enormous changes that occur in the areal extent of Lake Eyre in Australia result from ENSO-related changes in inflow, with the greatest flooding occurring during La Niña phases (Kotwicki and Allan 1998).

Some glacier fluctuations are controlled in part by El Niño. Glacier retreat in the tropical Andes can be attributed to increased ablation during the warm phases of ENSO (Francou *et al.* 2000). Conversely, further south, in the southern Andean Patagonia of Argentina, El Niño events have led to increased snow accumulation, causing glaciers to advance so that they create barriers across drainage, creating glacier-dammed lakes (Depetris and Pasquini 2000).

El Niño impacts upon tropical cyclone activity, and the differences in cyclone frequency between El Niño and La Niña years is considerable (Bove *et al.* 1998). For example, the probabilities of at least two hurricanes striking the US is 28 per cent during El Niño years, 48 per cent during neutral years and 66 per cent during La Niña years. There can be very large differences in hurricane landfalls from decade to decade. In Florida, over the period 1851–1996, the number of hurricane landfalls ranged from 3 per decade (1860s, and 1980s) to 17 per decade (1940s) (Elsner and Kara 1999). Given the importance of hurricanes for slope, channel and coastal processes, changes of this type of magnitude have considerable geomorphological significance. Mangroves, for example, are highly susceptible to hurricanes, being damaged by high winds and surges (Doyle and Girod 1997).

Streamflows and sediment yields may also be affected by El Niño events. One area where there have been many investigations of the links between ENSO and streamflow is in the western United States. There is a tendency for the southwest to be wet and the north-east to be dry during the El Niño warm phases (Negative Southern Oscillation Index), and vice versa for La Niña (Cayan *et al.* 1999). There is some evidence that the effect on streamflow is amplified over that on precipitation. A study of sediment yields in southern California showed that during strong El Niño years severe storms and extensive runoff occurred, producing sediment fluxes that exceeded those of dry years by a factor of about five. The abrupt transition from a dry climate to a wet climate in 1969 brought a suspended sediment flux in the rivers of the Transverse Range of 100 million tons, an amount greater than their total flux during the preceding 25-year period (Inman and Jenkins 1999). The wet period from 1978–1983 caused a significant response on alluvial fans and in channels in desert piedmont areas (Kochel *et al.* 1997).

Phases of high sediment yield may themselves have geomorphological consequences. It has been argued, for example, that Holocene beach ridge sequences along the north coast of Peru may record El Niño events that have occurred over the last few thousands of years. The argument (Ortlieb and Machare 1993) is that heavy rainfall causes exceptional runoff and sediment supply to coastal rivers. This, combined with rough sea conditions and elevated sea levels, is potentially favourable for the formation of beach ridge sequences. The high sea levels caused by El Niño, often amounting to 20–30 cm, can contribute to washover of coastal barriers (Morton *et al.* 2000).

Heavy rainfall events associated with ENSO phenomena can cause slope instability. Some of the most distinctive landslides in the south-west of the USA have occurred during El Niño events, and they can be especially serious if the heavy rainfall events occur on slopes that have been subjected to fires associated with previous drought episodes (Swetnam and Betancourt 1990).

On the other hand, exceedingly wet years can in due course cause a great increase in vegetation cover on slopes that may persist for some years and so lead to more stable conditions. In the arid islands of the Gulf of California, for example, plant cover ranges from 0–5 per cent during ‘normal’ years, but during rainy El Niño periods it rises to 54–89 per cent of the surface available for growth (Holmgren *et al.* 2001). Wet ENSO events can provide rare windows of opportunity for the recruitment of trees and shrubs. Such woodland can be resilient and, once established, can persist.

ENSO can be associated with intensified drought conditions and so can influence the activity of aeolian processes, particularly in areas which are at a threshold for dust entrainment or dune activation. Such areas will be those where in wet years there is just enough vegetation to stabilize ground surfaces. In the USA, dust emissions were greatly reduced in the period 1983–84 following the heavy rainfall of the 1982 El Niño (Lancaster 1997). Likewise, Forman *et al.* (2001) have reconstructed the history of dune movements in the Holocene in the USA Great Plains. They have found that phases of dune activity have been associated with a La Niña-dominated climate state and weakened cyclogenesis over central North America.

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A.S. GOUDIE

ELUVIUM AND ELUVIATION

For soils which exist in areas where the water balance is such that rainfall is greater than evaporation, the excess water drains downwards under the influence of gravity. This percolating water can carry material in solution, a process known as leaching. Additionally, material in the form of very fine particles can be moved down in suspension and this is referred to as eluviation – or a 'washing' of particles out of an upper soil horizon, the material being referred to as eluvium. Eluviation can be referred to as mechanical eluviation to distinguish it from losses occuring in solution. When the material becomes redeposited further down the soil profile it is referred to as illuvium, or that material which is washed in to a new lower location by illuviation. The process overall results in the upper soil horizons having a coarser texture, and therefore a greater porosity and permeability, and can result in a finer textured, and sometimes compacted, layer below.

In the lower soil profile, the redeposition of the eluvial material takes place within voids or on the walls of channels, forming a coating of clay round coarser particles in a skin of material where the clay particles are often oriented parallel to each other round the large particle. Such a clay skin is referred to as a cutan, derived from the Latin *cutis*, meaning a skin, coating or rind (cf. cuticle) (Brewer 1964). Such a deposition contributes to the decease of soil pore size and can thus impede further drainage.

Alternatively the eluvial material can be washed out of the soil profile in downslope moving waters such as throughflow or return overland flow. Here the eluvium may be redeposited in or on the soil at the slope foot or washed out of the hillslope system, contributing to the suspended sediment load of rivers and thus forming a constituent of the denudation system of the hillslope in a drainage basin. Whether the eluvial material is redeposited within the soil or reaches the river is largely a matter of the porosity and permeability of the soil and the overall water balance, thus eluviation and hillslope loss to a river is more characteristic of permeable soils in climates with a moderate or high rainfall whereas in less permeable soils and/or with climates with lower rainfall the eluvium is more likely to stay in the soils as redeposited illuvium. Thus, eluviation can form a significant denudation process where there are permeable soils and regoliths. Ruxton (1958) calculated that denudation of hillslopes in the Sudan by eluviation was almost as significant as that by the removal of material in solution, with around 25 per cent of removal by the former and around 35 per cent by the latter.

At the intermediate stage of deposition between the lower soil profile and loss to a river, the formation of clay plains at the foot of slopes along seepage lines can be quite significant. Ruxton (1958) reported such lateral sediment transport and deposition in the form of surface deposits of eluviated clay near the edges of weathered granite domes. The deposits were up to 500 m long, curved round the base of the slope, and up to 150 m wide, though with some longer tongues of deposition where there was evidence of greater water flow and some channelization. Steep (20°) slopes give way sharply to low-angle slopes of deposited fine material below. Here, there was evidently enough rainfall to wash the clay from the higher areas through the bedrock but insufficient runoff to transport the clay further than the slope foot.

Where there are landforms constituted from loose, unstable material, such as sand dunes or even under periglacial conditions with repeated frost heave and downslope movement, eluviation is not a dominant feature as the material is frequently in motion. However, if the material stabilizes – by vegetation growth on dunes or amelioration of climate – then eluviation can occur and this can be referred to as an eluvial phase of development for the soils and associated landform.

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STEVE TRUDGILL

ENDOKARST

Endokarst consists of the main part of karstic relief and contains carbonate rocks and cave systems (shafts, cavern, etc.). It involves all the underground features of the input karst and of the output karst. It is situated below the EPIKARSTic zone and is fully developed when the karstification is mature (Ford and Williams 1989). It can develop when the acid water from the surface (humic acid from biological activity and carbonic acid from CO_2 exchanges between atmosphere and rainwater) can reach the deepest part of the carbonate (or other karstifiable rocks) layers. It means that the openings of the stratification joints and fractures are large enough to allow the flow of water and suspended material. The development of endokarst is ruled by the competition between dissolution, carbonate precipitation and clogging with non-dissolved particulate material (Rodet *et al*. 1995). Infillings of endokarst conduits are frequently used to date the genesis and the evolution of cave systems (Maire 1990).

The thickness and the lateral extension of the karstifiable layers mark the boundaries of each endokarst. The world's largest cave systems explored by humans are known to exceed 10,000 km^2 and to reach some depths of more than 2,000 m.

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SEE ALSO: chemical weathering; epikarst; ground water; karst; palaeokarst and relict karst

MICHEL LACROIX

ENGINEERING GEOMORPHOLOGY

Engineering geomorphology deals with the geomorphic features of the Earth's surface, with special reference to their engineering properties. These properties include topography, rock units (lithology, rock mass strength, joint spacing, point load range, plasticity characteristics, compaction bearing strength, texture, etc.) soil units, water retention capacity, weathering, mass movement, erosion, etc. The results are very significant in sustainable land management (SLM) through combating processes of land degradation and DESERTIFICATION, and for obtaining a better planning system (see ENVIRONMENTAL GEOMORPHOLOGY).

Engineering geomorphological maps are prepared in a composite form from which collected, processed and stored data, required for a particular project, are extracted and analysed. Derivative maps are also obtained for specific purposes. An engineering geomorphological map (EGM) depicts the morphological and engineering properties of the terrain. It is very useful during the phase of policy or project formulation and implementation, the phase of development and construction, and the continuing management of a development, particularly in the context of civil engineering (see GEOMORPHOLOGICAL MAPPING; TERRAIN EVALUATION).

Various studies have been made in the field of APPLIED GEOMORPHOLOGY that particularly emphasize engineering applications (Cooke and Doornkamp 1974; Hails 1977; Verstappen 1982; Jha and Mandal 1997).

The methodology for the preparation of an engineering map (Table 16) involves three phases: (1) pre-fieldwork; (2) fieldwork; and (3) post-fieldwork. The pre-field phase includes the preparation of the base map from topographical sheets and cadastral maps. Delineation of the area is also done with high-resolution satellite imagery and aerial photographs at a 1:5,000 to 1:25,000 scale.

Table 16 Phases in engineering geomorphological mapping

Pre-fieldwork
Objectives
Framing of delineation rule
Delineation of the study area (topographical and cadastral maps, aerial photographs, and satellite imagery)
Classification criteria
Selection of engineering geomorphological properties
Selection of sites for collection of rock samples
Fieldwork
Field checks, scanning of engineering geomorphological properties
Collection of rock samples, soils, weathering, mass movement and erosional characteristics of rills and gullies
Rock fall, landslides, slope failures and scree zones
Post-fieldwork
Discussion
(Addition/alteration, etc.)
Field mapping corroboration

In the fieldwork phase all the rock units, soil units, tectonic elements, and active geomorphic processes are investigated. Also noted is the water retention capacity of the mass. Ultimately, the areal coverage and locations of occurrences are marked on the base map. The topographic features marked from the toposheets are also updated during field investigation. Rock samples are collected from different litho-stratigraphic units for laboratory analysis to obtain their geo-engineering properties.

The post-fieldwork phase involves laboratory testing of rock samples, transfer of field data relating to mass movement, weathering and erosion, and preparation of the final map and its interpretation. The final map can be prepared by transferring and plotting data obtained from different sources in a synthesized manner, and by using different symbols and colours. During this phase of work, aerial photographs and satellite images are consulted, in addition to the fieldwork, to identify and demarcate the exact location and boundaries of different rock mass strength units, soil units, zones of weathering and areas of active mass movements and erosion, etc. Finally, the engineering geomorphological map for the study area can be prepared.

Engineering geomorphological map preparation (see Figure 54) includes the following physical parameters: topography; rock units; tectonics; soil units; weathering; mass movement; and erosion. The engineering geomorphological map gives special emphasis to the following aspects of the rock mass: strength of intact rock; joint spacing; width of joints; bedding planes; gauge or infilling; the materials and water movements within the rock; point load range; water retention capacity; and compaction bearing strength.

Considering the above attributes of the rock units, the study area can be divided into four categories by superimposing the layering of information on the base map using a Geographic Information System (GIS).

1 Low rock mass strength unit (Rlo)
2 Medium rock mass strength unit (Rme)
3 High rock mass strength unit (Rhi)
4 Very high rock mass strength unit (Rvhi).

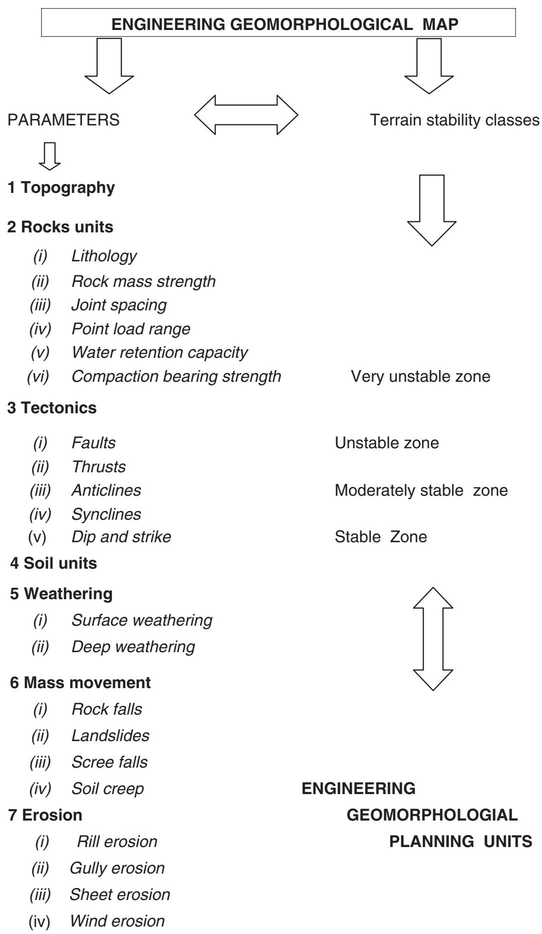

Figure 54 Parameters of the engineering geomorphological map

The engineering geomorphological map shows that all the parameters are functionally interrelated in the sample area. The result can be presented in four terrain stability classes: very unstable zone; unstable zone; moderately stable zone; and stable zone.

An engineering geomorphological map is an important tool in planning and development. It indicates the stability of terrain which is very significant in civil constructions, agriculture, industry, transportation networks and settlement establishment. As this type of map records all sorts of topographical, morphological, and geo-engineering data, so the development and planning of an area should be based on this type of mapping in which the land use and other planning aspects can be regulated according to the stability/suitability of the terrain.

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VIBHASH C. JHA

ENVIRONMENTAL GEOMORPHOLOGY

Environmental geomorphology is the practical use of geomorphology for the solution of problems where humans wish to transform or to use and change surface processes. According to Coates (1971, 1972–1975), this discipline involves the following issues and themes:

1. the study of geomorphic processes and terrain that affect man, including hazard phenomena such as floods and landslides;
2. the analysis of problems where man plans to disturb or has already degraded the land–water ecosystem;
3. human utilization of geomorphic agents or products as resources, such as water or sand and gravel;
4. how the science of geomorphology can be used in environmental planning and management.

Many other researchers have dealt with environmental geomorphology, in the discussion of both specific topics and the various applications of geomorphology in the forms of APPLIED GEOMORPHOLOGY, ENGINEERING GEOMORPHOLOGY and also engineering geology. It is not necessary to review the available literature here; it is sufficient to mention Tricart (1962, 1973, 1978), Verstappen (1968, 1983), Craig and Craft (1982) and Cendrero *et al*. (1992), among others.

More recently, Panizza (1996) defined *environmental geomorphology* as that area of Earth sciences which examines the *relationships between man and environment*, the latter being considered from the *geomorphological* point of view. It should be further specified that *environment*, in general, is defined (Panizza 1988) as the 'range of physical and biological components that have an effect on life and on the development and activities of living organisms'.

The geomorphological components of the environment may be schematically subdivided into: *geomorphological resources*; and *geomorphological hazards*. Geomorphological resources include both raw materials (related to geomorphological processes) and landforms: both of which are useful to man or may become useful depending on economic, social and technological circumstances. For instance, littoral deposits can become important, economically valuable and considered as geomorphological resources when used for sand quarrying. Similarly, a sea beach can acquire value and be considered as a geomorphological resource when utilized as a seaside resort. A landform can be considered a resource also from the scientific and cultural viewpoint: for example, a marine cliff can be seen as a model of geomorphological evolution.

GEOMORPHOLOGICAL HAZARDS can be defined (Coates 1972–1975) as the 'probability that a certain phenomenon of geomorphological instability and of a given magnitude may occur in a certain territory in a given period of time'. For example, in any one area, the possibility of a certain landslide occurring over a 50-year time span may be assessed. Hazard is therefore a function of the intensity/magnitude and of the frequency/probability of the phenomenon (Varnes 1984). The term ‘susceptibility’ as used in many mapping procedures (e.g. Brabb *et al.* 1972) corresponds to hazard by equating spatial probabilities to temporal probabilities.

In the context of the relationships with the environment, man represents: *human activity*; and *area vulnerability*. Human activity is the specific action of man which may be summarized under the headings of hunting, grazing, farming, deforestation, utilization of natural resources, engineering works, etc. Man’s interventions take place essentially on that thin layer of the Earth’s surface which makes up the interface between atmosphere and lithosphere where most energy exchanges and complex phenomena take place (Piacente 1996). Hardly ever are these phenomena confinable within preconstituted and rigid schemes, but nevertheless they can be summarized as follows (Castiglioni 1979): artificial forms, directly modelled by man’s activities; works aiming to divert, correct or upgrade natural processes; modifications of natural phenomena, indirectly resulting from man’s activities.

Area vulnerability is the complex of the inhabitants and all things that exist as a result of the work of man in a given area and which may be directly or indirectly sensitive to material damage. Included in this complex, we find the population, buildings and structures, infrastructures, economic activity, social organization and any expansion and development programmes planned for an area. In short, it corresponds to an ‘exposed element’. Vulnerability can also be defined (Varnes 1984; Einstein 1988) as the level of potential damage (ranging from 0 to 1) to a given exposed element, which is subject to a possible or real phenomenon of a given intensity: we prefer the first definition, which does not imply elements already included in the definition of hazard.

Considering the relationships between geomorphological environment and man, two main possibilities can be examined (Panizza 1992) (Figure 55):

1 Geomorphological resources in relation to human activity, where geomorphological environment is regarded as mainly passive in relation to man (active); in other words, a resource may be altered or destroyed by human activity (e.g. a mountain landscape that has been modified by a bulldozer). We define as *impact* these consequences of human activity on geomorphological resources. It consists of the physical, biological and social changes that human intervention brings about in the environment, the latter term being intended in its geomorphological elements. Therefore, this impact equals the ‘product’ of human activity and geomorphological resources.
2 Geomorphological hazard in relation to area vulnerability, where geomorphological environment is regarded as mainly active in relation to man (passive): in other words, a hazard may alter or destroy some buildings or infrastructures (e.g. a landslide or river

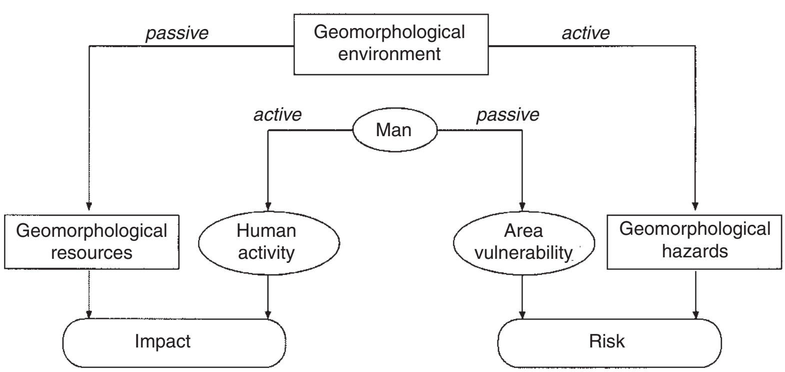

Figure 55 Relationships between geomorphological environment and man

erosion that cause road collapse). We define as *risk* these consequences of geomorphological hazard on a situation of area vulnerability. It is a natural risk connected to a geomorphological hazard: the term refers to the probability that the economic and social consequences of a particular phenomenon reflecting geomorphological instability will exceed a certain threshold. Therefore, this risk is equal to the 'product' of geomorphological hazard and an area's social and economic vulnerability. It corresponds to the term 'specific risk' by Varnes (1984), which expresses the loss due to a particular natural process.

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MARIO PANIZZA

EPEIROGENY

In his monograph on Lake Bonneville, Gilbert (1890: 340) formalized the definition of certain tectonic terms: 'Displacements of the Earth's crust which produce mountain ridges are called orogenic . . . the process of mountain formation is orogeny, the process of continent formation is epeirogeny, and the two collectively are diastrophism.' Bloom (1998: 43) concurred that there was need to describe non-orogenic tectonism (i.e. tectonic movements not associated with mountain belts) and so redefined epeirogeny as: 'continental vertical tectonic movement of low amplitude relative to its wavelength, not within an orogenic belt, that does not deform rocks or the land surface to an extent that is measurable within a single exposure.'

Such broad movements can be either positive (uplift) or negative (subsidence). Epeirogenic uplift can be attributed to MANTLE PLUMES beneath broad areas of continental crust and to such processes as glacio-isostasy. Rates of epeirogeny have generally been thought to be one or two orders of magnitude lower over similar time intervals to rates of orogeny, but studies of present rates of neotectonism suggest this is not invariably the case. In areas of active epeirogeny, river incision may occur (Wisniewski and Pazzaglia 2002).

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EPIKARST

The epikarst is the uppermost highly weathered layer of karst bedrock beneath the soil (Klimchouk 2000). It is also known as the subcutaneous zone (Williams 1983). Where there is no soil there is still an epikarst, for example beneath limestone pavements and alpine karrenfeld. The epikarst develops because rainwater is acidified by dissolving carbon dioxide in the atmosphere and especially in the soil, thereby producing weak carbonic acid. On percolating downwards from the surface into the bedrock, this water accomplishes most of its dissolutional work within 10 m of the surface, i.e. close to its main source of carbon dioxide. The result is that fissures in the limestone are especially enlarged by corrosion near the surface but taper with depth. Consequently, infiltration of rainwater into the karst is initially rapid, but vertical water flow encounters increasing resistance with depth as fissures become narrower and less frequent. This produces a bottleneck effect after particularly heavy rain, resulting in temporary storage of percolation water in a perched epikarstic aquifer.

Joints, faults and bedding-planes vary spatially within the rock because of tectonic history and variations in lithology. As a result the frequency and interconnectedness of fissures available to transmit flow also varies. Nevertheless, near the surface there is considerable interconnectedness in the horizontal plane; so recharging rainwater tends to be homogenized by lateral mixing. However, in the vertical plane some fissures are more favourable for vertical percolation than others, for example master joints that penetrate numerous beds and especially where several joints intersect. As a result these fissures develop as principal drainage paths. Water in the epikarstic aquifer flows laterally towards them and, as a result, they are subjected to still more dissolution by a positive feedback mechanism and so vertical permeability is enhanced. Water captured within the zone of influence of particular drainage routes becomes increasingly isolated as it percolates downwards from water elsewhere in the epikarst, and so despite the early homogenization it gradually acquires a water quality that reflects the residence time in the epikarst.

The saturated zone in the epikarst is especially well developed after heavy rain, when the epikarstic saturated zone is suspended like a perched aquifer above the main phreatic zone in the karst. The piezometric surface (water table) of the epikarst draws down over a preferred leakage path similar to the cone of depression in the water table over a pumped well. Streamlines adjust and resulting flow within the epikarst is centripetal and convergent on the drainage zone. The diameter of any solution doline that ultimately develops as a consequence of the focused dissolution is determined by the radius of the epikarstic drawdown cone.

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PAUL W. WILLIAMS

EQUIFINALITY

Equifinality is the principle which states that morphology alone cannot be used to reconstruct the mode of origin of a landform on the grounds that identical landforms can be produced by a number of alternative processes, process assemblages or process histories. Different processes may lead to an apparent similarity in the forms produced. For example, sea-level change, tectonic uplift, climatic change, change in source of sediment or water or change in storage may all lead to river incision and a convergence of form. The usage of the term in this way stems from Chorley (1962) but the related concept of converging landforms was developed earlier by Mortensen (1948), who pointed out that there are many convergences in the landforms of arid and polar regions even though their climates (and therefore by implication their geomorphic process assemblages) are so different.

Perhaps one of the better illustrations of the problem of equifinality concerns the origin of TORS. Four principal theories are held to explain identical landforms: (1) subaerial weathering causes spheroidal modification to the morphology of outcrops produced by differential erosion of the bedrock; (2) exposure of tors is due to a two-stage process: a period of prolonged subsurface groundwater weathering leading to decay of closely jointed rock and spheroidal modification of larger blocks, followed by a period of erosional stripping leading to exposure of the tor at the surface; (3) reduction in area of larger inselbergs by scarp retreat and the formation of pediments; and (4) tors are isolated as a result of freeze–thaw weathering, followed by solifluction over permafrost in a periglacial climate. In the last analysis, it is probable that all four processes are reasonable alternative explanations of the origin of this landform. It is generally accepted that no final descriptive definition is without ambiguity. Hence it is not possible to argue from the presence of this landform alone that a certain sequence of genetic events has occurred. This is the classical concept of equifinality.

Brunsden (1990) in discussing his Proposition 10 – the ability of a landscape to resist impulses of change tends to increase with time – notes that in spite of the existence of complex causes and complex responses in geomorphology 'there is within any tectono-climatic domain a tendency toward an all pervading unity and a repetitive but characteristic geometrical order and regularity.' He explains this tendency as resulting from preferential selection of stable forms; exponential decrease in rate of change; increasing effectiveness of barriers to change; constancy of process; persistence; convergence; over-relaxed systems; self-propagation; preferential fabric relief patterns; and process smoothing and extreme event accumulation.

Haines-Young and Petch (1983) suggest that the concept has been misused in that geomorphologists have invoked equifinality in order to avoid the hard question of specific mode of origin of the landforms in question and, they claim, too rapid an acceptance of equifinality may inhibit the development of general laws or may lead to detailed differences of form being overlooked. They suggest a redefinition of the concept as follows: 'a single landform type is said to exhibit equifinality when it can be shown to arise from a range of initial conditions through the operation of the same causal processes' (1983: 465). In this context, they commend Culling (1957) for his use of the graded stream as an example of equifinality in the sense that whatever the initial conditions, a graded stream will display a similar long profile.

Culling (1987) suggests that the word 'equifinality' is no longer useful because advances in our understanding of dynamic systems have opened up a new and richer world with its own more flexible vocabulary. The idea that a system will strive to arrive at similar positions in phase or state space despite differing initial conditions has become familiar to students of general systems theory. The existence of strange attractors in non-linear dissipative dynamic systems is also reminiscent of the older idea of equifinality, but there are equal evidences of chaotic motion in systems that are fully determined and predictable with accuracy in the immediate future. Because of the ubiquity of noise, all stable systems are transient. It is the recognition of complicated periodic behaviour at points far from equilibrium and its interaction with strange attractors that upsets a simple definition of equifinality. Culling proposes a complex topology of degrees of equifinality, which depart from the definition of equifinality *sensu stricto*. That definition remains 'that upon perturbation a system will eventually return to its initial position'. It is important to realize that such a condition is itself transient. Therein lies the essence of the flexibility of the new approach to the concept of equifinality.

Phillips (1995) discusses the value of viewing landscape evolution as an example of self-organization. Self-organization or self-regulation depend upon the dominance of negative feedback in the system. He points out that geomorphic systems give evidence of both self-organization (as in the case of at-a-station hydraulic geometry) and non-self-organization (as in soil landscape evolution). The challenge for geomorphology is how to distinguish between such systems. Self-similarity and equifinality are incompatible with non-self-organizing systems.

Culling (1987) argues for a variety of looser definitions of equifinality than that of the strict definition above. He suggests that a quasi-equifinality can be defined to include approximate return to initial conditions, by placing a restriction on the allowable magnitude of perturbation, by defining a system domain whose manifold has several local minima and in accepting the retention of certain ergodic and topologic properties as adequate criteria for equifinality.

The debate is reminiscent of the dynamic equilibrium debate of thirty years ago, but with two developments: the level of analytical sophistication has increased and, perhaps more importantly, the debate is open ended and does not yield a unique conclusion. 'In looking once again at the concept of equifinality, it is as if we had opened some magic casement to find, between chance and necessity, one dimension and the next, a whole new world of chaotic motions, strange attractors and periodic windows. With a wild surmise we gaze upon an ocean of discovery between two continents previously thought contiguous' (Culling 1987: 69).

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SEE ALSO: non-linear dynamics; tor

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EQUILIBRIUM LINE OF GLACIERS

The position on a glacier where seasonal accumulation equals seasonal ablation is termed the equilibrium line (refer to Figure 56). At this area on a glacier the net mass balance is at zero (no ice mass is lost or gained at this point). A glacier gains mass during winter as snow falls, causing a positive annual mass balance above the equilibrium line. Below the equilibrium line, the annual mass balance of glaciers is negative due to ablation during the summer melt season (Sugden 1982).

The equilibrium line altitude (ELA) for a particular glacier budget year is considered synonymous with the end of summer snowline (EOSS). The altitude of the annual EOSS averaged over many years, defines the steady-state ELA. The annual snowline position with respect to the long-term or steady-state ELA can be used as a surrogate or index of the annual mass balance changes of a glacier. Changes in glacier mass balance are a direct, undelayed response to changes in atmospheric conditions (Fitzharris *et al.* 1997) and hence can be a useful indicator of larger scale changes in global climate.

A commonly used method to work out the ELA calculates the area of the glacier, comparing this to the accumulation area. This is termed the accumulation area ratio (AAR). Glacier studies worldwide have demonstrated that the AAR for glaciers with stable ELA has a value of around 0.6 (Lowe and Walker 1997). Where it is possible to estimate the extent of late Pleistocene glaciers, this method can describe the change in ELA compared with the present, and consequently the changes in climate over time.

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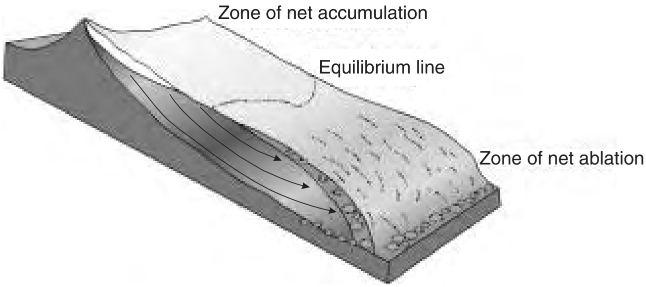

Figure 56 The longitudinal profile of a valley glacier, showing the area of seasonal accumulation and ablation, and the equilibrium line

EQUILIBRIUM SHORELINE

A beach face is in a state of DYNAMIC EQUILIBRIUM when the same amount of sediment is moved landwards by the stronger uprush as is moved seawards by the weaker backrush. This is accomplished by adjustments to the gradient and shape of the BEACH. The gradient is largely determined by the amount of water that percolates into the beach and is lost to the downrush, which is primarily a function of grain size. Pebble or shingle beaches (see SHINGLE COAST) are particularly steep because of rapid percolation and the consequently very weak backrush. Much less water percolates into fine-grained sandy beaches, however, and the weak gravitational effect of a gently sloping beach face is therefore able to compensate for small differences in the onshore and offshore transport rates. There is also a tendency for beach gradient to decrease as wave steepness increases, presumably because the greater velocity of the uprush makes it easier to carry sediment up the slope. The gradient of beaches, or portions of beaches, with the same grain size can therefore vary according to differences in exposure and wave steepness, while temporal variations in wave steepness explain why beach gradient changes during and following storms. There is also evidence to suggest that beach slope is partly determined by the height or energy of the WAVE, which would explain why, for the same wave steepness, slopes are generally greater in low than in high energy environments. The proportion of heavy minerals in a beach, which increases the resistance to removal by backrush, may also be significant in determining the equilibrium gradient.

It is difficult to know if beaches are in quasi-equilibrium, and it can also be difficult to define the slope of beaches that consist of more than one slope element, although it is usually measured in the swash zone where it is essentially linear. A variety of descriptive, empirical and mathematical models are concerned with the relationship between equilibrium beach slope, grain size and wave parameters. The first attempts to model beach gradient were largely statistical and concerned with correlations with wave and sedimentological parameters. Analytical models solve equations for equilibrium gradient on the assumption that there is no net sediment transport, whereas iterative models simulate beach development until an equilibrium slope has developed.

Bruun (see BRUUN RULE) suggested that the shape of equilibrium beach profiles can be represented by the power law:

$$h_x = A\ x^{2/3}$$

where h_x is the depth at a distance x offshore of the mean water line and A is a scale parameter that is largely determined by grain size or fall velocity. If there has been significant sorting, however, coarser sediment can make the shoreward portions of beaches steeper than model predictions, and finer sediment can make the seaward portions more gentle than predicted. Although other workers have provided alternate expressions for the geometry of the equilibrium profile, the use of a single equilibrium equation to represent all beach profiles has been criticized, and the concept of a profile of equilibrium has been questioned.

Beaches adjust their equilibrium morphology and sediments with variations in waves, tides and other influences (Short 1999). Two profiles represent the extremes of a fairly predictable range of forms that can be assumed according to the size or power of the waves. The distinction has been made between reflective profiles with wide berms or swash bars and steep foreshore slopes, sometimes with steps at the breaker line, and dissipative profiles with gentler foreshore slopes and longshore submarine bars (see BAR, COASTAL). Reflective profiles change into dissipative profiles during storms, when large waves move sediment seawards, whereas the reverse occurs when smaller swell waves move sediment back onshore. Frequent changes in wave power are responsible for cycles that are frequently much shorter than that between the two extremes, and wave environments therefore tend to generate globally distinctive beach state characteristics. The mid-latitudes, for example, have persistently high wave power and the beaches are generally kept in a highly dissipative state, whereas beaches in low swell or sheltered environments are normally in reflective states. Beach states also change with tidal level, and the microtidal model has to be modified for areas with a high tidal range. Equilibrium beach profiles and beach states have been modelled as a function of the relative tidal range (RTR) – the ratio of tidal range to breaker height and the dimensionless fall velocity of the sediment, with high RTR values representing tide-dominance and low values wave-dominance.

Coastlines trend towards an equilibrium state in the longshore as well as in the cross-shore direction. The distinction can be made between swash- and drift-aligned equilibrium beach forms (Plates 39 and 40). Swash-aligned beaches are parallel to the incoming wave crests and net LONGSHORE (LITTORAL) DRIFT is at a minimum. Drift-aligned beaches are parallel to the line of maximum drift and sediments can be carried great distances in one direction. Swash-aligned beaches are associated with irregular coasts where longshore transport is impeded, and the important wave trains reach the shoreline almost normally. Drift-aligned beaches develop where the initial coastal outline is fairly regular, or where important sediment-moving waves approach the coast at an angle.

Drift alignment and dynamic beach equilibrium require a constant supply of sediment, as for example when a coastal cell is coupled to the mouth of an ESTUARY. Static beach equilibrium can be attained in several ways (Carter 1988):

1. through swash-alignment, when sediment movement is restricted to cross-shore transport;
2. by strong wave height gradients or the interaction of two wave trains causing the longshore current velocity to become zero; and
3. by the alongshore grading of beach sediments in such a way that the strength of the current at each place is too low to entrain it.

All three situations are common, and in some cases equilibrium is attained through a combination of options.

Plate 39 Swash-aligned beach at San Martinho do Porto, Portugal

Plate 40 Drift-aligned beach, St Petersburg, Florida

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EQUILIBRIUM SLOPE

Equilibrium slopes are hillslopes which are characterized by an equilibrium of forces that compensate each other. An equilibrium slope exists 'if the amount of material that is removed from an areal unit of the surface per time unit is equal to the amount of material that is supplied to this areal unit during the same time' (Ahnert 1994). This definition follows the conception that a geomorphic system, e.g. a hillslope, is under equilibrium conditions if the 'mass budget' of that system does not change. Hillslopes are in equilibrium conditions if the processes acting upon the hillslope are in equilibrium. Each change of this equilibrium will result in adjustments of the acting processes towards a new equilibrium.

This definition goes back to the fundamental work of Grove Karl Gilbert on the Henry Mountains (Gilbert 1877). In this publication Gilbert introduced the concept of equality of action: 'Erosion is most rapid where the resistance is least, and hence as the soft rocks are worn away the hard are left prominent. The differentiation continues until an equilibrium is reached through the law of declivities (slope gradient). When the ratio of erosive action as dependent on declivities becomes equal to the ratio of resistances as dependent on rock character, there is equality of action.' This situation is called 'dynamic equilibrium' because the system equilibrium is attained by mechanisms of self-regulation, where a change of process components caused by a change of input will result in a compensation between these process components by negative feedbacks (see DYNAMIC EQUILIBRIUM). The negative feedback between processes governing the system causes adjustments to the changes of inputs (e.g. a climatic change or a BASE LEVEL lowering) towards a new equilibrium. In this way the two central aspects (1) of mass transport rates and (2) negative feedback mechanisms were established in geomorphology.

Gilbert (1877) distinguished two types of transport laws of hillslopes: weathering-limited and transport-limited regolith removal. Weathering-limited transport occurs where the weathering rate is lower than the transport capacity of the hillslope forming processes, so that the regolith is removed and slope development is related to the weathering rate of rocks. In this case the slope system is in non-equilibrium, the inverse of equilibrium. The material supply by different slope processes (weathering, slope wash, soil creep, etc. from upslope) is smaller than the potential rate of removal. These form elements can be found in arid and semi-arid environments, in mountain areas and on free faces of cliffs and all stream channels in bedrock. Transport-limited transport occurs where the weathering rate is higher than the transport capacity of the hillslope-forming processes, so that regolith accumulates and the transport processes operate at their full capacities.

The concept of equilibrium slopes has been applied to numerous investigations in geomorphology related to slope evolution by river incision and undercutting followed by mass movements if internal frictional threshold angles of the regolith are crossed. Examples are given, for instance, by the pioneering research of Strahler (1950), who related statistically maximum valley-side slope gradients with the frictional threshold angles of up to 1-m thick regolith cover. Similar equilibrium approaches were published by Young (1972) and Carson (1975).

Further approaches are related to finding characteristic form slope profiles (equilibrium profiles) for a range of transport processes by empirical modelling transport capacity relationships. Kirkby (1971) used this relationship to derive characteristic equilibrium slope profiles for soil creep, soil wash without and with gullying and rivers. Ahnert (1976) developed a more complex computer model which generates five different equilibrium slope profiles related to splash (convex), suspended load-wash (convex–straight–concave), point-to-point wash (rolling) (convex–straight), plastic flow and viscous flow (convex–straight) processes. These models are based on equilibrium assumptions concerning mass transport rates of different processes and feedback mechanisms between them.

There are extreme events that destroy the equilibrium on hillslopes, e.g. by removing the regolith by extreme rainfall, landsliding, gully erosion or vegetation change. In this situation the process rates change significantly in time, which means that the system is in a state of disequilibrium. If the entire slope system has the tendency towards a dynamic equilibrium, negative feedbacks adjust the slope after these external impacts. The period of recovery from this event depends on the constitution of the system itself and on the magnitude of the external impact. If the regolith coverage has been removed and the bare rocks are exposed, system response will result in an adjustment by an increase of the rate of the weathering processes.

In a recent debate Ahnert (1994) and Thorn and Welford (1994) reviewed different equilibrium definitions and concepts in geomorphology. The authors were especially concerned with a high degree of confusion generated by different types of equilibrium concepts used. Based on the very clearly defined concept of Ahnert (1994), Thorn and Welford (1994) suggested the use in the future of a mass-based equilibria concept which is based on field data, namely mass volume and mass flux. They suggested one should abandon the term 'dynamic equilibrium' and use the term 'mass flux equilibrium' to avoid associations with former definitions and concepts and to reach a clearer coupling with other disciplines.

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SEE ALSO: dynamic equilibrium; equilibrium shoreline; hillslope, form; hillslope, process

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ERGODIC HYPOTHESIS

Ergodicity is an idea developed in physics. In studying the movement of molecules in a macroscopic system (such as a room full of air), physicists faced a difficult problem: innumerable molecules move very fast compared with the time taken to observe them. To overcome this problem, they devised the ergodic theorem and the ergodic hypothesis, the word ergodic coming from the Greek *ergon* ('work' or 'energy') and *hodos* ('road') and meaning a 'path of constant energy'.

To appreciate the ideas behind ergodicity, take the case of people in a maze. Now, a maze is a network consisting of a number of links joined at nodes. Imagine one person entering the maze, which has no exit, and wandering around long enough to have entered all possible links at least once. By keeping a record of the path taken by the person, it is possible to calculate two probabilities. First, is the probability of the person's being in a particular link after a given time. Second, is the probability that the person has spent so many minutes in a particular link. Alternatively, in a new experiment, imagine that a large number of people enter the maze (again the maze has no escape route). After sufficient wandering (sufficient for an equilibrium to obtain), the probability of a person's being in a particular link may be given as the ratio of the number of people in that link to the total number of people in the maze. Therefore, by taking an aerial picture of the maze, it is possible to say how much time a person would spend in each link had he or she wandered around the maze for a long time. The first case specifies the relative amount of time spent by one person in each link; the second case specifies the relative number of people in different parts of the maze in an instant of time after equilibrium prevails. The system is ergodic when these two probabilities are the same. In formal language, the statistical properties of a time series of a phenomenon (the individual maze-wanderer) are essentially the same as a set of observations made on a spatial ensemble (the spatial distribution of collective maze-wanderers) at a single time. In other words, ensemble averages can replace time averages in large-scale statistical statements. The individual maze-wanderer exemplifies the ergodic hypothesis, the collective maze-wanderers the ergodic theorem, which states that sampling across an ensemble (the people in the maze) is equivalent to sampling through time for a single system (the lone maze person).

How does this reasoning apply to geomorphology? One might discover that of all slopes in a region, 9 per cent stand at 6 degrees. If ergodic conditions apply, then the ergodic hypothesis would predict that the region would have 6-degree slopes for 9 per cent of its lifetime. In practice, few geomorphic applications of ergodicity make such quantitative statements about time using spatial data. This dearth of applications results largely from the strict conditions required for ergodic arguments to hold, including the difficulty of finding equilibrium landforms in an environment that is constantly changing. The few geomorphological applications that do meet the stringent statistical demands of the ergodic assumptions include studies of geomorphic magnitudes and frequencies, 'threshold' hillslopes, and the growth of drainage basins (see Paine 1985); river channel evolution (Zhang *et al.* 1999); and a general analogy between statistical thermodynamics and the transfer of mass within a landscape (Scheidegger 1991: 254).

A far commoner practice in geomorphology is to study change through time by identifying similar landforms of differing age at different locations, and then arranging them chronologically to create a time sequence or topographic chronosequence. Such space–time, or – more strictly – location–time, substitution has proved salutary in understanding landform development. Two broad types of location–time substitution are used. The first looks at equilibrium ('characteristic') landforms and the second looks at non-equilibrium ('relaxation') landforms.

In the first category of location–time substitution, the assumption is that the geomorphic processes and forms being considered are in equilibrium with landforms and environmental factors. For instance, modern rivers on the Great Plains display relationships between their width–depth ratio, sinuosity and suspended load, which aid the understanding of channel change through time (Schumm 1963). Allometric models are a special case of this kind of location–time substitution (see Church and Mark 1980).

Studies in the second category of location–time substitution, which look at developing or 'relaxation' landforms, bear little relationship to the ergodicity of physics. The argument runs that similar landforms of different ages occur in different places. A developmental sequence emerges by arranging the landforms in chronological order. The reliability of such location–time substitution depends upon the accuracy of the landform chronology. Least reliable are studies that simply assume a time sequence. Charles Darwin, investigating coral-reef formation, thought that barrier reefs, fringing reefs and atolls occurring at different places represented different evolutionary stages of island development applicable to any subsiding volcanic peak in tropical waters. William Morris Davis applied this evolutionary schema to landforms in different places and derived what he deemed was a time sequence of landform development – the Geographical Cycle – running from youth, through maturity, to senility. This seductively simple approach is open to misuse. The temptation is to fit the landforms into some preconceived view of landscape change, even though other sequences might be constructed.

More useful are situations where, although an absolute chronology is unavailable, field observations enable geomorphologists to place the landforms in the correct order. This occasionally happens when, for instance, adjacent hillslopes become progressively removed from the action of fluvial or marine processes at their bases. This has happened along a segment of the South Wales coast, in the British Isles, where the Old Red Sandstone cliffs between Gilman Point and the Taf estuary have been affected by a sand spit growing from west to east (Savigear 1952). In consequence, the westernmost cliffs have been subject to subaerial denudation without waves cutting their bases the longest, while the cliffs to the east are progressively younger.

Relative-age chronosequences depend upon some temporal index that, though not fixing an absolute age of landforms, enables the establishment of an interval scale. For example, the basin hypsometric integral and stream order both measure the degree of fluvial landscape development and are surrogates of time (e.g. Schumm 1956).

The most informative examples of location–time substitution arise where absolute landform chronologies exist. Historical evidence of slope profiles along Port Hudson bluff, on the Mississippi River in Louisiana, southern USA, revealed a dated chronosequence (Brunsden and Kesel 1973). The Mississippi River was undercutting the entire bluff segment in 1722. Since then, the channel has shifted about 3 km downstream with a concomitant stopping of undercutting. The changing conditions at the slope bases have reduced the mean slope angle from 40° to 22°.

Location–time substitution does have pitfalls. First, not all spatial differences are temporal differences because factors other than time exert an influence on landforms. Second, landforms of the same age might differ through historical accidents. Third, equifinality, the idea that different sets of processes may produce the same landform, may cloud interpretation. Fourth, process rates and their controls may have changed in the past, with human impacts presenting particular problems. Fifth, equilibrium conditions are unlikely to have endured for the timescales over which the locational data substitute for time, especially in areas subject to Pleistocene glaciations. Sixth, some ancient landforms are relicts of past environmental conditions and are in disequilibrium with present conditions.

Many geomorphologists substitute space for time to infer the nature of landform change. Only a handful of these adhere to the statistical assumptions of ergodicity. Nonetheless, the loose application of the ergodic reasoning, as seen in location–time substitution, is a productive line of geomorphological enquiry.

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RICHARD HUGGETT

ERODIBILITY

Erodibility is the resistance of surface material to erosion. It is usually restricted to soils or REGOLITH, and to water or WIND EROSION OF SOIL. Both water and wind erosion are complex processes, but when other factors are constant, erosion rates still vary due to differences in soil resistance. Erodibility is influenced by climate and is a complex, dynamic characteristic, which changes significantly over annual, seasonal or irregular time intervals, or even during a single storm. Nevertheless, the concept is useful for small-scale field or hillslope processes of rainsplash, sheetwash (see SHEET EROSION, SHEET FLOW, SHEET WASH) and rill erosion. It is more difficult to use with processes such as GULLY erosion, which involve very different spatial and temporal controls.

The erosion sub-processes affected by soil erodibility, are entrainment (by which soil particles are picked up), and transport. The relevant properties vary with the erosion process, and affect the erosive force available and resistance. In rainsplash erosion the entraining force is the kinetic energy of raindrop impact (see RAINDROP IMPACT, SPLASH AND WASH), converted to an upward force, while sheet wash and rill erosion involve runoff for both entrainment and transport. Raindrop impact can also disrupt particles and change their resistance to movement. The effectiveness of the upward force depends on soil particle size and mass. Poesen and Savat's (1981) experiments showed an entrainment:particle size relationship resembling the Hjulstrom curve for flowing water, particles between 0.1 and 0.25 mm diameter requiring least energy for movement. These results apply to non-cohesive particles with uniform density, such as quartz sands, where there is a direct relationship between particle size and mass. On such soils, erodibility can be assessed by standard particle size analysis techniques.

The relationship between particle size and erodibility is more complex when the surface is largely composed of aggregates. Aggregates are mixtures of mineral and organic matter, joined by electrostatic charges, microbial muscilages, hydrous oxides and carbonates. These materials and the volume of pore space are quite variable, so aggregate density is also variable. Microaggregate ($<$0.25 mm diameter) density is usually much higher than for macroaggregates, which may exceed 10 mm diameter (Oades 1993). Some large aggregates have densities below 1 $g.cm^{-1}$, can float on water, and are more easily eroded than small aggregates or mineral particles. The relationship between aggregate size and mass is not linear or direct, and size:entrainment relationships can be very complex.

These relationships are further complicated as aggregates can disintegrate during rainfall, if subjected to stresses exceeding the strength of 'stabilizing' agents. Under rainsplash, three dominant stresses occur. Raindrop energy, which can disrupt aggregates, is most significant, but SLAKING and differential hydration swelling may also cause breakdown. Aggregate strength is derived mainly from *cohesion*, due to electrostatic forces that bond clay and humus particles. Bond strength depends on total clay and humus content, on cation adsorption capacity, and on the cations adsorbed. Bivalent cations, such as calcium, cause flocculation and strong bonding, yielding small, highly resistant aggregates. Monovalent cations, such as sodium, cause particle repulsion, yielding weak, easily dispersed aggregates.

These interactions ensure that aggregates vary greatly in size, shape, stability and density, even in a single soil. As many soils consist largely of aggregates, it is their properties, rather than those of mineral particles, which most strongly influence erodibility. Aggregation is also affected by extrinsic factors such as wetting–drying and freeze–thaw, and by soil organic matter, changing with inputs of plant litter and decomposition by microbial activity. It is normal for regular or irregular seasonal changes in aggregation to be superimposed on short-term aggregate dynamics during or between storm events (Bryan 2000).

This discussion has emphasized the role of particles or aggregates as individual units. In fact, they only behave this way in recently disturbed or dispersive soils. The most common cause of soil disturbance is tillage. After tillage, disturbed soils gradually regain *coherence* due to weathering, compaction and crusting (see CRUSTING OF SOIL) by raindrop impact. This may occur in one rainstorm, or may take many months, but strongly affects erodibility. Erodibility usually declines as soils regain coherence, as sufficient force is required to overcome soil coherence and to entrain loosened particles.

Erodibility of coherent soils is determined by *soil shear strength*, the resistance to *interparticle failure*, defined by the Coulomb equation as:

$$r = c + zy \tan o$$

The active components are internal friction (o) which integrates surface and interlocking friction of particles, and (c) cohesion. Cohesion is explained above, while internal friction depends on the strength, heterogeneity of mineral particles and aggregates, and on overburden pressure (zy) at the point of potential failure. In surface soils acted on by rainsplash, sheetwash and rill erosion, overburden pressure is negligible and shear strength is dominated by cohesion. Both cohesion and internal friction are strongly influenced by soil water content. Cohesion is modified, either positively or negatively, by water molecules between particles. In completely dry soils, these are absent and soils are usually non-coherent and highly erodible. As soil water content increases, thin water films form (often <1 micron in thickness) which are viscous and hold mineral particles together. As these become thicker, viscosity declines and cohesion is reduced. Internal friction also declines in wet soils, as positive pressure in water-filled pores counteracts overburden pressure. The strength of both coherent soils and soil aggregates is thus greatly reduced when soil water content approaches saturation, and erodibility increases.

The role of soil water means that erodibility also depends on soil hydrological properties. The overall control is climatic, but soil hydrological properties determine the proportion of water reaching the surface that infiltrates the soil, and how long the soil remains wet after a storm. Soil *infiltration capacity* depends on surface porosity, but on bare soils under intense rainfall, soil crusting or sealing often produces a thin almost impermeable surface layer. Once water infiltrates, its distribution and ultimate drainage depend on soil permeability, which is controlled by the *soil structure*, the arrangement of soil material and pore space, which determines such features as pore space continuity.

The properties described affect erodibility for all the processes discussed, but the relationship is somewhat different in each case, because of the role of surface water layers. Rainsplash often occurs without any surface water layer, and a significant water layer can reduce or eliminate erosive energy. In sheet wash and rill erosion, it is the *shear stress* exerted by runoff which causes entrainment and transport. As the existence and depth of the surface water layer is determined by the ratio of rainfall:infiltration capacity, this means that for sheetwash and rill erosion, shear stress is partially controlled by soil properties. The magnitude of shear stress exerted by runoff is also affected by water distribution, increasing significantly when runoff is concentrated on a small surface area. Flow concentration depends on surface roughness, which is controlled by vegetation and on bare surfaces by soil particle and aggregate heterogeneity. These properties determine whether, under raindrop impact, the surface becomes progressively rougher or smoother, and will therefore affect the distribution and magnitude of runoff shear stress.

The complexity of the interacting processes and properties involved means that erodibility is not a single, simple, measurable soil property (Lal 1990), but reflects the collective interaction of many soil properties. Nevertheless, it is often necessary to attempt to assess erodibility by one or several simple measures. Many attempts have been made to identify or combine soil properties as *indices of soil erodibility* (Bryan 1968). No single measure is successful in all cases, but several measures can be effective if the precise nature of the dominant erosion process is clearly identified.

Measures of aggregate water stability, such as wet-sieving, are useful for rainsplash erosion, particularly on disturbed soils, while soil shear strength measured with a *vane shear* apparatus can be useful on coherent soils. *Soil consistency*, assessed by Atterberg limits, can indicate crusting potential. As sheetwash and rill erosion are more complex processes, it is more difficult to isolate a reliable index, but measures based on range of aggregate or particle size may be promising. In all cases, however, the high temporal and spatial variability of erodibility must be recognized.

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RORKE BRYAN

EROSION

Commonly speaking the term erosion (from Latin *erodere* = to gnaw away) is often used to indicate the overall exogenic process or group of processes that are directed at levelling off Earth relief, in contrast with the antagonist endogenic processes (crustal movements and volcanism) that build it up. In this very wide meaning erosion includes: acquiring materials from the higher elevations, moving them from one place to another (*transport*) and leaving them in lowlands (*deposition*).

Actually, it is the opinion of all scientists that erosion cannot include deposition. In fact, in a more technical language, the term erosion – although variously defined – usually excludes the processes whereby transported materials are set down. In the most broad and common of the technical meanings, erosion includes all exogenic processes, excluding WEATHERING and MASS MOVEMENTS, that involve the entrainment of loose weathered materials by a mobile agent, the removal of bedrock particles by the impact of transported materials, the mutual wear of rock fragments in transit and the transportation of acquired materials (Thornbury 1954).

Sometimes the term is restricted by excluding transportation; in this case DENUDATION is the more general term. More rarely, erosion indicates exclusively the entrainment of loose materials by a mobile agent.

Erosional agents and processes

Erosional processes are performed by mobile agents that draw their energy from solar radiation and act in one or more ways, constantly driven by the force of gravity. The principal erosional agents are running water, glaciers, wind and sea waves (Figure 57). In some cases they complete the same process, in some others a given process is accomplished by a distinctive agency that operates according to its physical peculiarities. Beside the cited 'natural' agents, humans can be rightly considered an important erosional agent too. Nowadays anthropogenic activities are so widespread and marked that they deeply modify the Earth's surface, often in an irreversible manner (see ANTHROPOGEOMORPHOLOGY).

The most common processes performed by natural erosional agents are shown in Table 17. The entrainment of rocky particles by erosional agents can be both chemical and mechanical. The first action (CORROSION) implies the work of a solvent and therefore it is typically accomplished by running water or waves; it is less important than mechanical action.

Mechanical removal takes place with different modalities, depending on the erosional agent. *Hydraulic action* comes from the pressure and hydraulic force of flowing water or sea waves that allow the acquisition of rocky particles. CAVITATION is a particular process operated by running water; it is still poorly documented and would represent a mechanism through which hydraulic action has a direct role in bedrock breakage. This process occurs when an increase in flow velocity and the following decrease in pressure cause the formation of bubbles that implode emitting jets of water capable of fracturing solid rocks. Moving ice acquires materials by *plucking* (or *quarrying*); through this process glaciers that move forward can remove large rocky fragments already detached from the bedrock by the freezing of water circulating inside cracks. *Overdeepening* is the process whereby glaciers erode to levels below regional BASE LEVEL related to fluvial systems;

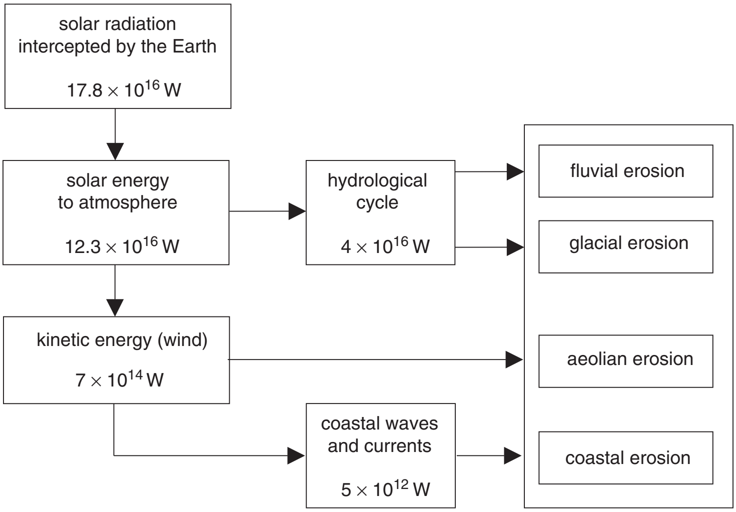

Figure 57 Source and flow of energy available for the different kinds of erosional processes. (From Summerfield 1991: 21, simplified and modified.)

Table 17 Erosional agents and their relevant erosional processes

Erosional Agent	Erosional processes			
	Entrainment of rocky materials	Erosion by transported materials	Wear of transported materials	Methods of transportation
Running water	Hydraulic action (Corrosion)	Abrasion	Attrition	Traction Suspension Solution
Glacier	Plucking or quarrying	Abrasion	Attrition	Traction
Wind	Deflation	Corrosion or abrasion	Attrition	Traction Suspension
Waves and currents	Hydraulic action (Corrosion)	Abrasion	Attrition	Traction Suspension (Solution)

Source: From Thornbury (1954: 47, modified)
Note: The less effective processes are indicated in brackets

however, it would be more correct to consider overdeepening as one of the effects of glacial erosion rather than an erosional process (Castiglioni 1979). The turbulent eddy action of wind is responsible for *deflation* and produces effects that are similar to those deriving from hydraulic action of waters.

Erosion accomplished by transported materials (named ABRASION) is due to the continuous collisions and friction by particles in transit on bedrock. All erosional mobile agents carry out this process. Abrasion due to running water is operated by solid materials of any size (up to large boulders, depending on the flow velocity) that can be transported as BEDLOAD by the water flow. Breaking waves that throw solid particles against the shore perform the same abrasive effect. EVORSION is a particular kind of abrasion due to the action of running water. It is caused by the erosional action of vortices and eddies on stream rocky beds. When a stationary eddy rotates a pebble, a small hollow is produced; this process leads to the formation of POT-HOLES (evorsion hollows), which contribute to valley deepening. In glacial environments abrasion is the friction produced on the bedrock by the debris carried along in the basal parts of glaciers; in its broader meaning it can include STRIATION, i.e. the bedrock *scoring* and *polishing* that reduces the rock surface roughness (Benn and Evans 1998). Aeolian abrasion derives from the repeated impacts of sand grains, silt particles and dust on rock surfaces; it is more properly named *corrasion*.

The wear of transported solid particles (*attrition*) takes place through repeated reciprocal knocking and collisions among the materials in transit: the result is a progressive decrease of particle sizes. At the same time the rocky particles tend to assume particular shapes, depending on the different ways in which each agent accomplishes transportation.

Mobile agents carry out transportation in three different ways. *Traction* consists of the rolling, sliding, pushing or jumping (in which case SALTATION is the specific term) of transported particles that are swept along on or immediately above a bottom surface. *Suspension* is a mode of transportation by water and wind; it implies the holding up of transported particles by the upward currents that develop in turbulent flows like those of running water and moving air (see SUSPENDED LOAD). *Solution* is a kind of 'chemical' transportation that is restricted to the action of running water and waves.

Erosional processes produce distinctive erosional landforms; furthermore, each erosional agent develops its own typical assemblage of landforms, depending on its mode of shaping Earth's relief. Erosional landforms are particularly striking features of the landscape; for this reason, perhaps, they have been considered with more attention than depositional landforms that, although interesting from a morphogenetic point of view, are usually less attractive.

The close links between erosional agency, accomplished process and produced landforms were recognized also by the ancients. Leonardo da Vinci at the end of the fifteenth century wrote: 'Every valley is created by its river and the proportion between one valley and another is the same as that between one river and another.' Once the concept that distinguishing features of landforms depend on the geomorphic process responsible for their development was fully understood, the genetic classification of landforms became possible. This scientific advancement made it possible to study the Earth's physical landscape not only from the descriptive point of view, but also considering the possible interpretation as to its geomorphological history. The genetic interpretation of erosional landforms, however, to be satisfactory must take into account the possible homologous or converging landforms, i.e. those landforms that although generated by different processes show similar features. Moreover it must be kept in mind that the genesis of most erosional landforms cannot be attributed to a single process, although it is rather simple to identify the dominant one.

The work of erosional agents produces peculiar assemblages of landforms that take on distinctive aspects depending on the stage of their development. The recognition that landforms change in time sequentially is the basis of the concept of the *Geographical Cycle* by Davis (1899). Once this concept had imposed itself, geomorphological interpretations made a new step forward. In fact, if properly applied, the geographical cycle affords a useful reference scheme to predict the possible future evolution of Earth's physical landscape (see CYCLE OF EROSION).

Erosion factors

Erosional landform features strictly depend not only on the way the exogenic agents operate, but also on a series of factors that control both the nature and the rate of erosion. The most important factors of erosion are lithology, tectonics, climate, vegetation and humans.

LITHOLOGY

Lithology strongly controls erosional processes as rock ERODIBILITY relies on it; as a consequence it influences the speed of erosional processes. In this perspective rocks are often referred to as 'hard' or 'resistant' or 'weak' and 'non-resistant' to erosional processes. The same erosional process can operate in a differentiated way where resistant rocks crop out next to no resistant rocks: as the erosional process proceeds, an uneven surface originates where more resistant rocks, slowly and hardly eroded, stand higher above less resistant rocks, which are more quickly and easily eroded. To some extent differential erosion can produce INVERTED RELIEF. The effects of differential erosion are particularly evident on stratified and differently erodible rocks. In this case the result of erosion is the formation of steep, abrupt faces of rock that mark the outcrop of the more resistant layers; the steep faces of a CUESTA, the rock terraces of a step-like slope or the scarp of a MESA are typical products of differential erosion. The concept of more or less erodible rocks is a relative one; in fact a rock can be resistant to one process and non-resistant to another. Therefore lithology has an influence also on the typology of the erosional processes.

TECTONICS

Tectonics influences erosional processes in different ways. Faults (see FAULT AND FAULT SCARP) and FOLDs can bring into contact rocks with different erodibility and then favour differential erosion. Furthermore they can directly influence the response of rocks to erosion, thus conditioning erosion rate. In fact rock erodibility depends not only on lithological characteristics but also on rock attitude (dip-slopes are less resistant than scarp-slopes) and on the degree of tectonic deformation (the higher the deformation of rocks, the higher their erodibility). Tectonic joints and faults can influence both the intensity of erosion and the location of the resulting landforms (see JOINTING). For example, fluvial erosion acts more powerfully where joints and faults create zones of weakness in the rocks than in other directions. As a consequence the orientation of stream valleys often follows closely the directions of these discontinuities. The sensitivity of tectonic discontinuities towards erosional processes may be so great that the morphological effects of differential erosion can help in the identification of discontinuities of small entity or affecting plastic lithologies (Belisario *et al.* 1999).

Tectonic uplift has also an important role in controlling the effectiveness of erosional processes. Uplift and erosion, together with relief, are the fundamental components of geomorphodynamic systems (see SYSTEMS IN GEOMORPHOLOGY) and are functionally related to one another in a negative feedback process (Ahnert 1998). When uplift prevails relief increases, and as a consequence erosion rate is faster. Increased erosion rate can eventually balance the building processes; in this case mountains do not change in elevation. When erosion overcomes the effects of uplift, elevations begin to lower and consequently the erosion rate slows down until the whole process comes to an end (Figure 58).

CLIMATE

Climate controls erosional processes both directly and indirectly. The direct control is exerted by the climate elements – temperature, rainfall and wind – that show a wide variability not only from one part to another of our planet, but also within very restricted areas, as for example from one slope to another on the same mountain. This wide variability of climatic conditions affects WEATHERING processes that weaken the rocks, predisposing them to subsequent erosional processes. Furthermore it favours the action of one erosional agent with respect to others: fluvial erosional processes become dominant in shaping the Earth's surface where rainfall amount is enough to guarantee the perennial channelled flow of waters, wind erosion is particularly effective where humidity is low, and glaciers can operate only where temperatures are such as to allow the fall and accumulation of snow. Besides this quite obvious influence, climatic conditions also control the way the different erosional processes operate with each other; these considerations are the basis of CLIMATIC GEOMORPHOLOGY which examines the systems of morphodynamic processes and their reciprocal interactions, in relation to the different climatic conditions.

Climate not only affects the typology of erosional processes but also the different behaviour of rocks. Under different climatic conditions the same rock may exhibit a different degree of resistance towards erosional processes and therefore it can be shaped into a variety of landforms. Granitic rocks are a good example. Depending on the climatic conditions and therefore on the dominant erosional processes they can be eroded into: the sharp peaks of the Monte Bianco massif, the low relief of INSELBERGs that dominate the savannah and steppe regions of South Africa, the ellipsoidal hollows (TAFONI) which originate from chemical corrosion and from the sweeping action of wind, the large round-topped mountains (piton) of tropical regions, etc. The close relationships among climatic conditions, erosional processes and landforms help to reconstruct the climatic variations that occurred in the past by

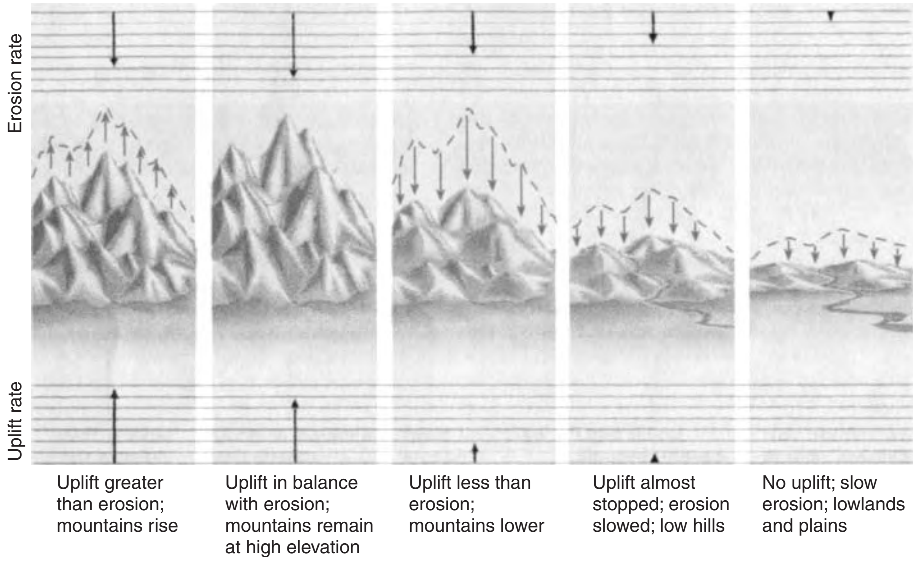

Figure 58 Negative feedback process relating uplift, erosion and mountain elevation. (From Press and Siever 1994: 364, modified)

examining the imprints left on the Earth's landscape by the prevailing erosional processes.

VEGETATION

The indirect influence of climate on erosion is largely related to the way it affects the amount and type of vegetation that, in its turn, has an important control on the EROSIVITY of some erosional agents. A thick vegetation cover inhibits surface runoff, thus restraining the action of running water; furthermore it obstructs the free flow of winds and therefore reduces the effectiveness of aeolian erosional processes. Root structures have a double influence: they can enhance the resistance of loose materials towards erosion or they can cause the breakage of solid bedrock, thus making erosion easier. As a whole, vegetation constrains erosional processes more frequently than it favours them. The limiting effect of vegetation on erosional processes varies as a function of the kind and density of the vegetation cover, and reveals itself because it enhances the stability of surface materials. Vegetation is also an important factor of pedogenesis, as it supplies the organic matter necessary for humus formation; therefore it has a role both in forming soil and in protecting it from erosion. This protective action is of primary importance to inhibit SOIL EROSION. When climatic conditions are such as to assure a dense and persisting vegetation cover erosional processes are slowed down: in these conditions (referred to as biorhexistasy) soils can develop and stay in place; on the contrary when climate is unfavourable to the development of vegetation, erosional processes are largely widespread and soils are easily removed (rhexistasy).

HUMAN IMPACT

If it is true that *humans* are nowadays powerful erosional agents, they are also an important factor of erosion. Anthropogenic activities are sometimes directed to undo or reduce erosional processes accomplished by 'natural' agents, as, for example, in the case of coastal defences built to inhibit sea erosion. More frequently, however, they produce the opposite effect and make erosion rate faster: in very densely inhabited areas, for instance, the extensive use of asphalt and concrete favour surface runoff and then erosion due to running water.

All the erosion-controlling factors play their role together; therefore their effects can interfere with each other in many possible ways so that the overall control on erosional processes is widely differentiated both in space and time. An example that clarifies the response of erosional processes to the complex constraints imposed by erosional factors is afforded by some more careful considerations about soil erosion. Once pedogenetic processes have led to formation of soils, they are exposed to the action of exogenetic erosional agents that start consuming them. When soil erosion proceeds normally, equilibrium conditions are attained: the rate at which soil is eroded equals the rate of soil formation. If this equilibrium is broken, erosional processes can become faster than pedogenetic processes; as a result, soil erosion is accelerated. The conditions more favourable to start *accelerated erosion* occur where weak rocks (such as clay or marls) crop out on areas affected by abundant and irregular precipitation that favours erosion by running water; under these conditions, for instance, BADLANDS originate. Wherever these natural predisposing conditions are added to deforestation and faulty land use connected to anthropogenic activities, accelerated erosion attains its maximum intensity. Under these conditions the soil erosion rate exceeds the soil formation rate. As a result soils get thinner and can completely disappear. In some cases erosion becomes so severe that it can be compared to the process of DESERTIFICATION: the irreversible process whereby soils lose their fertility because of the destructive effects of some anthropogenic activities.

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SEE ALSO: fluvial geomorphology; glacial erosion; granite geomorphology; tectonic geomorphology; wind erosion of soil

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EROSIVITY

A measure of the potential ability of a soil to be eroded by particular geomorphological processes. Erosion is a function of erosivity on the one hand and of erodibiltiy (the vulnerability of a soil to erosion) on the other.

Water erosion susceptibility is related to various rainfall erosivity indices. Rainfall intensity, rainfall amount and antecedent conditions are all important controls of erosivity, but as Morgan (1995: 27) has remarked: 'the most suitable expression of the erosivity of rainfall is an index based on the kinetic energy of the rain. Thus the erosivity of a rain storm is a function of its intensity and duration, and of the mass, diameter and velocity of the raindrops'. In recent years RAINFALL SIMULATION has been used to assess the response of soils to storms with different characteristics.

Wind erosion susceptibility has also often been determined using indices based on wind velocities and durations above certain threshold velocities (e.g. Skidmore and Woodruff 1968) and portable wind tunnels have been employed to assess the response of different ground surfaces to different wind velocities.

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A.S. GOUDIE

ERRATIC

Erratics are rock fragments carried by a glacier, or in some cases by floating ice, and subsequently deposited at some distance from the outcrop from which they were derived. For this reason their lithology differs from the surrounding rocks and sediments – hence the term 'erratic'. Some erratics are large blocks, that lie free on the surface and form interesting landscape features. Glaciologists, however, mainly use the term for the exotic components embedded in tills (see GLACIAL DEPOSITION), encompassing both large clasts and fine-grained rock fragments.

Scientific investigation of erratics started during the first half of the nineteenth century, when most geologists thought that they were swept into the northlands by the universal flood. At the same time their transportation and distribution by former widespread glaciers was first suggested, and later in the nineteenth century it was universally agreed.

Some erratics form landmarks because of their spectacular dimensions. One of the largest erratics measures 45 m by 20 m by 10 m and is estimated to weigh 16,500 tons. It is part of the Foothills erratic train, a series of boulders stretching over 400 km along the eastern foothills of the Rocky Mountains.

Erratics not only give evidence of the existence of former glaciers; especially erratics in tills provide a powerful tool for many glacial investigations. Such studies are based on the identification of 'indicator erratics'. Indicator erratics are those for which a definite source area is known. They form 'indicator trains' or, in cases of shifting ice divides and ice flow directions within an ice sheet, 'indicator fans' trailing downglacier from the source rock. Indicator trains and fans are enriched in the distinctive component relative to the till underlying or enclosing it. The concentrations of indicator erratics vary systematically along former ice flowlines. Within indicator outcrops, concentrations increase rapidly downglacier, reflecting the addition of new material from the glacier bed, but concentrations drop off rapidly down-ice of the outcrop margin. The up-ice and down-ice limits of an indicator plume are known as the 'head' and the 'tail', respectively.

Erratics can be used to reconstruct the pattern and history of ice flow in studies of ice sheet dynamics as long as erratic transport histories are not blurred by repeated glaciations involving total redistribution of previously deposited materials (Benn and Evans 1998). The study of erratic dispersal patterns furthermore can provide important clues to the location of mineral outcrops or ore bodies, because the erratic plumes are much larger than their bedrock sources, making them easier targets to find (Kujansuu and Saarnisto 1990). In Denmark, Germany and the Netherlands till units of different age show differently composed indicator assemblages and counts of the erratics derived from various western, central or eastern Scandinavian source areas are here successfully employed in stratigraphical investigations (Ehlers 1996).

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ESCARPMENT

The term escarpment, or scarp, has been applied traditionally to a steep, often single slope, of considerable length, that dominates a section of landscape. An escarpment thus can be distinguished from the two flanking walls of canyons. For example, south of Sydney, Australia, the coastal escarpment forms a long, virtually continuous wall, but it is outflanked by canyons which extend more than 100 km further inland. Another notable instance of a valley cut well back from an escarpment is the Sognefjord, which extends about 200 km inland from the coastal edge of the Norwegian highlands. The lengths of escarpments vary from a few kilometres to the subcontinental scale of mega-escarpments, or Great Escarpments, such as the Drakensberg of South Africa, while their heights vary from a few tens of metres to several thousand metres. A distinction is generally drawn between denudational escarpments and fault scarps (see FAULT AND FAULT SCARP), although this may be no simple exercise in areas of essentially homogeneous crystalline rocks.

The majority of denudational escarpments have formed as a result of differential rock resistance to erosion. Probably the most outstanding examples form the sequence of the Vermillion, White, Grey and Pink Cliffs, known as Great Staircase, which rises from the rim of the Grand Canyon of the Colorado River to the high plateaux of southern Utah. The treads of this staircase are cut mainly in softer strata between the sandstones which form the cliffs. Major sandstone escarpments also occur in the Adrar and Borkou areas of the Sahara. But such features are not limited to arid lands, for the great ramparts of the Roraima massif have developed in the humid tropics of Venezuela. Neither are they limited to sequences of sedimentary rocks. Major escarpments and benches also have developed in response to the differential resistance of volcanic strata in the Bushveld of Transvaal, and of sheeted granites intruded through metamorphic rocks in Madagascar. Some escarpments in granitic rocks were initiated at the boundary with less resistant rocks, and have retreated to their present positions. Others, such as the multiple steps in the Sierra Nevada of California, were initiated by differential weathering within a granitic mass.

While many escarpments can be attributed to the great resistance to erosion of particular types of rocks, others show no systematic relationship to lithology. For example, the coastal escarpment south of Sydney extends from sedimentary to metamorphic rocks, and thence to crystalline rocks. Escarpments of this type have developed in response to regional uplift. The most extensive of them, which are sometimes called Great Escarpments, occur on many tectonically PASSIVE MARGINs of continents. Notable examples are the Drakensberg, the Western Ghats of India, the escarpments of east Australia, the Serra da Mantiqueira of Brazil, the coastal escarpment of Norway, the escarpment of east Greenland and the Torngat Mountains of the uplifted margin of Labrador.

Great Escarpments are not limited to passive margins, for they also occur on tectonically active continental margins. Collision of the Australian and Pacific Plates has resulted in some 20 km of uplift in about the last 10 million years along the west coast of the South Island of New Zealand. And, although uplift has been virtually matched by erosion, the flank of the Southern Alps rises in a steep, heavily dissected wall from the narrow coastal plain. Likewise, collision of the Pacific, North American and Cocos Plates has resulted in major uplift that, together with rapid erosion, has produced the great escarpments of the Sierra Madre Occidental and Sierra Madre Del Sur on the western flank of the Mexican highlands.

Less extensive, though nonetheless impressive, escarpments have developed as a result of regional uplift in continental interiors. The classic examples are along the margins of the *Mittelgebirge* of central Europe, such as the Massif Centrale of France and the Erzgebirge of Germany. However, even more impressive escarpments occur along the margins of uplands in central Asia, as for example on the northern side of the Bogda Shan in north-west China. Erosion in response to the regional uplift in north-east Africa has resulted in high escarpments cut largely in volcanic rocks along the west flank of the Ethiopian plateau. The Kaibab Limestone escarpment of the Mogollon Rim on the southern flank of the Colorado Plateau is a notable North American example.

Much of the initial research on escarpments was carried out in the folded terrain of the Appalachians and north-west Europe, where they occur in association with homoclinal CUESTAs and HOGBACKs. As early as 1895 W.M. Davis pointed out escarpments in this type of terrain were second-stage features that did not develop until streams began to extend headwards along the strike of the folds. He attributed the retreat of escarpments not only to erosion on their steep faces, but also to lateral erosion of streams at the foot of escarpments. Although Davis's ideas, and especially his terminology, have been subject to much criticism (see SLOPE, EVOLUTION), the scheme that he proposed a century ago still provides the basis for the interpretation of many scarp and cuesta landscapes. However, major challenges to it have come from German geomorphologists.

Schmitthenner (1920) argued that the most important form of denudation in scarplands (*Schichtstufenlandschaft*) is the lowering of the backslope surface and the breaching of escarpments from the rear. He attributed this primarily to seepage down the dip, and the consequent development of swampy hollows, or *dellen*, by solutional processes. Strong support for the role of seepage leading to the lowering of escarpment crests and to the breaching of them from the rear has come from recent research on the Colorado Plateau and Australia.

Many stairways of multiple escarpments have been attributed to repeated uplift and erosion, but, as independent evidence of repeated uplift is often lacking, alternate hypotheses need consideration. W. Penck (1924) claimed that multiple scarps and benches (*Piedmonttreppen*) could form on a continuously rising and expanding dome. Penck's hypothesis provides a valuable warning that the relationship between tectonics and slope form may be complex (see SLOPE, EVOLUTION).

In recent decades prominent German researchers have expounded climatic explanations of escarpments (see MORPHOGENETIC REGION). Büdel (1982) argued that escarpments in the tropics are essentially the result of deep weathering and the subsequent stripping of regolith, and referred to them as 'etchplain stairways'. He also extended this climatic interpretation to temperate lands by claiming that most of the escarpments of south-west Germany were formed in a similar fashion to etchplain stairways under a past 'tropicoid climate'. The structural influences so clearly expressed in many of these escarpments were dismissed as only an 'arabesque' in the general two-stage development by deep weathering and subsequent stripping. According to this theory, scarplands are sculpted predominantly by areal downwearing, and scarp retreat is minimal.

In striking contrast, conclusions drawn from studies in southern Africa, especially those of L.C. King, emphasize the dominant role of the retreat of escarpments over long distances. According to King (1953), the most important processes are sheet wash on pediments at the foot of scarps, and mass failure and gully erosion on the steep slopes. He argued that 'scarps retreat virtually as fast as nick-point advance up rivers, so that the distribution of successive erosion cycles bears no relationship to the drainage pattern whatsoever'. He argued also that retreating scarps resulted in isostatic (see ISOSTASY) uplift, generally in the form of large-scale warping, initiating a new cycle of scarp retreat.

Although many escarpments may have retreated over considerable distances, some have apparently not done so. The Blue Ridge Escarpment on the east flank of the Appalachian Mountains is a major feature, which lies about 350 km inland from the coast. Rather than having retreated from the coast, however, this escarpment may have been maintained in approximately its present position over many millions of years by uplift along an ancient continental margin preserved deep in the crust. It thus seems to be in a long-term state of dynamic equilibrium in which denudation has been largely balanced by the slow rise of the underlying rocks. On the other hand, some escarpments are essentially fossil features that have been exhumed by erosion (Plate 41). Sedimentary evidence indicates that the Arnhemland Escarpment, cut in Proterozoic rocks in the Northern Territory of Australia, is a coastal cliff-line that was buried during Cretaceous times and subsequently exhumed. Since being exhumed, parts of this escarpment have been almost stationary, and the most active parts have retreated only a few kilometres.

Assessing which explanation is best suited to a particular escarpment may be no simple task, and indeed none of them may be entirely satisfactory. The final appeal must be to field evidence rather than to conceptual constructs.

The rates at which escarpments retreat vary greatly, and seem to depend on lithological resistance, tectonic activity and the intensity of erosion which itself is largely controlled by climate. Many major escarpments are thought to retreat about 1 km per million years. However, the distance of retreat from dated basalts show that escarpments in temperate east Australia have retreated at rates of only about 25 to 170 m per million years, and the rates for some of them have been even slower. Low rates of only about 70 m to 150 m per million years have also been recorded in subarctic Spitzbergen. Average rates of scarp retreat on the Colorado Plateau since Miocene times have been only about 160 m per million years, but earlier than that the rates were about 1.5 km to 4 km per million years. The great reduction has been attributed to the incision of streams and thus the cessation of very active lateral planation at the foot of scarps. Denudation

Plate 41 Proterozoic sandstone capping schists on the Arnhemland Escarpment, Northern Territory, Australia

on the Arnhemland Escarpment increased by an order of magnitude during the late Quaternary, but in this case the change seems to have been climatically controlled. Moreover, the evidence from Arnhemland prompts caution in extrapolating long-term rates of retreat from relatively short-term records.

Although most research has been carried out on escarpments above sea level, submarine escarpments are of far greater magnitude. The fall below the continental shelf off east Australia is more than five times greater than the height of the escarpment onshore, and contrasts of similar, or greater, magnitude occur along most continental margins.

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R.W. YOUNG

ESKER

Derived from the Irish word *eiscir* (ridge), an esker is an elongate sinuous ridge composed of glacifluvial sediments and marking the former position of a subglacial, englacial or supraglacial stream. The routing of former meltwater channels in glaciers, and their association with ice-marginal landforms and sediments is indicated by the overall form of eskers. There are four major types of esker (Warren and Ashley 1994: (1) continuous ridges (single or multiple) that document tunnel fills; (2) ice-channel fills produced by the infilling of supraglacial channels; (3) segmented ridges deposited in tunnels during pulsed glacier recession; and (4) beaded eskers consisting of successive subaqueous fans deposited in ice-contact lakes during pulsed glacier recession. In planform, eskers also come in a wide variety of types. These include, single ridges of uniform cross section or of variable volume, single ridges linking numerous mounds (beads) and complex braided or anabranched systems. Although individual eskers or esker networks may stretch over hundreds of kilometres of former ice sheet beds it is unlikely that they formed in tunnel systems of that length. Rather, they were probably deposited in segments in the ice sheet marginal zone of ablation and each segment was added as the ablation zone migrated towards the ice sheet centre. In some locations eskers lie on the bottoms of bedrock meltwater channels (TUNNEL VALLEYs or Nye channels) indicating that erosion by subglacial meltwater was followed by deposition, possibly due to waning discharges, and that subglacial conduits were remarkably stable features.

Most eskers are aligned sub-parallel to glacier flow, reflecting the flow of meltwater towards the glacier margin. Eskers that were deposited as subglacial tunnel fills may be the result of flow in pressurized conduits and therefore may possess up-and-down long profiles where they climb over topographic obstacles. This is due to the fact that flow in the conduits was driven by the ice surface slope rather than by the glacier bed topography (Shreve 1972). If conduits or tunnels switch to atmospheric pressure, as occurs beneath the thinner ice nearer to the glacier snout, then the water will follow the local bed slope and so any resulting eskers will be deposited transverse to glacier flow (*valley eskers*).

The former englacial position of some eskers is indicated by the occurrence of buried glacier ice or almost complete disturbance of the stratified core. Englacial or supraglacial construction of eskers will result in the draping of the features over former subglacial topography after glacier melting. These apparent up-and-down long profiles must not be confused with true subglacial eskers deposited under pressure and therefore also crossing topographic high points. Largely intact internal stratigraphies are typical of subglacial eskers.

Eskers are often well stratified but contain a variety of sediment facies. Particles are usually not far-travelled, most being no further than 15 km from their source outcrop. The wide range of BEDFORMS observed in esker sediments reflect and document the large variations in meltwater discharge on both diurnal and seasonal timescales. Rhythmicity or cyclicity in the sediments is often represented by fining-upwards sequences separated by erosional contacts. Each fining-upward unit records an individual discharge event of maybe only hours in duration. The occurrence of large cross-bedded sequences may document deposition in deltas in subglacial ponds. Where tunnels collapse and/or change shape or streams change size or position, one depositional sequence may be truncated or partially infilled by another. Apparent anticlinal bedding in some eskers is thought to be the result of sediment slumping down the esker flanks as the supporting ice walls melt back.

The segments or beads on eskers are interpreted as the products of ice marginal deposition at the mouths of subglacial tunnels. At each tunnel mouth the glacifluvial sediment being carried through the tunnel or conduit is deposited in SUBAERIAL or subaqueous fans/deltas due to the sudden drop in meltwater velocity as it leaves the confines of its ice walls. Beads may also accumulate in subglacial cavities that develop as offshoots to the main tunnel. Where well-integrated esker networks carry large volumes of debris to the glacier margin they may link up with extensive ice-marginal depositional systems that include ice-contact deltas, subaqueous fans and MORAINES.

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DAVID J.A. EVANS

ESTUARY

Estuaries are unique ecosystems that provide spawning grounds for many organisms, feeding stops for migratory birds and natural filters to maintain water quality. Estuaries have value to humans for shipping and boating, settlements, erosion protection, recreation, mineral extraction and release of waste materials. Estuaries are generally considered areas where salt water from the ocean mixes with freshwater from land drainage but there are many definitions for the term (see Perillo 1995) reflecting the complex physical and biological processes present. Estuaries may be classified or described on the basis of numerous criteria, including entrance conditions (Cooper 2001), stage of development and degree of infilling (Roy 1984), morphology (Pritchard 1967; Fairbridge 1980), tidal range (Hayes 1975), vertical stratification and salinity structure (Cameron and Pritchard 1963). All these criteria affect the evolution of estuaries and the nature of transport of sediment and biota. The most common definition is Cameron and Pritchard (1963: 306) who define an estuary as ‘a semi-enclosed coastal body of water having a free connection with the open sea and within which sea-water is measurably diluted with fresh water derived from land drainage’. The boundaries of an estuary can be defined by salinity (ranging from 0.1‰ at the head of the estuary and 30–35‰ at the mouth) or sedimentary facies and the processes that shape them. For example, Dalrymple *et al.* (1992) defined the upper boundary as the landward limit of tidal facies and the lower boundary as the seaward limit of marine facies.

From a geologic perspective, today’s estuaries are recent features. Estuaries are a product of inherited factors (i.e. lithology) that influence the configuration of the estuarine basin and sediment type and availability; broad-scale controls such as climate and sea-level rise that influence rates of discharge and inundation; and, contemporary processes (wave, tide and river) that influence hydrodynamics and sediment transport. The position of estuaries is a result of fluctuations in sea-level rise, with sea-level elevation at or above current levels during interglacial periods and up to 150 m below present levels during glacial periods. Present-day estuaries are the result of sea-level rise and inundation of coastal lowlands following the last glacial period and the sea-level stillstand that began approximately 6,000 years ago. More recent regional sea-level histories have revealed both lowering and rising of sea level from stillstand levels.

Classification

Estuaries can be broadly classified as drowned river valleys, fjords, bar-built, and those formed by faulting or local subsidence (Pritchard 1967; Fairbridge 1980). Drowned river valley estuaries are found along the east coast of the United States (i.e. Delaware Bay and Chesapeake Bay) and in England (i.e. Thames and Mersey estuaries), France (i.e. Seine) and in Australia (i.e. Batesman Bay). Rivers eroded deep V-shaped valleys during the last glacial period that were subsequently inundated when melting ice sheets caused a rise in sea level. The planform and cross section of these estuaries are often triangular or funnel-shaped. In systems where sedimentation rates are less than rates of sea-level rise the river valley topography is maintained. Bar-built estuaries have a geologic history similar to drowned river valleys (the result of glacial incision and subsequent inundation by sea-level rise) but recent marine sediment transport (alongshore or cross shore) results in the creation of a barrier or spit across the mouth. The inlet at the mouth is small relative to the size of the shallow estuary created behind the barrier. In some cases the barrier may restrict exchange of water between the ocean and estuary except during high tides. Examples of this type of estuary can be found in the United States (i.e. Mobile Bay and Pamlico Sound) and in Australia (i.e. Clarence and Narooma Estuaries). Fjords are glacially-scoured U-shaped valleys that were subsequently inundated by a rising sea level. Most fjords possess a shallow rock sill near the mouth that forms an estuarine basin. Fjord-type estuaries are found in upper latitudes (i.e Oslo Fjord, Norway and Puget Sound, USA). Some estuaries are formed in valleys that were created by processes such as faulting (i.e. San Francisco Bay, USA) or subsidence.

Estuaries are located in micro-, meso- and macro-tidal environments. Planform morphology is an important control on the variation of tidal range and the magnitude of the tidal current within an estuary (Nichols and Biggs 1985). Convergence of the estuarine sides causes the tidal wave to compress laterally. In the absence of bed friction, the tidal range will increase. In the presence of friction, the tidal range will decrease. The relationship between convergence and friction control the amplitude of the tide within the estuary. In cases where convergence is greater than friction the tidal range and strength of the tide will increase toward the head of the estuary (hypersynchronous estuaries). In cases where convergence is less than friction the tidal range will decrease throughout the estuary (hyposynchronous estuaries).

Morphology

Wave, tide and river processes control the location of marine and river sediments in the estuary and the morphology of the sedimentary deposits. Conceptual models of estuarine morphology classify estuaries according to the relative contribution of these processes (see Dalrymple *et al.* 1992; Cooper, 1993) and are based, in part, on regional studies of estuarine sedimentation and morphology. These studies include tide-dominated, macro-tidal estuaries (Dalrymple *et al.* 1990) and micro-tidal estuaries in wave-dominated (Roy 1984) and river-dominated (Cooper 1993) environments.

Tide-dominated estuaries are found in macro-tidal environments (tidal range >4 m). They are generally funnel-shaped with wide mouths and high current velocities. Dalrymple *et al.* (1990) characterized the sedimentary characteristics of the macro-tidal Cobequid Bay–Salmon River Estuary, Canada. The axial sands are characterized by the presence of elongate tidal sand bars in the lower sector of the estuary that trend parallel to the dominant current direction. Sand flats and braided channels are located in the middle sector of the basin and a single channel is located in the river-dominated head of the estuary. Tidal currents are at a maximum in the inner part of the estuary. Sediments decrease in size from the mouth to the head. Dominant direction of sediment transport is landward with accumulation in the upper sector at the head of the estuary.

Wave-dominated estuaries are generally found in micro-tidal (tidal range <2 m) environments (Roy 1984). In general, these estuaries have an upper sector near the head, where river processes, sediments and bedforms dominate, a lower sector near the mouth, where wave and tidal processes and marine sediments dominate, and a middle sector, where tidal currents dominate and both river and marine sediments are present. High wave and tidal energies at the mouth of an estuary can deposit sediment and restrict or completely prohibit exchange of water between the ocean and the estuary.

Mixed wave-tide estuaries (such as those in meso-tidal environments with a tidal range of 2–4 m) can be found behind barrier islands (Hayes 1975). The dominant sand bodies in meso-tidal estuaries are the deltas (ebb and flood) formed by tidal inlet processes. Within the estuary are meandering tidal channels and point bars and marsh deposits.

River-dominated estuaries do not display the characteristic downstream facies changes observed in wave- and tide-dominated estuaries, and energy levels may remain similar along the axis of the river valley (Cooper 1993). River-dominated estuaries can range from those completely dominated by river processes (river channels) to those that experience some marine inputs at the mouth.

Shoreline environments

Estuarine shoreline environments often occur in small isolated reaches with different orientations and with great variability in morphology, vegetation and rate of erosion. This variability results from regional differences in fetch characteristics, exposure to dominant and prevailing winds, variations in subsurface stratigraphy, irregular topography inherited from drainage systems, differential erosion of vegetation or clay, peat and marsh outcrops on the surface of the subtidal and intertidal zones, small-scale variations in submergence rates, effects of varying amounts of sediment in eroding formations and effects of obstacles to longshore sediment transport, such as headlands and coves, that define drift compartments (Nordstrom 1992). Differences in the gradient of wave energy between the low-energy (upper) and high-energy (lower) shorelines in an estuary and between the high-energy (windward) and low-energy (leeward) sides of an estuary also contribute to differences in the types of estuarine environments and their dimensions. Saltmarsh is likely to form on alluvium in the upper reaches of the estuary, on the upwind side of the estuary or on the downwind side of the estuary in the lee of headlands that provide protection from breaking waves. Beaches are likely to form on the downwind side of estuaries because there is sufficient energy in the locally generated waves to erode coastal formations or prevent plants from growing in the intertidal zone.

BEACHES may be unvegetated or partially vegetated and are composed of sand, gravel or shell. The dominant processes of sediment reworking on beaches in estuaries is usually locally generated waves, although refracted and diffracted ocean waves may be present. The best development of beaches occurs where relatively high wave energies have exposed abundant unconsolidated sand or gravel in the eroding coastal formations. Adequate source materials occur where these formations are moraine deposits, submerged glacial streams, coarse-grained fluvial deposits, and sand delivered by ocean waves and winds, such as the estuarine shorelines of spits and barrier islands. Beach formation is favoured where high ground protrudes into relatively deep water, where wave refraction and wave energy loss through dissipation on the bay bottom is minimal. Ocean waves that enter the estuary usually create beaches close to the inlets. Sediment transported into the estuary by ocean waves may form spits in the lee of headlands in the estuary. Beaches created by waves generated within estuaries are most common in shoreline re-entrants, where sediments can accumulate over time. Other beaches occur where sand is plentiful on the bayside of barriers enclosing the estuary, particularly on former recurves, subaerial overwash platforms and former oceanside dunes (Nordstrom 1992).

Beaches may form on the bay side of eroding marshes from coarse-grained sediment removed from the eroding substrate. Beaches may precede and favour marsh growth by creating spits that form low-energy environments landward of them. Both processes create a beach ridge shoreline that combines features characteristic of beach shorelines and marsh shorelines. Peat, representing the substrate of former marsh, is often exposed in outcrops on eroding beaches transgressing marshes. These outcrops are resistant because of the presence of fine-grained materials trapped by upward growth of the marsh and the binding effect of vegetation.

Dunes (see DUNE, COASTAL) form within estuaries only where beaches are sufficiently wide to provide a viable sediment source or where the shoreline is stable enough to allow ample time for slow accretion or to prevent wave erosion. Wave energy must be sufficient to prevent colonization of intertidal vegetation, but erosion cannot be too great for aeolian forms to survive. Onshore aeolian transport occurring between moderate-intensity storms may create only a thin aeolian cap on top of the backbeach or overwash platform.

Marshes are components of the intertidal profile affected by waves and tidal flows, and they bear many similarities with beaches, including the potential for cyclic exchange of sediment between the upper and lower parts of the profile and a tendency to buffer energy in a way that resists long-term morphological change (Pethick 1992). Marsh shorelines differ from beaches in that they are characterized by fine sediment sizes, low gradients and dissipative slopes. The occurrence of marshes, like beaches, depends on their environmental setting and mode of origin, defined by factors such as bedrock geology, availability of sediments and recent sea-level rise history (Wood *et al.* 1989). Examples of distinctive morphological marsh units determined by macro-scale differences within estuaries include fluvial marshes, occurring on the upper estuarine margins of rivers; bluff-toe marshes that form at the base of coastal bluffs; backbarrier marshes found behind barrier islands and spits; and transitional marshes where freshwater peatlands are colonized by saltmarsh (Wood *et al.* 1989).

Competition for human resource values of the estuarine shoreline has led to elimination of many natural environments. The conversion of some of the environments (especially bay bottoms and marshes) is now severely restricted by land use controls in many countries, but many estuaries are still threatened by human activities.

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N.L. JACKSON

ETCHING, ETCHPLAIN AND ETCHPLANATION

The word 'etching' generally means corroding a surface by aggressive reagents and is used in geomorphology to describe progressive rock decomposition which occurs within deep weathering profiles. In particular, it is applied to situations where rocks differ in their resistance to chemical decay and consequently thickness of a weathering mantle is highly variable over short distances. Removal of products of deep weathering will expose bedrock surface, the topography of which is the direct result of etching, thus it is an 'etched surface'. At an early stage of development of geomorphology, when focus on planation surfaces and peneplains was pre-eminent, etched surfaces were visualized as surfaces of rather low relief and thought of as a special category of a PENEPLAIN, produced by deep-reaching rock decay followed by stripping of weathering mantle. For surfaces of this origin, the term 'etchplain' was proposed by B. Willis and E.J. Wayland, working in East Africa in the 1930s. Accordingly, the process of producing an etchplain through weathering and stripping has later become known as 'etchplanation'.

The impact of the concept of etching and etchplanation on general geomorphology was initially limited, mainly due to the association of etchplains with peneplains, remoteness of original study areas, and restricted access to early publications. Furthermore, no applications for extra-tropical areas were offered and etchplains were considered as of local importance and specific for low latitudes. Proposals to restrict the usage of the term 'etchplain' to areas of exposed rock, i.e. completely stripped of weathering products, added to the slow progress and general downplaying of the significance of etchplains in geomorphology.

The situation began to change with the arrival of the paper by Büdel (1957) which is noteworthy for a number of reasons, although the very term 'etchplain' was not used. First, Büdel made clear that he was applying the weathering/stripping concept to entire landscapes rather than to limited areas within them or to individual landforms. Second, he suggested that many upland surfaces in middle and high latitudes are inherited Tertiary etchplains, and hence extended the applicability of the concept outside the tropics. Third, he pointed out that transition from the phase of dominant weathering to the phase of dominant stripping might be associated with major environmental changes, whose profound impact for landform development was realized only later. Fourth, it contributed to the appreciation of the concept by the Central European geomorphological community, which was reflected in numerous detailed studies shortly after.

Realization of the crucial role of deep weathering and SAPROLITE development in shaping most tropical landscapes, achieved in the 1960s, led to the expansion of the original ideas of Wayland and Willis, so different types of landscapes could be described. The proposed classification, in the form subsequently modified by its author himself (Thomas 1989), include:

- *Mantled etchplain*: weathering mantle is ubiquitous and virtually no bedrock is exposed. Weathering progressively attacks solid rock at the base of the mantle, moulding the etched surface which is to be exposed later, but the mantle can also be relict.
- *Partly stripped etchplain*: develops from mantled etchplain through selective removal of the weathering mantle and exposure of bedrock surface, but part of the original saprolite remains. The proportion of areas still underlain by saprolite may vary from 10 to almost 100 per cent.
- *Stripped etchplain*: most of the bedrock is exposed from beneath a weathering mantle and only isolated patches of saprolite are left (<10 per cent of the area). These characteristics conform with the original definition by Wayland.
- *Complex etchplain*: includes a few variants, in which deeply incised valleys may be present (incised etchplain), or removal of saprolite is accomplished by pedimentation (pedimented etchplain), or a new generation of weathering mantles begins to form (re-weathered etchplain).
- *Buried etchplain*: one which has been covered by younger sediments or lava flows.
- *Exhumed etchplain*: one which has been re-exposed after burial.

One important terminological problem has been noted, that a stripped surface is rarely a plain but tends to show some relief, which reflects differential progress of etching (see Figure 59). This happens in particular, if bedrock is lithologically varied or various structures, e.g. fractures, are differentially exploited by weathering. In many granite areas, stripped surfaces are typified by domes, TORs, basins and boulder piles, and to call them 'etchplains' would be inappropriate and misleading. Therefore the term 'etchsurface' is recommended for use wherever evacuation of weathering mantles reveals a varied topography.

Etchplanation, and in particular the transition from the weathering phase to the stripping phase, is commonly linked with major external changes experienced by a landscape, related either to a change in tectonic regime or to environmental changes. It is envisaged that mantled etchplains form and exist during long periods (up to 10^9 yr) of stability, whereas stripping is initiated by uplift, or climatic change towards drier conditions, and is accomplished over much shorter timescales (10^5–10^7 yr). In this view, major external disturbances are essential for the formation of etchplains and static nature of planate landsurfaces might be implied. This position is contradicted by field evidence of geomorphic activity, hence an idea of 'dynamic etchplanation' has been introduced to emphasize ongoing landscape development through etching and stripping (Thomas and Thorp 1985). Key points made are simultaneous weathering and removal of its products, lowering of both interfluves and valley floors, continuous sediment transfer, redistribution and temporal storage of weathering products, and the importance of minor environmental disturbances.

From the 1980s onwards, following the progress in weathering studies, the etching concept has been extended away from tropical plains to the much wider range of settings. Emphasis on the process of deep weathering rather than on the ultimate form of a plain has made it possible to see geomorphic development of many low-latitude mountain ranges of moderate relief as being accomplished by differential etching. Deep weathering is facilitated by strong groundwater movement, steep hydraulic gradient, tensional fracture patterns and numerous lines of weakness within bedrock, whereas landslides play a major part in removal of the saprolite. Realization that

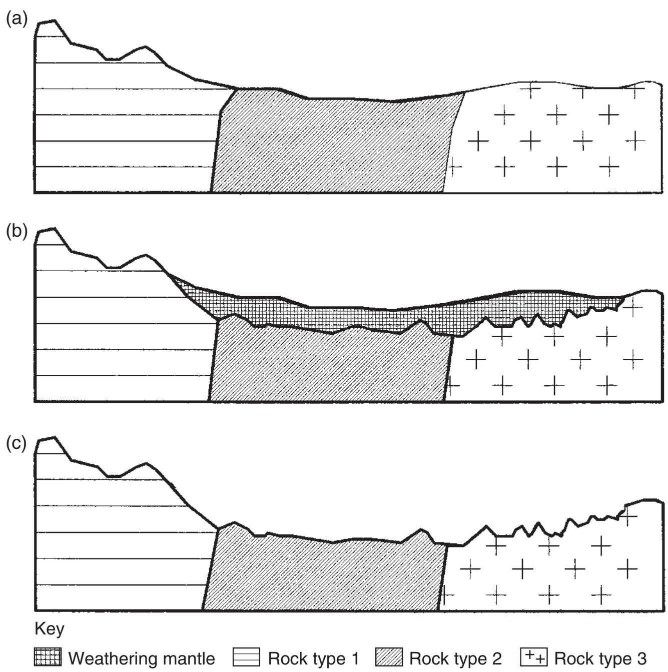

Figure 59 Depending on (a) bedrock characteristics and their susceptibility to (b) selective deep weathering, etching may produce surfaces of (c) various types, for instance inselberg-dotted plains (middle) or multi-convex, hilly areas (right)

formation of thick sandy weathering mantles (see GRUS) can effectively take place outside the tropics has opened the way to interpret several middle to high latitude terrains, with no or extremely remote history of tropical conditions, as etchsurfaces or etchplains, even if the very term has not always been used (Pavich 1989; Lidmar-Bergström 1995; Migoń and Lidmar-Bergström 2001).

Over the years, the idea of etching and etchplanation has evolved from being a mere specific, 'tropical' variant of peneplanation to the status of an autonomic concept, capable of accounting for both planate and topographically complex landsurfaces, integrating tectonic and climatic controls, linking historical and process geomorphology. Despite what the name and early history suggest, it must not be regarded as focusing on explaining the origin of planation surfaces. Nor does it compete with other planation theories, as for instance pedimentation may be a means of stripping. To the contrary, long-term etching and stripping may, and in many places do, lead to the differentiation and increase of relief. Depending on local lithology, tectonic setting and environmental history, long-term etching may transform an initial landscape into a range of topographies, from plains to even mountainous. Therefore, it is the evidence of former or present deep weathering which is a prerequisite for a terrain to be identified as an etchsurface, and not any particular assemblage of landforms.

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SEE ALSO: granite geomorphology; inselberg; planation surface; tropical geomorphology

PIOTR MIGOŃ

EUSTASY

The concept of 'eustatic' changes in sea level, implying vertical displacements of the ocean surface occurring uniformly throughout the world, was introduced by Suess (1885–1909). Global changes in sea level depend in fact on a combination of factors (changes in the quantity of oceanic water, deformation of the shape of the oceanic basin, variations in water density, and dynamic changes affecting water masses) which operate globally, regionally or locally on different timescales.

The quantity of ocean water is controlled mainly by climate, that may cause the development or melting of huge continental ice sheets. According to IPCC (2001), the present volume of continental ice can be estimated at about 29 million cubic kilometres, equivalent to *c.*70 metres water depth in the oceans. At the time of the last glacial maximum, some 20 ka ago, when the global sea level was estimated about 120 metres lower than now, the continental-ice volume must have been more than double the present.

Daly (1934) stressed the importance of changes in sea level and of glaci-isostatic effects (see ISOSTASY) accompanying the last deglaciation phase, with uplift in areas of ice melting and subsidence in a wide peripheral belt. During the last decades, improved global isostatic models based on ice volumes and water depths have been developed (e.g. Lambeck 1993; Peltier 1994), demonstrating that ice-volume changes imply vertical deformation of Earth crust which is highly variable regionally.

Mörner (1976) supported anew the old notion of geoidal changes, suggesting that displacements of the bumps and depressions revealed by satellites on the ocean surface topography could cause differences between coastal areas in the relative sea-level history.

Recently, the analysis of satellite observations, especially by Topex/Poseidon, have shown that the level of the ocean surface can be closely correlated with sea-surface temperature. The resulting sea-surface topography is highly variable, with sea-level rise in certain areas, and sea-level fall in other areas. Steric effects, which depend on the temperature (and density) of the whole water column are also highly variable. Analysis of the dynamic behaviour of water masses and their displacements would bring similar results.

SEA-LEVEL changes are therefore not uniform, but variable over several temporal and spatial scales. Sea level may therefore change from place to place in the ocean, and even more in coastal areas, where hydro-isostatic movements are controlled by the water depth on the continental shelf. In short, there is now general agreement that no coastal area exists where the local sea-level history could be representative of the global eustatic situation. Worldwide or simultaneous sea-level changes, therefore, do not exist. They are an abstraction.

In spite of this field evidence, the concept of eustasy is not an obsolete one, because the estimation of global sea-level changes, even if obtained with some approximation, may have many useful applications in geosciences, in relation to climate, tectonics, paleo-environmental, and also near-future environmental changes. If eustatic variations cannot be specified from coastal field data, the estimation of the changes in the quantity of ocean water is possible using geochemical analysis of marine sediments. The $\delta^{18}O$ content in fossil foraminifera shells cored from the deep ocean floor depends on the salinity and temperature of sea water at the time they lived. If benthic species are selected, temperature changes will be minimized and $\delta^{18}O$ will depend mainly on salinity, i.e. on the quantity of fresh water held up in continental ice sheets. Such calculation makes the estimation of approximate eustatic changes possible, with assumptions, and with an accuracy that depends on the resolution of geochemical measurements, i.e. with an uncertainty range, for sea level, of the order of ± 10 m. Such precision may seem relatively poor, if compared to what can be obtained at a local scale from the study of former shorelines data. In addition, tectonic, isostatic, steric and hydrodynamic factors are neglected. Nevertheless, continuous oceanic cores have the great advantage that they can cover long time sequences, making approximate estimations of eustatic oscillations possible for periods even longer than the whole Quaternary. According to Milankovitch's astronomical theory, major climatic changes have an astronomical origin, with cycles of near 100 ka for orbital eccentricity, 41 ka for orbital obliquity, and 23 ka and 19 ka for precession phenomena. The age of the climate oscillations deduced from oceanic cores is generally estimated with good accuracy, through calibration with selected astronomical (e.g. insolation at 65°N) curves.

Even with some approximation, eustatic oscillations can be very useful to coastal geomorphologists, e.g. to those who study sequences of datable raised marine terraces in uplifting areas. Each terrace, especially if made of coral reefs, can be considered to have developed when the rising sea level overtook the rising land, and therefore corresponds to a sea-level transgression peak. In this manner, the sea level relative to a stable oceanic floor can be extracted from each dated section if the uplift rate for that section is known (Chappell and Shackleton 1986).

Estimates of eustatic changes during the last century have been attempted by many authors, mainly using tide-gauge records. Discrepancies arising from different analysis methods (for a critical review, see Pirazzoli 1993) remain high enough, however, for IPCC (2001) not to choose between a recent sea-level rise of 1.0 mm yr^{-1}, or an upper bound of 2.0 mm yr^{-1}, or a central value of 0.7 mm yr^{-1} estimated independently from observations and models of sea-level rise components.

Satellite data are more reliable than tide gauges for global calculations. According to Topex/Poseidon, a global sea-level rise of 2.5 ± 0.2 mm yr^{-1} can be inferred between January 1993 and December 2000 (Cabanes *et al.* 2001). However, El Niño events produce significant oscillations in the global sea level trend and a few decades of additional records would be necessary before a reliable assessment of the present-day eustatic trend can be made with some confidence.

For the next century, eustatic predictions are based on climatic models and scenarios of greenhouse gas emissions. The most recent estimate (IPCC 2001) is of a global sea-level rise of 0.09 to 0.88 m for the period from 1990 to 2100, with a central value of 0.48 m. This estimate includes the variation in the quantity of oceanic water and steric effects, but is exclusive of vertical land motion and hydrodynamic effects. Therefore, it may have very little to do with the relative sea level experienced on a regional basis or at single sites.

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P.A. PIRAZZOLI

EVORSION

The erosion of rock or sediments in a river or streambed, by the impact of clear water carrying no suspended load. The process of evorsion often results in the formation of pot-holes (evorsion pot-holes) within the streambed, due to the action of eddies and vortices. The predominant processes involved in evorsion are hydraulic action and fluid stressing.

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STEVE WARD

EXFOLIATION

The shedding of material in scales or layers, it is often used interchangeably with sheeting or onion-skin weathering. Exfoliation of rock has

Plate 42 As a consequence of pressure release resulting from the erosion of overlying material, this granite near Kyle in Zimbabwe is being broken up into a series of curved sheets which parallel the land surface

been attributed to various causes including UNLOADING, insolation and HYDRATION (see INSOLATION WEATHERING). It is a process that has some applied significance and is, for example, a consideration in road, tunnel and dam construction where excavation can cause pressure release and fracturing to occur (Bahat *et al*. 1999). Exfoliation occurs at a variety of spatial scales from thin (<cm) scaling from boulders to mega-form some metres in size (Bradley 1963).

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A.S. GOUDIE

EXHUMED LANDFORM

Landforms that have been covered by sedimentary strata or volcanic rocks and then re-exposed are called exhumed. Exhumed landforms of different age are common within Precambrian shields. Oldest exhumed landforms in Australia are encountered below Proterozoic covers. Flat surfaces are often exhumed from below Lower Palaeozoic rocks on the Laurentian and Baltic shields, while etched (deeply weathered) more or less hilly surfaces extend from below Jurassic or Cretaceous rocks in Minnesota, USA, along parts of the Greenland west coast and in southern Sweden. Hilly relief is exhumed from Neogene sediments in south Poland. Glacially polished surfaces extend from below Upper Precambrian strata in northern Norway, Ordovician strata in the Sahara, and Permian strata on the Gondwana continents. Palaeokarstic features are exhumed from below Carboniferous strata in eastern Canada and in south Germany they make up the Kuppenalb, exhumed from a Cretaceous cover. Exhumed landforms give important information on past processes and their recognition is necessary for correct interpretation of present landscapes. Exhumed denudation surfaces are also important geomorphic markers for studies of Cainozoic uplift and erosion.

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KARNA LIDMAR-BERGSTRÖM

EXPANSIVE SOIL

Most clay soils experience a volume change on wetting and drying. The soils with the relatively inactive clay minerals, such as kaolinite, only produce a modest volume change; but soils containing montmorillonite and other smectite minerals, can have considerable changes of volume; expanding on wetting and shrinking on drying. This causes widespread construction problems because of damage to buildings and other structures, but also accounts for certain geomorphological features such as GILGAI.

The clay particles in soils carry electrical charges, and this accounts for their interesting relationship with water. Clay mineral particles tend to be negatively charged and to attract the cations in the soil water. These cations are hydrated because the negative end of the polarized water molecule is attracted to the charged ion. So, via the activity of the cations, the clay minerals attract water, and this confers on the clay systems the property of plasticity. This is measured via the plasticity index PI, which is low, perhaps around 20, for the inactive kaolinites, illites and chlorites, but high, perhaps up to 200, for the montmorillonites. It is the high PI systems which dominate in expansive soils. The structure of montmorillonite is such that water enters between the clay layers and generates considerable swelling force. The uplift pressure in undisturbed montmorillonite clays can be up to 0.1–0.6 MN m−2. These expansive pressures easily exceed the loads imposed by small structures such as single-family houses and single-storey schools. It is damage to these small structures which generates the vast costs caused by expansive soil effects. In the USA costs of over $2 billion a year are cited. This is about twice the cost of flood or landslide damage, and more than twenty times the cost of earthquakes.

In the USDA Soil Taxonomy system of soil classification the expansive soils fall into the order Vertisols, and they are defined as cracking soils; mineral soils that have been strongly affected by argillipedoturbation, i.e. mixing by the shrinking and swelling of clays. This normally requires alternate wetting and drying in the presence of >30 per cent clay, much of which, typically, is montmorillonite or some other smectite mineral. If not irrigated, these soils have cracks at least 1 cm wide at a 50 cm depth at some season in most years. Vertisols form the Black Cotton soils of north-west India; they form from the basalts of the Deccan plateau, under the influence of tropical weathering. These soils are classified as Usterts, i.e. ustic (dryish) vertisols; and so are the soils in east Australia which comprise the other large occurrence.

Regions underlain extensively by expansive clays can often be recognized by a distinctive microtopography called gilgai. Where undisturbed by humans, gilgai can be easily distinguished on aerial photography either as an irregular network of microridges, or, where the slope is greater than 1 per cent, as a pattern of downslope linear ridges and troughs. In Australia gilgai relief has been observed to reach over 3 metres. Gilgai microrelief can be used as a rapid means of mapping regions where a significant potential hazard from clay soil expansion can be expected.

The potential volume change of soils is controlled by a number of factors: (1) the type of clay, amount of clay, cations present and clay particle size, (2) density; dense or consolidated soils swell the most, (3) moisture content; dry soils swell more than moist soils, (4) soil structure; remoulded soils swell more than undisturbed soils, (5) loading; schools and houses with lightly loaded foundations are the most susceptible.

There are a variety of tests for expansive soils but one of the most reliable is the oedometer (consolidometer) test. In this test compacted soils are loaded and then wetted and the expansive pressures produced are measured. A simple classification can be produced:

$0–0.15\,\mathrm{MN\,m^{-2}}$	= non-critical
$0.15–0.17\,\mathrm{MN\,m^{-2}}$	= marginal
$0.17–0.25\,\mathrm{MN\,m^{-2}}$	= critical
$>0.25\,\mathrm{MN\,m^{-2}}$	= very critical

It is possible to recognize expansive soils in the field; some factors to look out for are:

Under dry soil conditions

- Soil hard and rocklike; difficult/impossible to crush by hand
- Glazed, almost shiny surface where previously cut by scrapers, digger teeth or shovels
- Very difficult to penetrate with auger or shovel
- Ground surface displays cracks occurring in a more or less regular pattern
- Surface irregularities cannot be obliterated by foot pressure

Under wet soil conditions

- Soil very sticky; exposed soil will build up on shoe soles
- Can be easily moulded into a ball by hand; hand moulding will leave a nearly invisible residue on the hands after they dry
- A shovel will penetrate soil quite easily and the cut surface will be very smooth and shiny
- Freshly machine scraped or cut areas will tend to be very smooth and shiny
- Heavy construction equipment will develop a thick soil coating that may impair their function.

These high PI, high montmorillonite soils can be stabilized by lime addition. This causes cation replacement and the soils become more rigid.

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IAN SMALLEY

EXPERIMENTAL GEOMORPHOLOGY

Experimental geomorphology is the study, under experimental conditions, of a representation of a selected geomorphological feature or landscape. The term ‘representation’ is intended to cover full-scale features, scale models (hardware models), and numerical constructs. This definition raises the question ‘what constitutes a geomorphological experiment?’ Writing in a geomorphological context, Church (1984: 563) defined a scientific experiment as ‘an operation designed to discover some principle or natural effect, or to establish or controvert it once discovered’. This activity differs from casual observation in that the phenomena observed are, to a critical degree, controlled by human agency, and from systematically structured observations in that the results must bear on the verity of some conceptual generalization about the phenomena. That definition leads to specific criteria for an experiment:

1. There must be a conceptual model of the processes or relations of interest that will ultimately be supported or refuted by the experiment, giving rise to:
2. Specific hypotheses about landforms or landforming processes that will be established or falsified by the experiment. (If the conceptual

model is a well-developed theory, the hypotheses will constitute exact predictions.)

To test the hypotheses, three further conditions are required:

3 Definitions must be given of explicit geomorphological properties of interest and operational statements of the measurements that will be made on them (sufficiently completely that replicate measurements might be made elsewhere);
4 A formalized schedule must be established of measurements to be made in conditions that are controlled insofar as possible to ensure that the remaining variability be predictable under the research hypotheses;
5 A scheme must be specified for analysis of the measurements that will discriminate the possibilities in (2).

The critical condition is the fourth one. Under this rubric, Church recognized two types of geomorphological experiment:

(a) Intentional, controlled interference with the natural conditions of the landscape in order to obtain unequivocal results about a limited set of processes that change the landscape;
(b) Statistically structured replication of observations in the landscape so that extraneous variability is effectively controlled or averaged over the experimental units.

Experiments of the second type entrain the capacity of statistical experimental design to discriminate and classify information in contexts where variability cannot actively be controlled. Ecologists face problems similar to those posed by geomorphology and have developed a sophisticated understanding of statistical experimentation (e.g. Hairston 1989) in order to deal with them.

The space and time scales associated with the development of most landforms effectively exclude them from experimental study. For this reason a definition of experimental geomorphology given by Schumm *et al*. (1987: 3) includes a careful exemption from strict experimental accountability. They wrote that experimental geomorphology is 'study, under closely monitored or controlled experimental conditions', accepting close observation as sufficient to establish a geomorphological experiment. Slaymaker (1991) similarly accepted 'formally structured', though not actively controlled, field studies as satisfactorily experimental exercises; exercises characterized by Church rather as formal case studies.

Schumm *et al*. (1987) also strongly implied that the objects of experimental manipulation would normally be small, or deliberately reduced scale, examples of field landforms. Experimental control is much more easily arranged in such cases, and their major summary book is entirely preoccupied with model studies. Models represent satisfactory experimental tests of hypotheses designed to explain features of the full-scale landscape provided that formal scaling criteria establish that the results can faithfully be extrapolated. Schumm *et al.* indicated no such requirement, even though there is a long history of scale model investigations in Earth science. Instead, they proposed two other perspectives. They suggested that reduced scale landforms be regarded simply as small prototypes. This does not release the investigator from formal scale constraints for extrapolation. They also explored the concept of model studies as analogues of full-scale systems or as studies in which there remains 'similarity-of-process' (after Hooke 1968, who elided the two perspectives). They argued that the model results might be extrapolated at least qualitatively to increase understanding of the full-scale landscape. Yet they recognized that changes in physical processes that undermine the supposed similarity may occur over large changes in scale. The approach, whilst it may be fruitful of ideas, suffers from inability to achieve exact predictions, or even to confirm unequivocal similarity of process, which disqualifies it as an experimental approach under the criteria given above.

There is, in fact, a tolerably well-developed body of conceptual and empirical studies of scale effects in geomorphology and hydrology (Church and Mark 1980), whilst formal scaling criteria have been investigated for hydrological and sedimentation processes at hillslope, channel and catchment scale. Formal scaling of generic model results (a generic model is one that captures the essential elements of a prototype whilst not conforming in inessential details with any particular prototypical example) ought to be possible in many problems.

In the field, one immediately faces the critical question whether experimental control can be adequately established. Geomorphological processes are driven by weather, which cannot feasibly be controlled at scales beyond that of a plot of order $10^2\,m^2$ (which might be enclosed). But variable forcing by weather is a fundamental feature of geomorphological systems, so one that is driven by artificially controlled weather is in some sense an unrealistic environmental system. Active manipulation should perhaps rather be focused on characteristics of the landform or landscape. Then it will usually be essential to establish a parallel reference or control case in which no manipulation is undertaken, in order that the effects of manipulation of the experimental system may be separated from the effects of variable weather.

It is helpful to differentiate landform and landscape studies. The former present more tractable space and time scales, and are more apt to represent environmental systems sufficiently simple to be amenable to control. At relatively small scales, successful experiments include plot studies of soil erosion, ground surface manipulations to investigate periglacial processes, and local applications of artificial precipitation or drainage adjustments to study effects on erosion or slope stability.

The centre of interest in geomorphology lies in transformations of landscape, which may be addressed through catchment experiments. There is far more experience with them than with any other full-scale experimental arrangement in Earth science (see Rodda 1976, for a historical perspective and critique). Much of the difficulty associated with the use of experimental catchments lies in establishing similarity between a treated catchment and its control, and in deciding how far observed results may be extended to the rest of the landscape. At the base of both issues lies the problem of establishing or measuring similarity between landscapes (Church 2003). Despite the known difficulties, the recognition of small catchments as the fundamental unit for most geomorphological process investigations ensures continuing effort to establish experimentally rigorous investigations at this scale.

Geomorphological experiments may be established inadvertantly. It is important not to overlook the potential value of landform or landscape manipulations undertaken for other purposes that nevertheless can be interpreted satisfactorily in terms of the requirements for an experiment. By this means, far larger systems than might ever be available for deliberate experimental manipulation may be studied. Examples include river rectification, certain water regulation projects, and certain changes in land surface condition. In such cases, it is important to establish a satisfactory reference comparison. Sometimes, this might be a before/after comparison in the same system; otherwise, a parallel reference case must be identified.

Numerical experimentation – the construction and operation of numerical models of geomorphological processes – represents a means by which complete control can indeed be gained over the conditions that drive landscape development. The penalty, of course, is uncertainty whether the numerical model faithfully represents all of the significant processes at work in the world. There also remain significant questions surrounding the means by which model outcomes may be compared with real landscapes, similar to those encountered in comparisons between real landscapes. Nevertheless, numerical modelling holds the promise to be an effective means to establish experimental control in geomorphological studies, especially ones concerning the development of entire landscapes over geomorphologically significant periods of time.

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MICHAEL CHURCH

EXTRATERRESTRIAL GEOMORPHOLOGY

The term 'extraterrestrial geomorphology' was not included in the 1968 *Encyclopedia of Geomorphology*. Indeed, at first reading, this term might seem to be an oxymoron. Should not a science of Earth forms (geomorphology) exclude those forms that are beyond the Earth (extraterrestrial)? The answer to this question depends on how one views the nature of science. Is a science more about methods and attitudes of study, or is it more about the organized accumulation of facts associated with specific subject matter? While organized fact accumulation might require precise definition as to the location of its subject matter, the methods and attitudes of geomorphology are readily conceived as extending to landforms on Earth-like planets (Baker 1993), if only better to understand Earth's landforms. To the extent that geomorphology emphasizes methods and attitudes for the study of landforms and landscapes, then it is no more restricted to studying Earth's landforms and landscapes than geometry is restricted to measuring the Earth's mathematical form.

Despite the immense excitement of planetary exploration during the 1960s, 1970s, and 1980s, there was a conspicuous lack of attention to planetary surfaces by mainstream geomorphologists. Nearly all the study of newly discovered landscapes was performed by scientists with very little background in geomorphology. More recently, increased attention to extraterrestrial geomorphology is indicated by Dorn's (2002) analysis of citations to late twentieth-century research works in geomorphology. Two of the 'top ten' cited geomorphology papers in recent years were directly concerned with topics in extraterrestrial geomorphology.

Historical and philosophical perspectives

It was not long after Galileo Galilei (1564–1642) first used a telescope to observe curious circular depressions on the Moon that Robert Hooke performed the first known geomorphological experiments to explain the origin of those depressions. Hooke was an intellectual adversary of Sir Isaac Newton, and, unlike Newton, he had a great interest and considerable talent for geology and geomorphology. In 1665 he compared the newly discovered lunar craters to (1) the cooled surface crust of melted gypsum, which was disrupted by bursting bubbles, and (2) the impact of musket balls and mud pellets into a clay-water target material. Using analogy as his mode of reasoning, Hooke hypothesized that the lunar craters formed either by (1) internal heat that melted and disrupted its surface crust (today we would describe this process as volcanism), or (2) impacts by particles from space (today these would be described as meteor impact craters). The controversy over these two origins for lunar craters actually continued until the 1970s, when it was finally resolved in favour of the impact hypothesis on the evidence of the lunar rocks returned by space missions.

Analogical reasoning was extensively employed by the geomorphologist Grove Karl Gilbert in his studies of (1) lunar craters, to which he correctly ascribed an impact origin (Gilbert, 1893), and (2) a crater in northern Arizona (now known as Meteor Crater; Plate 43), to which he incorrectly ascribed a volcanic origin (Gilbert 1896). The limitations of analogic reasoning in extraterrestrial geomorphology continue to the present day, as ably summarized by Mutch (1979):

1 Many landforms cannot be assigned a unique cause. Rather, the same landforms may be generated by different combinations of processes that converge to the same result.
2 The photointerpreter is artificially constrained in his analysis by his range of familiarity with natural landscapes. Because of these limitations, the student of extraterrestrial landforms must know as much as possible about the origin of landforms in general.

Planets, moons, and other objects

The term *planetary geomorphology* (Baker 1984) is also used for many of the topics covered by this

Plate 43 Meteor crater in northern Arizona. With a diametre of 1.2 kilometres, the crater formed about 25,000 years ago when an iron meteor struck the Earth at a velocity of about 11 kilometres per second

article, and it is true that planetary surfaces provide the sites of many landforms and landscapes (Greeley 1994). However, not all planets have rocky surfaces on which there are landforms. Moreover, there are many objects beyond Earth that are not planets, and many of these do indeed have landforms and landscapes. If we hold to the idea that one seeks to compare Earth's geomorphology to that of objects beyond Earth, then extraterrestrial geomorphology seems to be the appropriate term.

While future extraterrestrial geomorphology will surely extend to planetary objects in other solar systems, over a hundred of which have been discovered at the time of this writing, discussion here will be limited to the rocky objects of our own solar system. The inner planets, Mercury, Venus, Earth and Mars, all have rocky surfaces on which the effects of volcanism, tectonics and impact craters are in abundant evidence. Earth has a relatively large moon with a surface dominated by impact craters, the study of which has been directly facilitated by human visitation. Mars has two moons, but these are really captured asteroids, and are similar to many thousands of objects that occur throughout the inner solar system, mainly in the so-called 'asteroid belt' between Mars and Jupiter. The planets of the outer solar system, Jupiter, Saturn, Uranus and Neptune, are all giant gas balls, lacking any surface with landforms. Their satellites, however, are phenomenally rich in landscape complexity. Jupiter has four very large moons, Io, Europa, Ganymede, and Callisto, which form a kind of mini solar system, first discovered by Galileo's telescopic investigations. Ganymede and Callisto have heavily cratered surfaces on ice that is so cold it behaves as rock. Io's surface is dominated by volcanism that is much more active than any volcanism on Earth. Europa has a very young, nearly uncratered surface that is locally deformed because the icy crust of this moon overlies an immense ocean of liquid water. Other satellites of the outer planets are similar to asteroids in their surface character. Miranda, a moon of Uranus, looks to have been totally shattered by impact, and then accreted once more from the shattered remnants. The icy satellites of Uranus and Neptune are so cold that ices of ammonia and other volatiles comprise their rocky surfaces. A type of volcanism, known as 'cryovolcanism', was generated when these ices melted.

Titan, a moon of Saturn, has a diameter equal to about one-half that of Earth. It has an atmosphere that is slightly thicker than that of Earth, and, like Earth, the atmosphere is composed mainly of nitrogen. However, Titan is also extremely cold. The other main gas in its atmosphere is methane, and the great cold would lead to the condensation of that gas as a liquid on the surface of the satellite. Thus, Titan could have an ocean of methane and other hydrocarbons, or the liquid might just occupy lakes in the impact craters of a rocky surface, on which the 'rock' might be water ice. In any case, there is a very complex spacecraft, Cassini, which is scheduled to arrive in the Saturn system in July of 2004. The radar instrument on Cassini will permit penetration of the hazy atmosphere of Titan to reveal, for the first time to human observers, the landforms and landscapes of this haze-shrouded world.

Cratering and volcanism

Impact craters are the most ubiquitous landforms on the rocky and icy planets and satellites. Relatively low densities of impact craters indicate surfaces that are relatively young and unmodified relative to the 4.5 billion-year age of the solar system. Such surfaces comprise the dominant portions only of the large icy satellites Europa and Triton (a satellite of Neptune), the volcanically active satellite Io, and the planets Venus and Earth. In contrast, Mercury, Mars, the moon, and most planetary satellites have much of their surfaces covered with densely cratered terrains that formed over several billion years (Plate 44, p. 357). These surfaces are preserved because of minimal modification by active surficial processes related to atmospheric effects (exogenetic processes) and relatively localized volcanism and tectonic effects (endogenetic processes).

Volcanism is also very common in the solar system, though it does not dominate the landscapes of objects other than Io, Venus and Earth. All the rocky planets do have extensive volcanic plains, however. On Earth, these are hidden beneath ocean waters, and they were emplaced by seafloor spreading volcanism, mostly within the last 100 million years. Mercury has extensive intercrater plains, and both Mars and the moon have lowland plains that show evidence of being covered by lava flows. The lavas that formed these plains all seem to have been highly fluid, probably with basaltic compositions. Venus has some of the most extensive volcanic plains, and some of these are crossed by remarkable lava channels. The longest of these extends over 6,800 kilometres, making it longer than Earth's longest river (Baker *et al.* 1992).

Volcanic constructs, including large cones, shield volcanoes and calderas occur on Venus, Earth and Mars. Olympus Mons, a shield volcano on Mars, measures over 700 kilometres in diameter, and it rises to a height of 25 kilometres. It is only one example of extraterrestrial landforms that are much larger than their counterparts on Earth (Baker 1985). Though most extraterrestrial volcanic landforms are relict, the active volcanism of Io is spectacular. Eruptive plumes from the surface of Io were observed by the Voyager spacecraft to propel debris up to 300 kilometres above the surfaces and to deposit material up to 600 kilometres from the active vents. There are also active eruptive plumes on Triton, but the responsible process is probably more similar to that of a geyser than that of a volcano.

Tectonic landforms

Most of the rocky planet and satellite surfaces show evidence of structural deformation, with various fractures, graben and faults being the most common features. Mercury was deformed very early in its history by immense compressional forces that produced thrust fault landscapes. Mars has immense fracture zones and graben. However, only Earth exhibits the distinctive landforms associated with plate tectonics, including mid-oceanic ridges, transform faults and convergent continental margins with fold and thrust belt mountain ranges. Despite its density, radius and other geophysical similarities to Earth, the planet Venus does not show plate-tectonic landforms. This raises interesting questions about what makes plate tectonics unique to Earth.

Hillslopes and mass movement

Slopes occur on all the rocky planets. On airless bodies, only gravity and impact processes generate slope processes, but the atmospheres of Mars, Venus and Titan invite comparisons to other processes on Earth. A particularly interesting problem is the movement of extremely large (millions of cubic metres) slides or avalanches of rock and debris. Such masses on Earth have very high mobility over flat terrain. The cause of the very high mobility has been ascribed to the cushioning effect of air or water, reducing the effective pressure of the slide mass that would resist broad lateral spreading. However, these types of MASS MOVEMENT occur on the moon, which lacks both air and water. Many examples also occur on Mars, where air and water may have exerted influences.

Aeolian landscapes

While most extraterrestrial surfaces are airless, the atmospheres of Earth, Mars, Venus and Titan invite the interplanetary comparison of aeolian processes and landforms. Mars has the greatest variety of aeolian landforms. Crescent-shaped and transverse DUNES (see DUNE, AEOLIAN), wind ripples, YARDANGS, pitted and fluted rocks, and various dust streaks are all well displayed. There are also remarkable tracks produced by Martian dust devils. Aeolian bedforms also occur on Venus, which has an atmospheric pressure on the land surface that is ninety times that of Earth.

Channels, valleys and fluvial action

Besides Earth, fluvial action seems to have occurred only on Mars, and the most extensive fluvial processes were active in the remote past. The two main varieties of fluvial landforms on Mars are valley networks and outflow channels, morphological attributes of which are reviewed by Baker (1982, 2001). A great many of the valley networks occur in the old cratered highlands of Mars, leading to the view that nearly all of them formed during the heavy bombardment phase of planetary history, prior to 3 or 4 billion years ago. The outflow channels, in contrast, involve the immense upwelling of cataclysmic flood flows from subsurface sources, mostly during later episodes of Martian history. Much of the Martian surface is underlain by a thick ice-rich permafrost zone, a 'cryolithosphere', and the water feeding the outflow channels emerged from beneath this permafrost, possibly associated with volcanic processes (Baker 2001).

One of the most striking recent discoveries is that some water-related landforms on Mars are exceptionally young in age. This fact was prominently demonstrated by images from the Mars Global Surveyor orbiter showing numerous small gullies generated by surface runoff on hillslopes. The gullies were most likely formed by the melting of near-surface ground ice and the resulting debris-flow processes. The gullies are uncratered, and their associated debris-flow fan deposits are superimposed both on aeolian bedforms (dunes or wind ripples) and on polygonally patterned ground, all of which cover extensive areas that are also uncratered. Exceptionally young outflow channels and associated volcanism also occur on Mars. Data from Mars Global Surveyor show that localized water releases, interspersed with lava flows, occurred approximately within the last 10 million years. The huge discharges associated with these floods and the temporally related volcanism should have introduced considerable water into active hydrological circulation on Mars. It is tempting to hypothesize that the young outflow processes and volcanism are genetically related to other very young water-related landforms. The genetic connection for all these phenomena might well be climate change, induced by the water vapour and gases introduced to the atmosphere by both flooding and volcanism (Baker 2001)

Lakes, seas and 'oceans'

On Earth bodies of standing water include (1) lakes, in which the water is surrounded by extensive land areas, (2) seas, in which saline waters cover the greater part of the planetary surface, and (3) the ocean, which is the vast, interconnected body of water that covers about 70 per cent of Earth's surface. For Mars there is no direct geomorphological evidence that the majority of its surface was ever covered by standing water, though the term 'ocean' has been applied to temporary ancient inundations of the planet's northern plains. Although initially inferred from sedimentary landforms on the northern plains, inundation of the northern plains has been tied most controversially to identifications of 'shorelines'. New data indicate the presence of a regionally mantling layer of sediment, which seems to be contemporaneous with the huge ancient flood discharges of the outflow channels. Though the debate over the Martian 'ocean' has received much attention, even more compelling evidence supports the existence of numerous lakes, which were temporarily extant on the surface of Mars at various times in the planet's history.

Glacial and periglacial landforms

Evidence for past glacial activity on Mars is both abundant and controversial. The glacial features are also associated with periglacial landforms, which include debris flows, polygonally patterned ground, thermokarst, frost mounds, pingos and rock glaciers. On Earth most of these landforms develop under climate conditions that are both warmer and wetter than the conditions for cold-based glacial landforms (Baker 2001). The implications for past climatic change on Mars are profound because glaciers require substantial transport of atmospheric water vapour to sustain the snow accumulation that generates the positive mass balance needed for glacial growth.

The glacial landforms of Mars are erosional (grooves, streamlined/sculpted hills, drumlins, horns, cirques and tunnel valleys), depositional (eskers, moraines and kames), and ice-marginal (outwash plains, kettles and glacilacustrine plains). Of course, the landform names are all genetic designations, and ad hoc alternatives have been suggested for many. What is not ad hoc, however, is that all the glacial landforms occur in spatial associations, proximal-to-distal in regard to past ice margins, that would be obvious in a terrestrial setting. Areas of past glaciation on Mars (Kargel and Strom 1992) include the summits of very large volcanoes, uplands surrounding major impact basins (Plate 44), and the polar

Plate 44 Oblique view of the Martian impact basin Argyre, surrounded by mountainous uplands (centre) many of which contain glacial features (Kargel and Strom 1992). Note the high clouds in the Martian atmosphere on the horizon

regions, where the ice caps were much more extensive during portions of post-Noachian time.

The future of geomorphology

It has long been apparent that the modern frontier of geomorphology, both as a matter of physical discovery and as an intellectual challenge, lies in the comparative study of planetary surfaces. This was summarized rather distinctly by Sharp (1980), as follows:

> Planetary exploration has proved to be a two-way street. It not only created interest in Earth-surface processes and features as analogues, it also caused terrestrial geologists to look at Earth for features and relationships better displayed on other planetary surfaces. Impact cratering, so extensive on Moon, Mercury, and Mars, is a well-known example. Another is the huge size of features, such as great landslides and widespread evidence of large-scale subsidence and collapse on Mars, which suggests that our thinking about features on Earth may have been too small-scaled. One of the lessons from space is to 'think big'.

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SEE ALSO: astrobleme; crater; geomorphology

VICTOR R. BAKER

F

FABRIC ANALYSIS

Measures one or more parameter of the three-dimensional disposition of elongated rock fragments in sediments. Such fragments have length, breadth and thickness, defined as a, b and c axes. The fragments contain three projection planes, maximum, intermediate and minimum. The maximum plane contains a and b axes, the intermediate a and c axes and the minimum b and c axes. Measurements of the orientation and dip values for axes and planes combined with statistical analysis can identify processes and environments of deposition for many sediment types (Andrews 1971; Dowdeswell and Sharp 1986). For example, in an undisturbed lodgement till the a-axes of pebbles are strongly oriented in the direction of local ice flow and dip slightly up-glacier. On the bed of a river cobbles and boulders frequently exhibit imbricate structure in which the a-axes are normal to the water current and the maximum plane dips upstream. In a storm beach deposit the a-axes of cobbles or shingles (flat cobbles) are usually deposited normal to the direction of wave advance and the maximum plane dips seaward.

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ERIC A. COLHOUN

FACTOR OF SAFETY

The factor of safety, F, is defined as the ratio of the sum of resisting forces (shear strength) divided by the sum of driving forces (shear stress) of a slope:

$$F = \frac{\text{sum of resisting forces}}{\text{sum of driving forces}}$$

If at a location within a soil mass the shear stress becomes equal to the shear strength of the soil, failure will occur at that point. In this case $F = 1$. Where $F < 1$ the slope is in a condition for failure, where $F > 1$ the slope is likely to be stable.

Shear strength and shear stress were originally expressed by Coulomb in 1776. Shear strength of a soil is its maximum resistance to shear. Its value determines the stability of a slope. The knowledge of the shear strength is an essential prerequisite to any analysis of slope stability and the factor of safety. Coulomb postulated that:

$$\tau_f = c + \sigma \tan \varphi$$

where τ_f = maximum resistance to shear, c = cohesion of the soil, σ = total stress normal to the failure surface, and $\tan \varphi$ = angle of internal friction of the soil.

In 1925 Terzaghi published the fundamental concept of effective stress, $\sigma' = \sigma - u$ (with u = pore-water pressure), that water cannot sustain shear stress and that shear stress in a soil can be resisted only by the skeleton of solid particles at the particle contact points. Shear strength is expressed as a function of effective normal stress as:

$$\tau_f = c' + \sigma' \tan \varphi'$$

in which the parameters c′ (effective cohesion) and φ′ (effective angle of friction) are properties of the soil skeleton.

The factor of safety can then be expressed as

$$F = \frac{c' + \sigma' \tan \varphi'}{\tau_f}$$

This equation can be used for limit equilibrium methods in slope stability analysis (Duncan 1996). The calculation of F requires the description of a potential slip surface which is defined as a mechanical idealization of the failure surface. The critical slip surface is the one with the minimum value of F of all possible slip surfaces included in the limit equilibrium calculation.

Most natural hillslopes prone to landsliding have F values between about 1 and 1.3, 'but such estimates depend upon an accurate knowledge of all the forces involved and for practical purposes design engineers always adopt very conservative estimates of stability' (Selby 1993). In practice the highest uncertainties are related to soil water, especially with the spatial variability of PORE-WATER PRESSURE and seepage.

In geomorphology the factor of safety concept is essential to understand landscape stability. F is considered as ratio of landform strength resistence and the magnitude of impacting forces.

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SEE ALSO: shear and shear surface

RICHARD DIKAU

FAILURE

Within geotechnical geomorphology, the term failure implies the occurrence of a dislocation within a material, usually accompanied by detachment of a soil or rock mass. The most common case involves shear (see SHEAR AND SHEAR SURFACE) failure along a well-defined plane of rupture as a LANDSLIDE. In hard sedimentary strata, igneous and metamorphic rocks, detachments usually occur along planes of weakness defined by bedding, joints (see JOINTING), foliation and faults. The potential for movement is greatest where layers dip downslope. The resultant translational landslides are called rockslides. For dry layers, the kinematic criterion for failure is:

$$\phi' < \delta < \beta$$

where ϕ' is the friction angle along the layers, δ is the dip angle, and β the slope angle. The inequality shows that the weak layer must crop out on the slope (i.e. $\delta < \beta$), and the dip angle must exceed the friction angle. In most cases, the temporal variation of all three parameters is typically small. However, frictional resistance along joints can be abruptly reduced when water pressure increases. The steepest stable angle, θ_c, of a rock layer is then:

$$\theta_c = [(\gamma_{sat} - m\gamma_w) / \gamma_{sat}] \tan \phi'$$

where γ_{sat} is the saturated unit weight of the material, γ_w is the unit weight of water, and parameter m is the ratio of the saturated depth to the total slab depth. Similar principles apply to failures in colluvial materials where planar detachments called debris slides often occur at the interface between COLLUVIUM and harder materials below, such as bedrock. Debris slides also occur in glacial materials, for example at the interface between loose unconsolidated ablation till and denser basal till below. In softer rocks, such as clays, shales and mudstones, a lesser degree of structural control exists, and shear surfaces often run oblique to the direction of bedding as rotational failures. In these cases, the above assumption of a plane translational slide is no longer valid.

A more general way to assess the stability or proximity to failure of a soil or rock mass, of any geometrical shape, is the Mohr–Coulomb equation:

$$s = c' + (\sigma - u) \tan \phi'$$

where s is shear strength, c′ is COHESION, σ is total normal stress, u is PORE-WATER PRESSURE and ϕ' the angle of shearing resistance. Parameters c′ and ϕ' are material properties that control shear strength at the ambient EFFECTIVE STRESS. Pore pressure, u, is independent of these parameters, and is a function of moisture recharge from antecedent and ambient climatic events. The overall stability of a mass is assessed from its FACTOR OF SAFETY, $F = s/\tau$, where τ is shear stress. By definition, failure occurs when $F = 1.0$. Failure is most commonly caused by saturation, which causes an increase in shear stress concomitant with a reduction in frictional strength. This explains why so many landslides are associated with major rainstorms or snowmelt.

The above version of the Mohr–Coulomb equation applies to drained failures, which involve no excess pore pressures. In the case of rapid movements in low density, fine-grained, saturated soils (for example, QUICKCLAYS), collapse of the soil structure under shear loads causes significant excess pore pressures to develop. Such failures involve LIQUEFACTION, and must be analysed with reference to undrained (see UNDRAINED LOADING) strength parameters.

Failure may also occur by toppling and buckling of layers, especially in thinly bedded rocks. Toppling involves forward and downslope rotation of layers and is common where strata dip steeply into a slope. For single blocks the toppling criterion is:

$$b/h < \tan \delta$$

where b and h are the breadth and height of the block and δ is the inclination of the block's base. For flexural toppling, which involves downslope rotation and interlayer slip, the criterion is:

$$\alpha < \beta - \phi'$$

where pole angle $\alpha = (90° - \delta)$ is the angle of the normal to the plane, and δ is the dip angle. The inequality shows that toppling is most likely to occur in steeply dipping strata, but may be enhanced where slopes are undercut and steepened. Buckling tends to occur where ductile, thinly bedded rocks, such as argillite and phyllite, dip downslope slightly steeper than the slope angle. When the downslope compressive stress exceeds the bending resistance of the layers, buckling may occur.

Most slope failures involve more than one type of movement. For example, a landslide dominated by plane failure at its base may also involve buckling or forward toppling of material by compression at the toe area, and tensional failure at the headscarp. Transitions from one type of movement to another are also common, for example detachment of a saturated mass as a debris slide, followed by disintegration and fluidization as a DEBRIS FLOW further downslope.

Although the Mohr–Coulomb equation implies abrupt attainment of failure, many landslides probably involve slow creep movements prior to detachment. Deep-seated gravitational movements of the SACKUNG type probably involve prolonged, slow movements at depth. Such mountain scale masses total tens to hundreds of millions of cubic metres of material moving at millimetres to centimetres per year. The surface expression of such movements is typically tension cracks, uphill facing scarps and grabens, or is less clearly defined as masses of broken, dilated rock. Such movements may occur over centuries to millennia without the development of a landslide rupture surface. However, other cases are known to have terminated in large rock avalanches (STURZSTROMS). This suggests that a continuum of slope movement rates and types may occur over time at an individual site, a circumstance which is not easily encompassed by existing methods used to classify and analyse slope movements.

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MICHAEL J. BOVIS

FALL LINE

The topographical and geological boundary between an upland region of relatively high resistance crystalline rock and a lower region of weaker rock. Rivers transcending this boundary often develop waterfalls and rapids in parallel. A less frequent and appropriate use of the term is for the point where a river ceases to be tidal. The fall line is thus a geological and geomorphological boundary. Generally, streams and rivers upstream from the fall line have small floodplains and are of low sinuosity, whereas downstream from the fall line rivers and streams tend to possess larger floodplains and display high sinuosity.

The type example of a fall line is the eastern United States region, where the upland Piedmont Plateau (crystalline rock) meets the Atlantic coastal plain (weaker sedimentary rock). The junction is marked by rapids and waterfalls on each of the major rivers that transcend the zone (i.e. the Delaware, Potomac, James, Savannah, etc.).

The steep gradient of the American fall line has been accounted for in three main ways, as reviewed by Renner (1927) and Lobeck (1930: 454). First, the feature can be interpreted as a zone of monoclinal flexing or faulting (though faulting occurs on the fall line in few localities). Second, as an area where the rivers have eroded away the softer rocks of the coastal plain, forming knickpoints at the boundary with the resistant crystalline piedmont rocks. Third, as the intersection of two ancient peneplains, in which the older mid-Mesozoic erosion surface plunges beneath the coastal plain deposits that overlie the younger peneplain. The fall line represents a stripped part of the older peneplain and accounts for its steeper slope.

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STEVE WARD

FANGLOMERATE

A sedimentary rock consisting of heterogeneous fragments of assorted size deposited in an alluvial fan and subsequently cemented into a solid mass. The term was introduced by Lawson (1913) to describe the coarse upslope parts of ALLUVIAL FAN formations, though the term is also used more generally for conglomerates and breccias deposited on alluvial fans. They are composed of two main facies: water laid deposits and mass flow deposits. Fanglomerates are characterized by their parallel bedding and decreasing particle size downslope, alongside rapid fan thinning.

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FAULT AND FAULT SCARP

A *fault* is a surface or zone along which one side has moved relative to the other in a direction parallel to the surface or zone. The term is applied to features extending over distances of metres or larger, whereas those at the scale of centimetres are called shear fractures, and those at the scale of millimetres are microfaults. Most of them are brittle structures, although some may represent ductile deformation. For an inclined fault, the fault block above the fault is the hanging wall, and the block below the fault is the footwall. Faults are subdivided into dip-slip (showing slip parallel to the dip of the fault surface), strike-slip (of slip parallel to the strike of the fault surface), and oblique-slip faults (where slip is inclined obliquely on the fault surface). The dip-slip faults include normal faults, on which the hanging wall block moves down relative to the footwall block, and thrust (dipping $<45°$) or reverse faults (dipping $>45°$), on which the hanging wall moves up relative to the footwall block. Strike-slip faults are either right-lateral or left-lateral, depending on the sense of motion of the fault block across the fault from the observer. These faults are commonly planar and vertical. Both dip-slip and strike-slip faults frequently form a linked fault system which, in cross section, consists of flats and ramps, which cut through the footwall and detach a slice of hanging wall rocks.

Some normal faults are concave-upward faults of dip decreasing with increasing depth; they are called listric faults. These can join or turn into a low-angle detachment fault at depth. Small-scale faults parallel to the major fault and showing the same sense of shear are called synthetic faults; those of the conjugate orientation are antithetic faults. A downthrown block bounded on either side by conjugate normal faults is a graben, whereas a relatively uplifted block bounded by two conjugate faults is a horst. A half-graben is a lowered tilted block bounded on one side by a normal fault. Step faults are parallel faults on which the downthrown side is on the same side of each fault. Rotational movements between the two fault blocks result in varying throws along the fault strike, producing either hinge faults, where displacement increases from zero to a maximum along the strike, or pivot (scissor) faults, where one block appears to have rotated about a point on the fault plane. The traces of normal faults are either straight or slightly sinuous, depending on the fault dip, whereas the traces of thrusts are usually highly sinuous due to low-angle intersection with the ground surface.

Large-scale strike-slip faults are called transform faults when building segments of lithospheric plate boundaries, or trancurrent faults when they occur in continental crust and are not parts of plate margins. Strike-slip faults frequently form bends (curved parts of the fault trace) and stepovers, i.e. places where one fault ends and another, *en echelon* fault begins. A left bend or stepover in a right-lateral fault system (restraining bend) induces local compression (uplift; transpression), whereas a right bend or stepover in a right-lateral fault (releasing bend) produces local extension (subsidence; transtension). Displacement at extensional bends and stepovers forms rhomboidal, fault-bounded depressions, called pull-apart basins.

A *fault scarp* is a tectonic landform coincident with a fault plane that has displaced the ground surface. A residual fault scarp is a mature scarp, upon which the original tectonic surface has been obliterated by geomorphic processes. A fault-line scarp, in turn, results from differential weathering and erosion of the rocks on either side of the fault.

Scarps produced by normal faulting are usually located at the contact between bedrock in the footwall and Quaternary sediments in the hanging wall. Scarps associated with reverse faulting in solid bedrock are commonly overhanging and tend to collapse and/or be eroded; they are also more deeply embayed than their normal counterparts. Scarps associated with strike-slip faults are less prominent and are best developed in areas of uneven topography. In loose sediments, however, fold-limb (monoclinal or fold) scarps are formed, and usually are surface expression of blind thrusts.

Active normal fault scarps include: piedmont (simple) scarps, formed in unconsolidated deposits; multiple (complex) scarps, related to formation of a fault splay during a single faulting event; composite (multi-event, compound) scarps, up to a few tens of metres high, formed due to renewed slip on a fault; and splintered scarps, produced due to fault displacement distributed across overlapping *en echelon* segments. A piedmont scarp (Wallace 1977) includes a steep ($>50°$) free face, a moderately inclined (30–40°) debris slope, and a gently inclined (5–10°) wash slope (see SEISMOTECTONIC GEOMORPHOLOGY). Fault scarps in semi-arid climate degrade from gravity-controlled (10^2 yrs), through debris-controlled (10^3 yrs), to wash-controlled (10^5 yrs) slope due to either: decline, replacement, retreat or rounding (Mayer 1986).

Depending on the climatically controlled rate of removal of debris shed from the scarp, the Oregon or Basin and Range-type scarps, typical of semi-arid climate, and the Awatere or New Zealand-type scarps, formed in a more humid climate, have been distinguished. Faulting in bedrock is accompanied by fracturing and brecciation which can seriously modify the bedrock susceptibility to erosion. Fault rocks of contrasting resistance to erosion are typical of the Aegean-type fault scarps (Stewart and Hancock 1988), where normal faults in carbonate bedrock are underlain by different types of alternating compact and incohesive breccias. Degradation of such scarps proceeds differently as compared to the Nevada-type model of piedmont scarp.

Colluvial wedges shed from fault scarps can be dated by: ^{14}C, luminescence, dendrochronological, palynological, tephrochronological and weathering rates techniques. Scarps formed in loose sediments can be modelled mathematically by: linear regression, diffusion modelling and statistical analysis of scarp parameters.

Due to repeated episodes of faulting, bedrock fault escarpments, several hundred metres high, and fault-generated range fronts, several hundreds of kilometres long and up to 1 km high, can form. The range front morphology is determined mainly by the ratio of uplift to erosion. Range fronts in a humid climate may appear more degraded than range fronts with the same uplift rate in an arid climate. Active normal fault-generated mountain fronts frequently display triangular or trapezoidal facets (faceted spurs, flat irons) that form due to uplift and dissection of a normal scarp by gullies and whose bases are parallel to the fault trace. Flights of faceted spurs have been interpreted as a result of either episodic uplift, distributed faulting within the range-bounding fault, or even active landsliding.

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SEE ALSO: seismotectonic geomorphology

WITOLD ZUCHIEWICZ

FECH-FECH

A term applied in the Sahara Desert to fine silt with a powdery consistency and to fine superficial deposits with a low density that often contain evaporates. Progress across fech-fech is made difficult by the absence of cohesion of particles (where feet or wheels penetrate). Fech-fech can be classified from a genetic point of view into two main types:

- Fech-fech developed on Holocene lacustrine muds or fluvio-lacustrine sediments: soft zones within the lacustrine limestones, with 40 per cent of fine particles ($< 20\,\mu$) and a higher content of soluble salts, differentiate these sediments in an environment which is almost exclusively sandy.
- Fech-fech developed on clayey shales: the present-day weathering of shales leads to their superficial expansion, which is accentuated by the incorporation of aeolian detrital particles between the disconnected layers.

In addition to these two types, one also finds fech-fech on Quaternary regs with a denser structure ($1.5\,g\,cm^{-3}$) due to the formation of aggregates of silty and salty sand. Tracks can be preserved for a long time on these soft regs where they are underlain by a sandy layer.

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MOHAMED TAHAR BENAZZOUZ

FERRALLITIZATION

A process characterized by the aggressive leaching of a substrate as a consequence of intense tropical weathering, and in which the net effect is a relative accumulation of iron-rich (and commonly also aluminium-rich) compounds; in particular iron and alumina sesquioxides. Ferrallitization is an *in situ* process during which prolonged or intense weathering causes the breakdown of the primary constituents of a pre-existing soil or rock substrate.

Ferrallitization progresses upon rock substrates by the hydrolysis of their primary minerals. This leads to the individualization of the chemical elements of these minerals, the complete leaching of constituent alkali and alkali Earth elements, and the partial or total leaching of silica. Once breakdown commences and elements released, they become available for removal from the system, whilst less mobile constituents, such as Fe, Al, and Ti, remain behind as residual materials and form sesquioxides. Since these less mobile constituents are present in significant proportions within many common substrate lithologies (e.g. 12–18 per cent Fe_2O_3, 12–18 per cent Al_2O_3 in continental basalt, and 0.5–5 per cent Fe_2O_3, 12–15 per cent Al_2O_3 in granitic materials), the residuum readily becomes enriched in Fe and Al. Any silica remaining in the residuum is present either as corroded primary quartz crystals or grains, or else becomes combined into alteration products (e.g. kaolinite and gibbsite). Since, by definition, laterites (see FERRICRETE) form by *in situ* mineral breakdown, ferrallitization represents a key process in the development of lateritic weathering profiles.

MIKE WIDDOWSON

FERRICRETE

A horizon, at the land surface, made up of the cementation of near-surface materials by iron oxides, and often forming a resistant DURICRUST. Typically between 1–20 m in thickness, it can form laterally extensive sheets which may extend over a few, to hundreds, or even thousands of km^2. Consequently, it is perhaps the most widespread of all the duricrust materials. At outcrop it comprises a massive, interlocking fretwork of iron, and often aluminium compounds (i.e. sesquioxides) that bind together other lithological and pedogenic components.

Ferricrete has a long history of study by geologists, geomorphologists, pedologists and agronomists. Considerable effort has been directed toward determining the conditions under which it forms, and this has proved crucial in advancing many aspects of TROPICAL GEOMORPHOLOGY (Thomas 1994; Widdowson 1997). Moreover, chemical and physical durability of ferricrete has meant that it has often played a prominent role in evolution of tropical, and subtropical landscapes (e.g. McFarlane 1971; Bowden 1987; Widdowson and Cox 1996).

In its broadest sense, the term ferricrete can be used to describe any duricrust material in which the dominant bulk components are iron-rich compounds. However, whilst this may seem a straightforward definition, difficulties arise because the term has been employed to describe a wide range of terrigenous weathering and alteration products resulting from differing processes of formation (Ollier and Galloway 1990). Therefore, it becomes important to understand the differences and, where possible, make distinctions between genetically different types of iron-rich duricrust.

Since the nineteenth and early twentieth centuries, the terms ferricrete ('an iron-rich crust'; Lamplugh 1907) and laterite ('a highly weathered material rich in secondary forms of iron and/or aluminium'; (Buchanan 1807; Babbington 1821; Sivarajasingham *et al.* 1962; Plate 45) have been used interchangeably to describe iron-rich duricrusts of various genetic origins. This has led to considerable confusion. However, the problems of co-ordinating laterite and ferricrete description stem not only from investigation by a variety of different scientific disciplines, but also from the development of extensive anglophone and francophone descriptive terminologies. Nevertheless, it is evident from field studies that the majority of iron-rich duricrusts can be adequately described in terms of two genetically distinct types. Aleva (1994) distinguishes between those duricrusts in which an absolute iron enrichment occurs (i.e. those which receive a net input of iron), and those which attain their elevated iron contents through residual enrichment within the profile (i.e. no net input of iron).

Ferricretes are those duricrusts which incorporate materials non-indigenous to the immediate locality in which the duricrust formed. In many instances the transported materials can be readily

Plate 45 Laterite quarry near Bidar, south-east Deccan, India (with 1 m scale in lower right). Material beneath the indurated duricrust of a laterite profile is excavated, cut into large bricks, and allowed to harden in the sun. This is similar to the material first named 'laterite' by Buchanan (1807)

identified as pebbles or clasts derived from adjacent lithological terranes, or as fragments from indurated layers of earlier generations of laterite or ferricrete (Plate 46). Importantly, the term ferricrete should also be extended to those materials whose constituents have been substantially augmented by the precipitation or capture of elements and compounds from allochthonous fluids (i.e. those derived during the breakdown and mobilization of materials outside the immediate locality of ferricrete formation). Although it is the allochthony of the constituent materials of the ferricrete which justify its appellation, determining whether the introduction of such fluids has taken place, and confirming their allochthony, is often problematic. However, since ferricretes may develop as ferruginous foot slope accumulations or within topographic depressions, they can often be distinguished by the fact that they display an obvious discordance with the underlying substrate lithologies. In effect, they do not display the progressive weathering profile characteristic of many laterite profiles, and instead the ferricrete horizon sits upon relatively unaltered bedrock.

Laterites are iron-rich duricrusts which have formed directly from the breakdown of materials in their immediate vicinity, and so do not contain any readily identifiable allochthonous component. Lateritic duricrusts are typically manifest as the uppermost layers of *in situ* weathering profiles. Where these profiles are fully exposed, such as the widespread examples developed on basalt in western India (Widdowson and Gunnell 1999), they consist of an uninterrupted progression from unaltered bedrock, through the WEATHERING FRONT into SAPROLITE (in which structure and crystal pseudomorphs of the parent rock may still be recognized), and then upward through increasingly altered and iron-enriched zones that culminate as a highly indurated 'tubular' laterite at the top of the profile (Plate 47).

To summarize, ferricrete and laterite are not synonymous terms and should, wherever possible, be used to distinguish between fundamentally different types of iron-rich duricrust. This distinction is particularly important since it places constraints upon the type of processes operating during evolution of a duricrust, and the palaeoclimatic and morphological conditions existing at the time of its development. However, although emphasis is put upon establishing whether the iron component is allochthonous or autochthonous, distinguishing these two types of duricrust, both in the field and in hand specimen, can prove problematic. Problems arise because, once formed, ferricretes can begin to alter and evolve in response to prevailing climatic and groundwater conditions (Bowden 1997) and, over time, begin to exhibit some of the structural and textural features typical of lateritic weathering profiles. In effect these

Plate 46 Granular ferricrete surface comprising allochthonous materials derived from earlier generations of laterite and ferricrete, near Bunbury, Western Australia

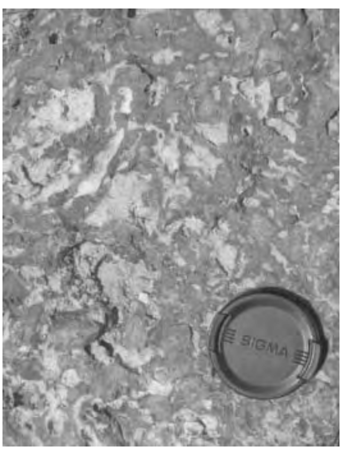

Plate 47 Indurated 'tubular' laterite sample from the top of an *in situ* weathering profile near Bunbury, Western Australia

'evolved' ferricretes become modified by a post-depositional weathering and ferrallitization overprint. Conversely, the role of allochthonous groundwater fluids, and associated lateral or downslope transport of elements and compounds, cannot always be excluded in the development of otherwise autochthonous laterite weathering profiles.

More recently, ferricrete and laterite duricrusts, together with Fe-rich palaeosols, have begun to acquire renewed importance as palaeoenvironmental indicators (e.g. Bardossy 1981; Thomas 1994; Tsekhovskii *et al.* 1995). The investigation of such materials within the geological record, together with appropriate mineralogical, geochemical and isotopic studies, can now reveal detailed information regarding past climatic and atmospheric conditions. For instance, geochemical and isotopic analyses of Proterozoic laterites from South Africa (Gutzmer and Beukes 1998), suggest not only an ancient oxidizing atmosphere, but also a hot and humid climate at *c*.2 Ga. Moreover, carbon isotope signatures preserved within these laterites may indicate the presence of an early terrestrial vegetation.

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SEE ALSO: duricrust

MIKE WIDDOWSON

FIRE

Wildfire is one of the most potent agents of geomorphic change, modifying processes and greatly increasing erosion and deposition rates in virtually all BIOGEOMORPHOLOGY environments. This entry covers the geomorphological role of fire in WEATHERING, soils, HILLSLOPE PROCESSes and river systems.

Although most texts and reviews list fire as an important weathering agent (Blackwelder 1927), most field studies are anecdotal or supported by little data with fewer experimental controls. Thus, the most important insights on fire weathering result from laboratory studies (Goudie *et al.* 1992) where experimentalists have learned that fire weathering depends heavily on rock physical properties, varies with different rock types, is faster in smaller rocks, and fire weathering rates increase with increasing water content.

An example of the impact of fire on rock weathering comes from the April–May 2000 ‘Coon Creek’ wildfire that burned around 37.5 km^2 of the Sierra Ancha Mountains, 32.3 km north of Globe, Arizona – including 25 sandstone and 19 diorite boulders surveyed in 1989 and resurveyed (a) after the burn, (b) after the summer 2000 precipitation season, and again (c) after the winter 2001 snow season (Dorn 2003, Plate 48). When stretched over cumulative boulder areas, erosion immediately after this single fire averaged >26 mm for sandstone and $>$ 42 mm for diorite. But averages are misleading, because sandstone and diorite boulders expressed bimodal patterns of erosion, where fire-induced weathering generated either (a) no

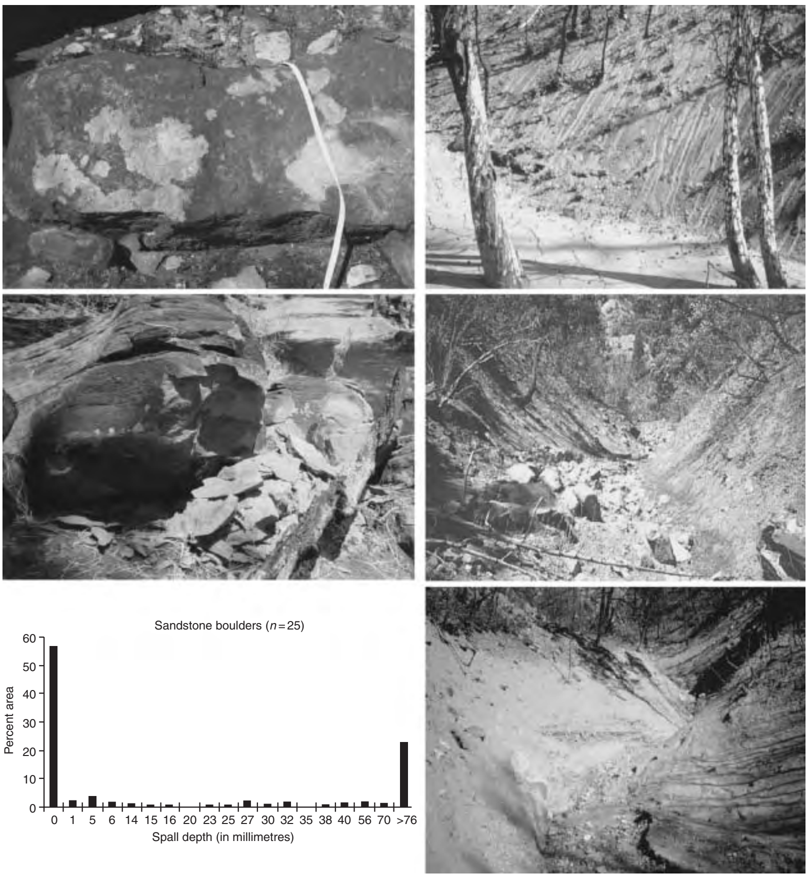

Plate 48 Left column: boulder weathering from the Coon Creek Spring 2000 fire, Sierra Ancha Mountains, Arizona. The top left image shows flaking of millimetre-scale spalls. The centre-left image shows where half of a boulder fragmented as a result of the fire. The graph on the lower left shows the overall bimodal pattern, whereby fire weathering produces erosion of small flakes or extensive slabs. Right column: fire-generated erosion from the 1995 Storm King fire, Colorado, courtesy of the US Geological Survey (Canon *et al*. 2001; see also http://landslides.usgs.gov/html_files/ofr95–508/index.html). The upper right image shows post-fire rill erosion. Other photos show in-channel conditions before (middle right) and after (lower right) passage of a debris flow

erosion to thin, millimetre-scale spalling or (b) massive spalls thicker than 7.6 cm. This field study confirms an earlier experimental finding that fire increases a rock's susceptibility to post-fire weathering and erosion processes (Goudie *et al.* 1992), since summer-time convective storms and subsequent winter snows continued to promote boulder erosion on the order of 1–5 millimetres. In addition to erosion of boulder surfaces, 85-metre- diameter boulders appear to have been fragmented into cm-scale clasts – suggesting that fire can modify hillslope evolution in locations where boulders are important controls on the evolution of slopes.

Wildfire generates extensive changes to soil systems (Morris and Moses 1987), perhaps the most important being the development of HYDROPHOBIC SOILS. Wildfires produce volatile hydrocarbons that penetrate soil up to 15 centimetres and make a water-repellent layer. In addition, fire ash decreases the ability of soils to adsorb water. Field checks involve digging a progressively deeper trench and applying water. Water that does not infiltrate immediately (within 10 seconds) indicates the soil is hydrophobic. Extreme hydrophobicity results in water ponding for more than 30 seconds. On unburned slopes, normal biogeomorphology processes decrease soil erosion, for example, by intercepting raindrop impacts, increasing infiltration and providing structural support. Hydrophobicity from burning decreases infiltration capacity, and increases OVERLAND FLOW and SOIL EROSION.

Even before it rains, burning enhances erosion by dust DEFLATION and dry ravel. Dry ravel is a type of granular MASS MOVEMENT where frictional and collisional particle interactions dominate flow behaviour, all not requiring rainfall. Dry ravel provides sediment to channels from particularly steep slopes, and this process is well documented after southern California fires.

Burning greatly increases surface runoff from precipitation, which increases the volume and velocity of the surface runoff. Higher discharge of surface water flows then result in the formation of RILLS and gullies (see GULLY) on hillsides. Fire-enhanced gullies and rills transport surface runoff and sediment to stream channels. Peak flows in the channel tend to occur with less of a lag time than those observed in unburned watersheds. Flood peaks tend to be much higher and more capable of eroding sediment stored in channels, leading to channel incision.

The sediment load of the fluvial system also changes after a fire. Sediment from a number of different sources may be incorporated into flows progressing down a hillside or channel. Sediment-water flows on burned slopes change the concentration, size distribution and/or composition of the entrained sediment to the point where a change in measurable yield strength takes place; this change is called HYPERCONCENTRATED FLOW. In hyperconcentrated flows, particles are deposited as individual grains from suspension, and the remaining fluid continues to move.

Fires also greatly increase DEBRIS FLOWS (Cannon *et al.* 2001; Swanson 1981). In contrast to streamflow or hyperconcentrated flow, debris flows host a sediment-water mixture that moves as a single phase. Deposition does not separate out particles, so debris-flow deposits have sharp, well-defined flow boundaries. The most recognizable deposits are levees lining flow paths and lobes of material at a flow terminus. Many terms have been used for the processes and deposits of debris flows, including slurry flow, mudflow and debris torrent.

Fire-enhanced debris flows start by landsliding or sediment bulking of surface water flows. Landsliding after burning tends to be more common in colluvial-filled hollows on slopes, where unconsolidated thick deposits of colluvium fail after rainfall. This landslide then mobilizes into a debris flow, where the debris-flow path can then be traced up to a landslide-scar source.

Sediment bulking tends to occur in the surface layer of hydrophobic soils. Hydrophobic soils create a condition where excess water that cannot penetrate deeply saturates the upper few centimetres of soil. This surface material then fails as small-scale debris flows. In addition, water runoff can incorporate so much loose material that sediment concentrations get high enough for the flow to behave as a debris flow. Sediment bulking is probably the most important debris-flow producing process after a fire.

GEOMORPHOLOGICAL HAZARDS are not limited to the first few rainstorms after a fire. Research by Ramon Arrowsmith in the Phoenix, Arizona, region indicates enhanced flash flooding potential decades after a brush fire. Even in forested regions, the supply of loosened material continues to deliver dry ravel sediment, hyperconcentrated flows and debris flows to stream systems for years after a fire.

The link between wildfire and increased erosion leading to large sedimentation events was made as early as 1949 by P.B. Rowe and colleagues working in southern California. They developed the concept of a 'fire–flood sequence' that has been studied extensively in a wide variety of river settings including alpine forests such as Yellowstone (Minshall *et al.* 1998), Mediterranean scrub (Shakesby *et al.* 1993) and even desert ranges (Germanoski and Miller 1995). In Yellowstone, for example, Minshall *et al.* (1998) found extensive RILL development, GULLY formation and MASS MOVEMENTs in burned watersheds during the summer of 1989, when post-fire heavy rains and snowmelt generated widespread 'black water' conditions and increased BEDLOAD and SUSPENDED LOAD. After monitoring Yellowstone streams for a decade after its massive wildfire, Minshall *et al.* (1998) stress that post-fire stream studies can yield misleading insights after only a few years since massive stream reorganization can take place seven to nine years after the fire event.

The study of fire remains associated with soils and sediment, called pedoanthrocology, provides important insight into prehistoric geomorphic changes associated with fires. Studies of fire-induced ALLUVIAL FANs, of fire remains within uneroded soils, and diagenesis of organic remains into such forms as vitrinite and inertinite provide geomorphologists with insights into palaeoecological conditions that may have influenced the geomorphic landscape seen today (Siffedine *et al.* 1994).

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RONALD I. DORN

FIRST-ORDER STREAM

STREAM ORDERING is based on the premise that stream size is related to the area contributing to runoff. This provides a method of ranking the relative size of streams within a catchment. The term first-order stream originates from ideas initially proposed by R.E. Horton in the 1930s (Horton 1932, 1945). Horton devised a method of classifying links in a stream network using a system of ordering. Under such a scheme the smallest unbranched streams in a catchment are designated first order. The combination of two first-order streams results in a second-order stream and so forth through successively larger links as additional streams join the network (Figure 60a). This original idea soon led to a proliferation of ordering schemes each providing a development

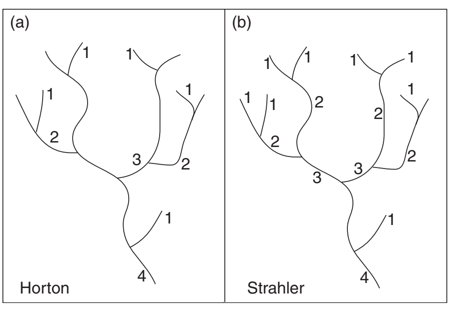

Figure 60 Comparison of stream and segment ordering methods: (a) Horton, (b) Strahler

or refinement of previous ones. Of particular note is the Strahler scheme (Strahler 1952) that begins like the Horton scheme, with the smallest channels being classified as first-order links; but higher order links are only generated when two links of equivalent order are joined (Figure 60b). The highest order generated by this mechanism is often used to classify drainage basins, e.g. a third-order drainage basin. Stream orders vary from the smallest first-order streams to the world's largest rivers that approach twelfth order (Mississippi, Amazon).

The hydrologic response of a stream channel is in part a function of its stream order. Stream order can be used to quantify other aspects of a watershed. These include the Bifurcation Ratio, R_b. The bifurcation ratio (R_b) is defined as the ratio of the number of streams of any order (N_i) to the number of streams of the next highest order. Horton (1945) found that this ratio is relatively constant from one order to another. Values of R_b typically range from the theoretical minimum of 2 to around 6. Typically, the values range from 3 to 5. The bifurcation ratio is calculated as

$$R_b = N_i/N_{i+1}$$

These are important geomorphic parameters that describe the structure and functioning of drainage basins. In the past, calculating these measures was extremely time consuming as catchment boundaries need to be carefully defined. However, these analyses are now routinely undertaken using GIS, which has the potential to provide rapid, accurate and automatic recognition of stream network links (Morris and Heerdegen 1988). This is often based on the topographic definition of streams based on contour crenulation and headwater divide delimitation. In this respect, an advantage of the Strahler scheme is that it retains the same common nomenclature for all similar sized channel links. Thus first-order streams are consistently identified as the smallest channels in a catchment. This is useful because streams with similar attributes, and a similar relative position in the network, are grouped in the same order. Hence, first-order streams tend to have common characteristics. These common characteristics are, however, dependent on the scale at which the channel links are defined, e.g. whether they are mapped from published maps or surveyed in the field. This raises important issues about consistency in definition of network properties (Blyth and Rodda 1973; Mark 1983) and highlights the property that most stream networks are dynamic, so the extent of the network varies in time from storm to storm and across seasons, years and decades.

Topography is not the only criterion used to distinguish first-order channels. First-order streams may also be defined on the basis of flow duration sufficient to sustain aquatic biota year round. In this respect, a first order channel must be by definition permanent, connected to the main stream network and convey runoff from a defined CONTRIBUTING AREA.

The greatest frequency of first-order streams tend to be found in the headwaters of catchments where channels tend to be small, confined, have steeper slopes and individually contribute only small amounts of stream discharge (Wohl 2000). In terms of the overall network, first-order channels defined by a Strahler ordering scheme commonly represent 50–60 per cent of the total stream length in a third-order drainage basin (Strahler 1964). During storms or prolonged wet periods the size of the network will expand and first-order channels may extend up hillslopes as ephemeral water flows are maintained for short periods. The extension of the permanent first-order network beyond the channel head represents a dynamic link. The coupling between the channel head and the network of hillslope hollows upslope usually defines a diffuse topographic network of zero-order basins (Dietrich *et al.* 1987). These zero-order basins are small unchannelled valleys. These form HILLSLOPE HOLLOW networks on slopes which focus runoff and sediment transport via saturated overland flow and gully and debris flows. In general terms, as stream order increases sediment yield per unit area tends to decline as HILLSLOPE-CHANNEL COUPLING becomes less effective.

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SEE ALSO: drainage basin; GIS; runoff generation

JEFF WARBURTON

FISSION TRACK ANALYSIS

Fission track analysis (FTA) is a thermochronometer that provides detailed information on the thermal history of rocks, most usually for temperatures below 350 °C (using zircon) and below 110 °C (using apatite). When a rock has cooled rapidly from its temperature of formation (e.g. a rapidly cooled lava), the technique may provide the age of formation of that rock (hence ‘fission track dating’) but the technique can be applied in any situation in which low-temperature thermal history is required and the appropriate minerals are present. In geomorphological applications, the technique exploits the increase in temperature with depth in the Earth’s crust (the geothermal gradient). This temperature increase means that the low-temperature thermal history of an apatite or a zircon now at the Earth’s surface (or in a drill hole) is a record of that mineral’s passage through the crust to the sampling point (surface or drill hole). The principal application of FTA in geomorphology is therefore to elucidate the long-term DENUDATION that brings the target mineral(s) to the Earth’s surface. For a surface temperature of 20 °C and a geothermal gradient of 25 °C km^{-1}, FTA in apatite provides a denudational history for the upper *c.*4 km of the crust (i.e. below about 110 °C). (URANIUM-THORIUM)/HELIUM ANALYSIS ((U-Th)/He analysis) in apatite provides a shallower denudational history from a lower temperature of *c.*75 °C. In geomorphological studies, in which the final stages of crustal denudation leading to the present topography are of interest, the thermochronometers most often used are the lowest temperature (i.e. shallowest), namely, apatite FTA and (U-Th)/He analysis. If all three low-temperature thermochronometers (i.e. the two in apatite plus zircon FTA) all yield essentially the same ages, then it is clear that denudation (and the associated cooling of the crust through the three thermochronometers’ temperature ranges) have occurred very rapidly.

FTA relies on counting the number, and measuring the lengths, of minute damage paths (defects or ‘tracks’) produced when the heavy daughter products of ^{238}U fission in a mineral’s crystal lattice travel away from each other at high speed through the lattice. The tracks are only *c.*5 nm in diameter and are widened slightly by etching in a weak acid during sample preparation, so as to make them visible under microscope. Etched tracks are about 1–2 μm in diameter and up to about 16 μm long. The tracks are produced continuously at a known rate, dependent on the U-content.

In order to reconstruct, from the sample’s cooling history, the denudation necessary to bring the sample to the Earth’s surface, a knowledge of the geothermal gradient at the time the sample was exhumed is necessary. This geothermal gradient provides the crustal depths associated with the temperatures from which the sample was exhumed. The ‘ancient’ geothermal gradient is usually unknown and an ‘appropriate’ geothermal gradient is often assumed based on likely modern analogues of the tectonic and thermal setting of the sample locality at the time of exhumation. If a vertical profile of FT samples is available (for example, from a drill hole or from a mountain side), the gradient of the elevation–age profile provides the geothermal gradient.

In simple terms, the number of tracks is a function of the time since the sample cooled sufficiently for the tracks to be retained (i.e. cooled below about 110 °C in apatite), and the U-content of the mineral in the areas of the grain in which the tracks have been counted and their lengths measured. The fission track age is derived by the application of the standard radiometric dating formula but with the amount of decay product (‘daughter’) of the dating technique’s radioactive decay system replaced by the number of tracks.

Lower and/or more variable rates of denudation through time result in more complex cooling histories, which can be elucidated using frequency distributions of track lengths (the ‘track length distribution'). Tracks form continuously as a result of ^{238}U fission but in apatite, for example, they are annealed (repaired) geologically instantaneously above a temperature of about 110 °C. Below 110 °C, apatite fission tracks are only partially annealed and are increasingly retained at temperatures down to surface temperature. This temperature range in which tracks are partially annealed (repaired) is the partial annealing zone (PAZ), and there is a range of views as to the effective lower limit of the PAZ. Strictly, fission tracks may be annealed even at room temperature but some authors set the effective lower boundary of the PAZ at *c.*60 °C. Track annealing is by repair at the ends, resulting in shorter tracks. The duration of the sample's residence in the PAZ is therefore reflected in the frequency distribution of track lengths, shorter track lengths reflecting longer residence time in the PAZ.

Statistical temperature–time paths can be calculated to match the measured fission track age and track length distribution, giving a complete description of thermal history of the apatite below 110 °C, and hence of the sample's trajectory to the surface as a result of denudation. Figure 61 shows the ways in which different fission track ages and track length distributions reflect different cooling histories. In A, the sample cooled very rapidly at 100 Ma ago, and the fission track age (99.8 Ma; the upper number of the three within the plot) is essentially the same as the age of the cooling event. The rapid cooling is reflected in the long mean track length (15 μm; the middle number in the plot) and the very low standard deviation of the track length distribution (1.07 μm; the third number). The track length data have a high, unimodal, narrow distribution in the histogram of the track length distribution.

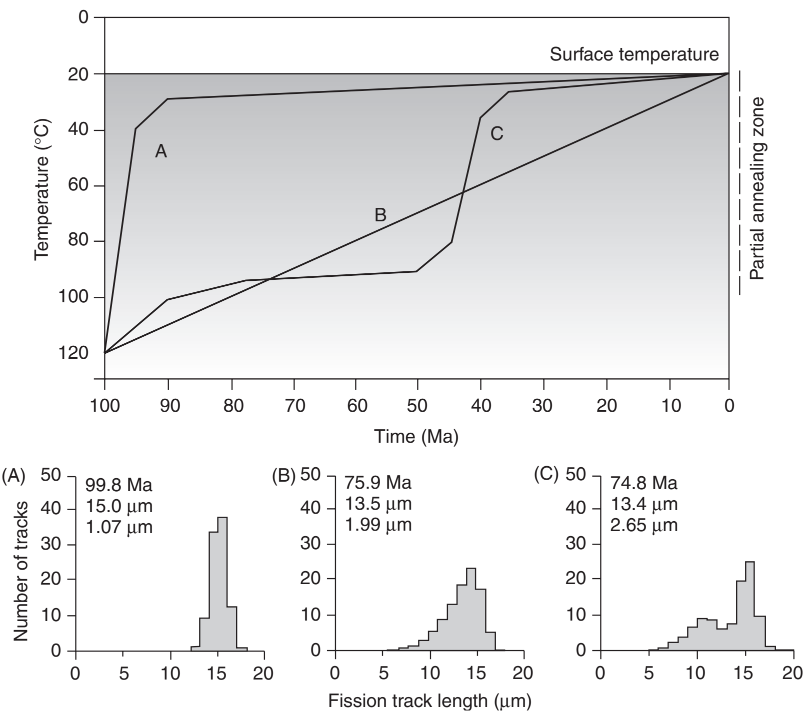

Figure 61 Fission track ages and track length in relation to cooling histories (based on Gleadow and Brown 2000: figure 4.3)

Sample A's very rapid cooling could be the result of very high rates of subaerial denudation (probably combined with ongoing rapid uplift to drive the denudational exhumation) or it could be associated with tectonic denudation in which very high rates of uplift lead to detachment of slabs of crust which slide away by gravity along decollements, thereby cooling the underlying crust. Sample B has experienced steady cooling from 100 Ma ago to the present. The average track length is shorter (13.5 μm) and the distribution broader (s = 1.99 μm), both measures being reflected in the broader histogram with a longer 'tail' into the short track lengths (reflecting the greater time that tracks have spent in the PAZ after formation). Note how B's fission track age (75.9 Ma) bears no obvious relation to the cooling event or to any particular depth in the crust for determining a rate of denudation. The determination of the rate of denudation requires modelling of the cooling history of the sample (in effect determining the cooling history, as in the upper diagram, from the age and track length distribution in the lower diagram). In the more complex cooling history of C (two discrete cooling events: one between 100 and 90 Ma and the second at about 45 Ma), the fission track age (74.8 Ma) relates to neither cooling event. The track length distribution is broad and bimodal, with the upper mode (long track lengths of c.15 μm) reflecting the 45 Ma cooling event and the lower mode reflecting annealing (shortening) of tracks formed after the first cooling event.

There are several inferential and logical steps involved in converting FTA data to a geomorphological history and an amount of denudation (e.g. Gleadow and Brown, 2000). Notwithstanding the uncertainties and assumptions associated with these steps, FTA has been successfully applied to elucidate long-term landscape development in a range of settings. Application in active orogenic settings of FTA in conjunction with higher temperature thermochronometers, such as the $^{40}Ar/^{39}Ar$ system, and lower temperature systems, such as *(uranium-thorium)/helium analysis* in apatite, has very convincingly demonstrated that denudation of these settings is very rapid. When various thermochronometric systems yield the same rates of denudation, it is argued that there is a dynamic equilibrium between denudation and the ongoing tectonic uplift necessary to drive the flux of crust through the Earth's surface where it is removed by denudation at the same rate as uplift. The processes and sequences of events associated with lithospheric extension and subsequent PASSIVE MARGIN development have also been widely elucidated using FTA. FTA data along many passive continental margins, especially the data from closest to the new margin, exhibit rapid cooling events (long, unimodal track length distributions) at about the time of break-up. These FTA data are interpreted in terms of rapid denudation of the nascent or new continental margin at about the time of breakup, in response to one or more of the following: thermally driven active or passive uplift and denudation of the pre-breakup rift shoulders; rapid denudation of the new margin in response to the new BASE LEVEL for denudation that is provided by the formation of a new ocean basin adjacent to the margin; and ongoing flexural isostatic uplift of the new margin in response to this accelerated denudation.

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PAUL BISHOP

FJORD

A deeply incised trench or trough excavated in bedrock by long-term glacial erosion and occupied by the sea during periods of glacier recession. Spectacular fjordic scenery occurs along the coasts of British Columbia in Canada, Alaska, southern Chile, Greenland, northern and eastern Iceland, Spitsbergen, Fiordland in New Zealand, the Canadian arctic islands and western Scotland. The longest fjords are Nordvestfjord/Scoresby Sund in Greenland (300 km), Sognefjord in Norway (220 km) and Greely Fjord/Nansen Sound in the Canadian arctic (400 km).

Troughs and fjords have distinctive cross and long profiles, referred to as U-shaped but best approximated by the formula for a parabola:

$$V_d = aw^b$$

where w is the valley half width, V_d is valley depth and a and b are constants. However, true cross profiles deviate from this mathematical parabola largely due to the production of breaks in slope by pulsed erosion through time. These effects have been modelled by Harbor *et al.* (1988) by imparting a valley glacier on a fluvial, V-shaped valley. Basal velocities below a glacier are highest part way up the valley sides and lowest below the glacier margins and centre line. By assuming that the erosion rates are proportional to the sliding velocity, the greatest erosion occurs on the valley sides, thereby causing broadening and steepening of the valley. The development of the steep sides of troughs and fjords is aided by PRESSURE RELEASE or dilatation in the bedrock. This is the development of fractures parallel to the ground surface. Such fractures weaken rock masses, thereby facilitating subsequent subglacial erosion. Dilatation is most likely to take place immediately after deglaciation when the glacier overburden has been removed and freshly eroded rock surfaces are exposed.

Overdeepenings along fjord and trough long profiles separated by sills or thresholds appear to represent areas of increased glacier discharge such as at the junctions of tributary valleys or where fjord narrowing occurs. The area of deepest erosion in a fjord marks the location of the long-term average position of maximum glacier discharge. Fjord mouths are often characterized by STRANDFLATS, likely due to the fact that the erosion capacity of the outlet glaciers is severely reduced due to glacier buoyancy and eventual ice flotation in the sea in addition to the flow divergence induced by the more open topography.

The planform of many fjords clearly reveals fluvial or structural origins. For example, the sinuous forms and dendritic patterns of some fjords suggest that they are glacially overdeepened preglacial fluvial valleys and rectilinear fjord networks have been linked to large-scale structural features such as faults and grabens. Moreover, the close association between linear fjord alignments and intersecting lines of regional fracture have led to purely tectonic theories for fjord initiation. The survivial of preglacial landforms and sediments on upland areas between fjords demonstrates that the deep glacial incision is selective, hence the use of the term *selective linear erosion* to describe the development of fjord and trough systems. It is most likely that pre-existing valley systems, whether fluvial and/or tectonic in origin, will contain thicker ice during glaciations and therefore act as major conduits for glacier flow from the centres of ice dispersal, especially if they are oriented parallel to regional glacier flow. Greater ice thicknesses and concomitant preferential ice flow down such valleys will result in greater frictional heat, increased pressure melting and widespread basal sliding. Conversely, on the plateaux between fjords, cold-based ice will dominate and protect underlying preglacial features from glacial erosion. The occurrence of a preglacial land surface on the plateaux surrounding Sognefjord has allowed the calculation of a fjord erosion rate (Nesje *et al.* 1992). Approximately 7,610 km^3 of material has been removed from the fjord by glacial erosion, yielding erosion rates ranging from 102 to 330 $cm\,kyr^{-1}$ depending upon the amount of time that glaciations have dominated the region.

The dimensions of fjords appear to be scaled to the amount of ice that was discharged through them, several researchers having demonstrated that relationships exist between fjord size and glacier contributing area. The strength of these relationships also varies between regions. For example, Augustinus (1992) demonstrated that fjords in British Columbia are 2.5 times deeper and 2.4 times longer than New Zealand fjords even though the contributing areas are comparable in size. This suggests that glacial erosion is more intense in British Columbia probably due to the fact that water depths are shallower offshore than in New Zealand and are therefore less capable of floating the fjord glaciers. In addition, the lengths of the former British Columbia fjord glaciers were much greater than those of the New Zealand palaeo-glaciers, the former having been nourished by an inland ice sheet.

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DAVID J.A. EVANS

FLASH FLOOD

Flash flood denotes an abrupt rise in the discharge of a river or stream, providing an event of short duration. The term has conventionally been associated with ephemeral flow regimes in which the majority of events are rain-fed. The flood is discrete. It impinges on a channel bed that is initially dry and is exhausted within a short interval – a few hours in the case of small drainage basins or a few days where basin size involves longer travel time. Because of this, flash floods are commonly associated with deserts and semi-deserts of low to middle latitudes, flow in high-latitude deserts resulting rather from the slow release of water in the form of snowmelt – a seasonal freshet lasting continuously for weeks or months. However, the term has been used more widely to describe a sudden significant increase in discharge where the annual flow regime is intermittent or even perennial. In environments such as those with a Mediterranean-type climate, flow dwindles seasonally so that flood runoff in summer may add dramatically to a pre-existing trickle. In these circumstances, the perception of an observer will be that the rising limb of the flood hydrograph is steep, occupying tens of minutes rather than hours, a pattern which contrasts with that more typical of runoff during the wet season. To warrant the descriptor, the magnitude of the flood peak will also have been remarkable in causing nuisance or damage, or, indeed, loss of human life. It is debatable whether, here, the flood hydrograph should be described as 'flashy' or 'flashier' in relation to the norm rather than given the epithet 'flash flood'.

In desert or semi-desert environments, where dry channel conditions are the norm, (see WADI) flash floods are usually remarkable regardless of magnitude. Here, in contrast with Mediterranean or temperate environments where events of this type are a feature of the 'dry season', floods are more often than not associated with incursions of monsoonal airmass (as in the Sonoran Desert and the Saudi Arabian peninsula) or with the regular latitudinal shift of the Inter-Tropical Convergence Zone (as in East Africa). In such areas, these are events of the rain season(s), but rainfall is extremely uncertain, making events even more memorable if, by chance, they are witnessed. Although widespread, low intensity rainfall can generate runoff in these areas, flash floods are more likely to be the product of wandering cellular convective storms. The wetted 'trail' or footprint of these is usually only a few kilometres across. Atmospheric dynamics dictate that such storm systems that are capable of releasing rain of sufficient magnitude and intensity will be separated by several tens to several hundreds of kilometres. The likelihood that a drainage basin will receive sufficient rainfall to generate runoff (see RUNOFF GENERATION) depends upon its position in relation to the trajectory of each storm. Small basins (in the order of $10^1\,km^2$) may experience an event only infrequently, perhaps staying dry even though floods occur in the vicinity. In this case, it may be that a flood series for one basin is developed from events in years that do not contribute to the series of a neighbouring stream. In basins of moderate size (several $10^2\,km^2$), the storm cell is frequently smaller than the basin. Indeed, in this case, the flood may move down-channel into parts of the basin that have not experienced rain. This is a circumstance that provides the greatest danger for the unwary and is not uncommonly the cause of human mortality, especially where the dry river bed has provided an apparently convenient location for overnight encampment.

The significance of the variable spatial coincidence of storm and drainage basin is that the flood hydrograph can take on a variety of shapes. This is, in part, because different sub-basins may contribute to each event and storms may move up or down catchment, depending on local atmospheric dynamics, so affecting the gathering time of contributions from each tributary. This means that it is more difficult to define a typical flood hydrograph (as in, e.g. unit hydrograph analysis) in a desert or semi-desert setting, not only because the frequency of events is low but also because each runoff event may possess unique characteristics. There is, however, evidence from one drainage basin in the Asir Escarpment of Saudi Arabia that flood volume can be approximated from a parameter such as flood peak discharge, and finite element models of rainfall runoff have been developed with some success for predicting flood waves in small basins in Oman and Arizona. Despite these, the relations between runoff and storm characteristics such as rainfall amount and intensity are often chaotic so that predictability of event frequency and magnitude is low even if rainfall is being monitored by spatially inclusive means such as radar. In all but rare instances where research catchments have been established, rain gauge density and disposition will be either inadequate or, more often, non-existent.

Flash floods are undoubtedly dramatic, if only because of the stark contrast between the event itself and the much longer intervening period when the channel is dry. In southern Israel, at the eastern edge of the Sahara's hyper-arid core, long-term monitoring has revealed that the frequency of events is, on average, much less than one a year, but there can be periods of several years when no runoff occurs. In the semi-arid northern Negev, with a rainfall of *c*.200–300 mm per year, the number of events that occurs in moderate-sized basins ranges from zero to seven. Here, on average, an ephemeral channel is occupied by flow for about 2 per cent of the year, or about seven days.

Perhaps the most dramatic aspect of flash floods is the arrival of a bore. This may be the first that the observer is aware of rainfall, which may have occurred well up-catchment. Field monitoring in semi-arid areas, where vegetation may be sparse at the start of the rain season, has shown that time to ponding is short – typically in the order of a few minutes, depending on the infiltration capacity of the local soils – even under modest rainfall intensities. An observer caught out in the rain undergoes a curious sensation that the ground is moving as a glistening sheet of OVERLAND FLOW slips towards the channel system. Here, high drainage densities, developed in response to the easy and quick shedding of water, ensure rapid concentration of flow and the birth of a flash flood.

A flash flood bore takes on a number of forms. The rapidly advancing 'wall of water' is almost certainly a figment of imagination encouraged by the panic of moving to a place of safety. Indeed, the type of bore most commonly caught on camera is comparatively shallow, with low trailing water-surface slope. However, a few examples have been photographed where the bore reaches a height of about half a metre (Plate 49). Of those few measurements that have been made of bore advance, velocities range from 0.5 to 2 m s^{-1}, the rate depending directly on bore height. This is equivalent to a stiff walking pace for a human being and one might wonder, therefore, what reasons there are for the number of fatalities that are reported. The problem for those unfortunate to be caught napping is that, following the passage of the bore, water levels rapidly increase. One fully documented example has indicated an average rise of a quarter of a metre per minute, so that the water surface was at waist height within two and head height within ten minutes of the start of hydrograph rise. By this time, average flow velocity is in excess of 3 m s^{-1} and increasing to values greater than 5 m s^{-1} (Figure 62).

Plate 49 Flash flood bore in Nahal Eshtemoa, northern Negev, advancing over dry bed at about 2 m s^{-1}. Note that the immediate area has had no rain

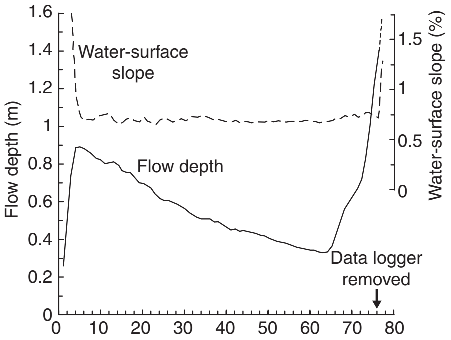

Figure 62 Hydrograph and water surface slope of the flash flood on Nahal Eshtemoa shown in Plate 49

However, a flash flood bore is not stealthy. If it advances over a gravel bed, the cacophony of grains being thrown against each other can be heard several hundred metres away. Flotsam is also characteristic of flash flood bores, the infrequent flow sweeping up LARGE WOODY DEBRIS and other organic matter that has fallen into the channel between events and adding to the general confusion that is already inherent. Indeed, some have reported hearing the clash of tree trunks, etc. several kilometres ahead of the bore's arrival.

Although the bore of a flash flood is its most spectacular feature, another unique but hidden characteristic is the loss of a significant fraction of flow to the dry bed. These are dubbed transmission losses. They are determined in part by the magnitude of the flow and hence the wetted perimeter. For a small heavily gauged ephemeral channel in Arizona, examples show that transmission losses to the bed in each kilometre of channel can account for as much as 6 per cent of the flow. This points to another important characteristic of flash floods in desert and semi-desert settings – many fail to reach the terminal ALLUVIAL FAN.

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IAN REID

FLAT IRON

Term used to designate relic slopes whose morphology resembles a reversed iron. They are also known as talus flat irons or triangular slope facets and develop at the foot of scarps in mesas and cuestas. The slope deposits, which locally contain datable charcoal, ashes or pottery remains, may grade in the distal sector to cover pediments or fluvial or lacustrine terraces.

The most widely accepted genetic model relates the development of talus flat irons to climatic changes. The accumulation processes in the slopes prevail during humid periods whereas the reduction in the vegetation cover during dry periods favours rilling and gullying processes. Successive climate changes give place to different generations of relict slopes whose relative chronology can be inferred from their spatial distribution. Up to five generations of flat irons have been identified in the three main Tertiary basins of Spain. The slope deposits of the flat irons dated with ^{14}C correspond to cold periods. The youngest facet generation fits with Upper Holocene Neoglaciation episodes and the two previous generations correlate to Heinrich events (H_3 and H_4) that indicate cold periods (Gutierrez *et al.* 1998).

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SEE ALSO: slope, evolution

M. GUTIERREZ-ELORZA

FLOOD

A flood is a flow of water greater than the average flow along a river. A flood may be described in terms of its magnitude. On a given river, for example, any discharge exceeding $1{,}000\,m^3/s$ might be designated a flood. A flood may also be described by its recurrence interval; a 100-yr flood occurs on average once every 100 years. Or a flood may be described as any flow that overtops the banks or LEVEEs along a channel and spreads across the FLOODPLAIN.

Floods along inland rivers result from precipitation or from a damburst. When water rapidly flows downslope from snowmelt, rain-on-snow, or various types of rainfall, the baseflow from subsurface water that keeps some stream channels flowing during dry periods is augmented by runoff (see RUNOFF GENERATION). As discharge increases in the channel during the rising limb of the flood, the channel boundaries may be eroded, and both SUSPENDED LOAD and BEDLOAD sediment transport are likely to increase. Once the input of runoff to the channel declines, sediment transport is likely to decrease and sediment may be deposited along the channel during the falling limb of the flood. Floodwaters in a channel commonly rise more rapidly than they fall during all types of floods, but this difference is most pronounced for damburst floods. Dams built by humans and naturally occurring LANDSLIDES, ice jams, glacial moraines, glacial ice dams, or beaver DAMS may fail suddenly, prompting catastrophic drainage of the water ponded behind the dam. Along with FLASH FLOODS, damburst floods are often the most unexpected and damaging floods. Damburst floods may have a peak discharge more than an order of magnitude larger than the peak discharges created by meteorological floods along a river. This large discharge may generate high values of STREAM POWER that cause substantial erosion and deposition along the flood path. OUTBURST FLOODS generated by the failure of natural dams ponding meltwater from the great continental ice sheets during the Pleistocene shaped such dramatic landscapes as the Channeled SCABLAND of the northwestern United States.

Floods along coastal rivers may also result from STORM SURGES, TSUNAMIS, or other anomalously large waves or tides backflooding upstream from the ocean. Low-relief coastal areas such as those found in Bangladesh may be particularly susceptible to such floods.

The largest measured historical floods generated by precipitation have occurred primarily between 40°N and 40°S latitude, usually near coastal areas where the onshore movement of warm, moist airmasses into the continental interior produces intense and widespread precipitation (Costa 1987). The envelope curve of maximum rainfall-runoff floods is mathematically described by $Q = 90A$ for drainage areas less than 100 square kilometres, and $Q = 850A^{0.357}$ for larger drainage areas, where Q is peak discharge in cubic metres per second and A is drainage area in square kilometres (Herschy 1998).

The importance of a flood relative to smaller flows in shaping channel and valley morphology will depend on the magnitude and duration of the hydraulic forces generated during the flood in comparison to the erosional resistance of the channel boundaries, and on the recurrence interval of the flood. A channel formed on bedrock or very coarse alluvium may have such high boundary resistance that only a flood generates sufficient force to erode the channel boundaries. This effect may be enhanced where a deep, narrow channel and valley geometry concentrate floodwaters such that flow depth increases rapidly with discharge, giving rise to high stream power. In contrast, a channel bordered by a broad floodplain will have much less increase in flow depth with increasing discharge, and the flood may not have a substantially greater capacity for erosion and sediment transport than do smaller flows along the channel. Channels in which geometry and sediment transport reflect primarily floods are likely to have a flashy hydrograph, abundant coarse sediment load, a high channel gradient, highly turbulent flow, and shifting, erodible banks (Kochel 1988). Geomorphic change during floods is likely above a minimum threshold (see THRESHOLD, GEOMORPHIC) of approximately $300\,\mathrm{W\,m^{-2}}$ of unit stream power for alluvial channels (Magilligan 1992). The threshold for bedrock channels may be expressed as $y = 21x^{0.36}$, where y is stream power per unit area and x is drainage area (Wohl *et al.* 2001). In steep channels with abundant sediment, flows may alternate downstream among water-floods, DEBRIS FLOWS and HYPERCONCENTRATED FLOWS.

Measures to reduce hazards to humans associated with floods date back several millennia. Such measures include impoundments to regulate water flow; channelization to increase the flood conveyance of channels; levees to confine floodwaters; warning systems to help alert and evacuate humans at risk; and engineering designs which reduce flood damage to structures. Despite this long history of river engineering and flood mitigation, property damage from floods continues to increase worldwide as population density and building in flood-prone areas increase, and as land uses across drainage basins alter runoff generation. Along rivers where alteration of the natural flow regime has reduced or eliminated floods, aquatic and riparian species adapted to flooding have declined in extent and diversity. River rehabilitation and restoration measures are now being applied to some of these rivers in an attempt to mitigate damages caused by the absence of floods.

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SEE ALSO: bankfull discharge; floodout; palaeoflood; sediment rating curve

ELLEN E. WOHL

FLOODOUT

A floodout is a site at the downstream end of a river where channelized flow ceases and floodwaters spill across adjacent, unchannelled, alluvial surfaces. The term has been most widely used in connection with ephemeral channels in arid central Australia (Tooth 1999a) but it has also been applied to discontinuous gullies (see GULLY), intermittent channels and perennial channels in semi-arid, subhumid and humid regions of eastern Australia and southern Africa (Fryirs and Brierley 1998; Tooth *et al.* 2002).

Floodouts form as a result of various factors including downstream decreases in discharge, downstream decreases in gradient, and aeolian or bedrock barriers to flow (Tooth 1999a). These factors commonly act in combination. For example, along many arid or semi-arid rivers, discharge decreases downstream owing to factors such as infiltration into normally dry channel beds, evaporation, hydrograph attenuation and a lack of tributary inflows. Gradient also commonly decreases owing to channel-bed AGGRADATION or lithological/structural factors, such as a change from a harder to a weaker lithology underlying the channel bed. In combination, these discharge and gradient decreases mean that unit STREAM POWER and sediment transport capacity also decrease, which in turn leads to a downstream reduction in the size of the channel and diversion of an increasing proportion of floodwaters overbank. This is often exacerbated by the presence of aeolian or bedrock barriers, such as longitudinal dunes (see DUNE, AEOLIAN) that have formed across the river course. Eventually, the channel loses definition and disappears entirely, and the remaining floodwaters spill across the floodout as a sheet flow (see SHEET EROSION, SHEET FLOW, SHEET WASH). This process is often referred to as ‘flooding out’ but strictly speaking the term ‘floodout’ and its derivatives should be used for the fluvial form only.

Floodouts can form in river catchments of widely different scale and thus the areas of floodouts vary considerably (*c*.1–1,000 km^2). The location and shape of floodouts, however, are often strongly influenced by local physiography. In central Australia, for instance, floodouts in the northern Simpson Desert are narrow (<500 m) features where rivers terminate between longitudinal dunes and occasional bedrock outcrops but on the relatively unconfined Northern Plains they can reach up to several kilometres wide (Tooth 1999a,b).

‘Floodout zone’ is a related but broader term that encompasses both the lower reaches of the channel and the floodout itself. Geomorphological and sedimentary features commonly associated with floodout zones include distributary channels, splays, waterholes, PANs, PALAEOCHANNELs and various fluvial-aeolian interactions (Tooth 1999a,b). In addition, two basic types of floodout can be distinguished (Tooth 1999a): (1) terminal floodouts, where floodwaters spill across the unchannelled surfaces and eventually dissipate through infiltration or evaporation; and (2) intermediate floodouts, where floodwaters persist across the unchannelled surfaces and ultimately concentrate into small ‘reforming channels’. Reforming channels commonly develop where the unchannelled floodwaters become constricted by aeolian deposits or bedrock outcrops, or where small tributaries provide additional inflow, and they either join a larger river or decrease in size downstream before disappearing in another floodout (Tooth 1999a; Tooth *et al.* 2002).

Formation of a floodout is just one possible end result of the broader processes of channel ‘breakdown’, ‘failure’ or ‘termination’ that can also occur where channels disappear in playas, in permanent wetlands, or on the surfaces of ALLUVIAL FANs. Floodouts, however, are predominantly alluvial features which are normally dry except after flood events or heavy local rains, and thus they differ from saline playas or organic-rich, saturated wetlands. Furthermore, the relatively low gradients (<0.002) and fine-grained deposits typical of floodout zones distinguish them from alluvial fans. Floodout zones have many geomorphological and sedimentological similiarities with ‘terminal fans’, a term that has been applied to the distal reaches of some inland arid and semi-arid river systems where numerous distributary channels decrease in size downstream and grade into unchannelled, fan-shaped deposits (Mukerji 1976; Kelly and Olsen 1993). Downstream of intermediate floodouts, however, channels can reform, thus showing that floodouts are not necessarily terminal and, as floodouts are often confined laterally by aeolian deposits or bedrock outcrop, neither does alluvial deposition necessarily adopt a fan-shaped form (Tooth 1999a). As such, application of the term 'terminal fan' is inappropriate for many floodouts or floodout zones. In floodout zones, the disappearance of channelized flow means that FLOODPLAINs grade downstream into floodouts. Although the absence of channels makes it difficult to include floodouts within conventional definitions of 'floodplain' or existing floodplain classifications, nevertheless they can be regarded as part of a continuum of floodplain types (Tooth 1999b).

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SEE ALSO: alluvium; bankfull discharge; flood; hydraulic geometry

STEPHEN TOOTH

FLOODPLAIN

The floodplain is generally considered to be the relatively flat area of land that stretches from the banks of the parent stream to the base of the valley walls and over which water from the parent stream flows at times of high discharge. The sediment that comprises the floodplain is mainly ALLUVIUM derived from the parent stream with minor contributions from aeolian sediment or colluvium from the valley walls. During floods the channel width is increased to include some or all of the floodplain in order to accommodate the increased discharge with relatively smaller increases in velocity and depth than would be the case if the flood discharge were artificially confined within the channel. However, defining the extent of a floodplain at a locality in terms of the area inundated in floods of particular return periods poses problems, since flooding frequency may be a restricting factor. This can be especially problematic in arid and semi-arid areas.

It has been suggested (Wolman and Leopold 1957) that the active floodplain is the area subject to the annual flood (i.e. the highest discharge each year), though this can really only apply to rivers in humid regions. In reality, the active floodplain only forms part of the topographic floodplain, which encompasses the whole valley floor and includes parts of relict floodplains in the form of river terraces (see TERRACE, RIVER) (Plate 50). If the floodplain is defined in terms of the processes (including superfloods) that give rise to it, then the term polygenetic floodplain would apply to most since they result from changes in flow-regime and sediment supply over at least the recent geological past. Nanson and Croke (1992) have proposed the term genetic floodplain, which applies to a generally horizontally bedded landform built from alluvium derived from the present flow-regime of the adjacent stream. This does not take into account the geomorphic history of a floodplain and the processes that have influenced its construction over time, however. A floodplain is a functional part of the whole stream system and forms as

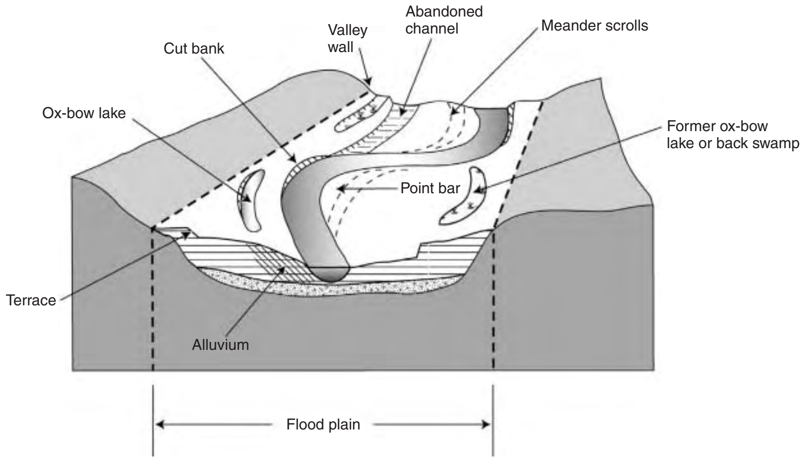

Plate 50 Features of floodplain topography

a byproduct of interrelated processes that, over time, give rise to variable flows and sediment loads derived from the drainage basin.

Floodplain formation

Floodplains are formed by processes that are active both within the channel of the parent stream and during overbank flow. The processes involved are lateral accretion, which takes place within the channel from the formation of bars by movement of relatively coarse bedload; and vertical accretion, which occurs on the floodplain surface due to deposition of finer material from the suspended load during overbank flow. The relative importance of vertical accretion in the formation of floodplains was considered negligible in comparison with within-channel processes (Wolman and Leopold 1957) though it has now been shown that overbank sedimentation can contribute significantly. For example, Nanson and Young (1981) have described the floodplain deposits of streams in New South Wales that have parts of their floodplains dominated by extremely cohesive overbank deposits that prevent the stream from migrating. In lowland rivers in the United Kingdom similar deposits have been described for the Severn (Brown 1987) and the Thames (Lewin 1984) where thick muddy deposits overlie sandy gravels.

Lateral accretion deposits are built up in the channel either as marginal bars that may form in an alternating sequence along relatively straight channels or as point bars that develop on the inside bends of meanders. If the channel migrates laterally by BANK EROSION on one side, the channel dimensions are maintained by compensating deposition on the other bank. The marginal or point bars grow laterally towards the direction of migration and increase in height, with sediment being deposited on them in low-angled, sloping layers overlain by finer material deposited at bankfull and flood stage flow. The sediments within the bars tend to fine upwards as they initially form from the stream's bedload sediment carried by secondary currents from the outer bank region towards the inner bank (Markham and Thorne 1992). Over time, in this way, a stream may rework the entire floodplain sediment as it migrates from one valley side to the other, leaving behind cutoff meanders as oxbow lakes or swamps and traces of the old meander paths as meander scrolls (Plate 50).

Vertically accreted sediment is added to the floodplain surface during overbank flow. As the water in the channel overflows the banks onto the floodplain surface, the flow velocity is reduced as the width of the channel is effectively increased by inclusion of the floodplain. This reduction, in conjunction with an increase in surface roughness if the floodplain is vegetated, causes suspended sediment to be deposited. As the flow of water on the floodplain is slower and shallower than that in the channel, there is a zone of turbulence near to the channel bank which results in a net transfer of momentum and sediment from the channel to the floodplain. The width of the turbulent zone depends on the relative difference in depths between the channel and floodplain flows but influences the nature of sediment deposition near to the channel. The sediment deposited near to the channel bank tends to be coarser and thicker than that deposited further onto the floodplain as transport competence declines with distance from the channel and away from the influence of the turbulent zone (Marriott 1996). However, the amounts and grain sizes deposited depend on sediment supply and duration and depth of overbank flow.

The channel banks can gradually become the highest points on the floodplain as the thicker, coarser overbank deposits build up to form natural levees. The stream may then deposit sediment on its bed during high flows that do not exceed BANKFULL DISCHARGE, resulting in the normal river surface being above the level of the floodplain. Both natural levees and artificial embankments set away from the channel as macrochannel banks to a two-stage channel, afford some protection from flooding. In extreme cases, floodwater may break through the levee, forming a crevasse channel and washing sediment from the channel and reworked from the levee onto the floodplain. These sediments form a crevasse splay of coarser material than the underlying floodplain alluvium. Coarse material can also be transferred from the channel during overbank flow due to the action of convection currents set up by turbulence at bends in the channel. This is because the flow of water on the floodplain tends to travel directly down valley and at meanders the direction of flow within the channel is at an angle to that on the floodplain (Knight and Shiono 1996). Bedload sediment can then be picked up and spread in a lobe downstream from the outer bank of the bend.

In arid and semi-arid areas floodplains are formed mainly during major flood events (superfloods) with recurrence intervals in the region of 10,000 years. These floods bring in material from highland areas of the catchment. Between these events the sediment is reworked by the more frequent relatively minor flash floods that occur in the ephemeral channels rather than new material being added. Studies of the streams of central Australia (Pickup 1991) show different landforms depending on the scale of flooding that gave rise to them. The superfloods result in large sandsheets and sand threads and the contemporary macrochannel system which has levees, FLOODOUTs and floodbasins wherein sediment is reworked during flooding or by aeolian processes in dry periods.

Floodplain classification

Floodplains can be classified according to their morphology rather than the manner in which they were formed or the processes active at present. The floodplain as a whole, together with its parent stream, can be considered as a macroform, whereas the mesoforms are the component parts of the channel, e.g. bar forms, and the floodplain (levees, crevasse channels and splays, backswamps and oxbow lakes). Mesoforms can influence flow and deposition patterns on the macroform. In this classification the microforms are the small-scale structures superimposed on the mesoforms, for example, ripples, dunes, shrinkage cracks (Lewin 1978).

A further classification that takes into account the genesis of the floodplain was suggested by Nanson and Croke (1992) and is based on the relationship between the ability of streams to entrain and transport sediment and the resistance to erosion of the bank sediment. Three classes are recognized: high energy streams with non-cohesive banks, medium energy streams with non-cohesive banks and low energy streams with cohesive banks. Within these classes further levels of classification can be set up based on primary geomorphic factors such as channel cutting and filling and lateral accretion on point bars, and secondary factors such as scroll bar formation and organic (peat) accumulation. The primary geomorphic factors depend on stream power and sediment load and can therefore identify different environments for floodplain formation.

Floodplains respond to changes in channel processes that result from alterations in flow regime and/or sediment supply (Schumm 1977) though the response may be at a slower rate. The sedimentary record of these changes is often incomplete as it is often complicated by episodes of erosion, so although floodplain formation is polygenetic, much of the evidence is destroyed. It could be suggested that, over geological timescales, all floodplains are polygenetic as external influences such as climate and relative base-level change are not constant.

Floodplain sedimentation rates

In humid areas rivers flood every 1–2 years, though it is mainly extraordinary events that are studied and documented due to their catastrophic effects on human activity and property. Studies of flood frequency and sedimentation patterns have, therefore, been carried out with the aim of attempting to predict, and thus avoid, the destructive effects of extreme events. As explained above, the floodplains of many rivers are composed of sediments accumulated from channel and flood activity and many studies record the thickness of deposits in various parts of the floodplain by extraordinary events or they give the nature of the sediment deposited. Generally figures are highly variable and are only useful as a general guide to likely sedimentation rates in the various floodplain environments indicated.

Flooding is an essentially random occurrence and it is difficult to sample satisfactorily from the floodplain surface during a flood. However, borehole data, sediment cores, ^{14}C dating, pollen analysis and radioactive nuclides (e.g. ^{137}Cs and ^{210}Pb) can all be used to estimate sedimentation rates over a period of time. As an example, Brown (1987) used some of these techniques to estimate that sedimentation rates on the River Severn floodplain in the UK over the past 10,000 years have been around 1.4 mm per year.

Floodplains act as storage space or sediment sinks for alluvial sediment. While they are being stored the sediment may be reworked by fluvial, aeolian, biological and/or pedological agents, often over considerable periods of time. The stored sediment may subsequently be eroded and re-incorporated into the sediment budget of the drainage basin. The residence time of sediment in storage will vary according to factors such as surface topography, climate and vegetation and the relative return frequency of major flood events. Originally, sediment storage on floodplains was studied so that SEDIMENT BUDGETs of drainage basins could be calculated; now, however, interest in the storage of sediment that was originally part of the suspended load of the parent stream has increased with an awareness of the ability of contaminants such as heavy metal ions and radionuclides to adhere to and be transported with this fine material.

Summary

Understanding the processes involved in floodplain formation is important because of the interaction between human activity and floodplain environments. These processes include both within-channel and overbank processes which rely on the interaction between channel and floodplain flows during flooding, and which account for the distribution of different sediment grain sizes across the floodplain. Some recent work has investigated the rate of sedimentation and sediment storage on floodplains using a multiproxy approach. As floodplains act as sinks for alluvial sediments, this work is particularly useful for studies of contamination.

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SUSAN B. MARRIOTT

FLOW REGULATION SYSTEMS

Flow regulation systems is a general expression including all works constructed along a watercourse in order to regulate the flow in the channels. Flow regulation systems can be planned for maintaining steady-state conditions along a river or for avoiding uncontrolled erosion and/or sedimentation processes. On the other hand, they can be planned for obtaining a more constant distribution of the discharge in the channel by reducing both highwater and minimum flow peaks.

The devices utilized for avoiding accelerated erosion or deposition in the watercourse consist of hydraulic works capable of maintaining flow velocities within suitable ranges of values, which should be adequately chosen for each river. Besides this, a whole set of other measures are planned and constructed for opportunely directing the flow stream in order to avoid the occurrence of maximum velocities in proximity of the banks (see BANK EROSION). The most commonly used remedial measures consist in the direct protection of the banks by means of walls, sheet piles, prefabricated structures, gabions, anchored geotextiles, loose debris and biofixing plants (e.g. planting cuttings, sowing herbaceous species, etc.).

Groynes are typical works which direct the water flow; they are made up of structures stretching both downflow and with a certain angle with respect to the mean flow and can be either linear or composite (such as sledge hammers or bayonet joints). Furthermore, various arrangements of boulders in the river bed are works which increase friction by reducing velocity and increasing turbulence.

Finally, check dams are works which reduce excessive riverbed slope profiles. They break the river's profile into several stretches with lower inclinations, fixing the level of non-erodable weir crests and introducing artificial steps in the longitudinal profile of the watercourse. Besides being constructed with very heterogeneous materials, check dams can be of various patterns and dimensions. When they produce a real impoundment upstream they should rather be considered proper DAMs. To this regard, it should be mentioned that in the Alps many dams constructed for hydroelectric purposes work also as reservoirs for the regulation of water discharge during considerable highwater or minimum flow events. Apart from the regulation of flow, dams across great rivers usually have multipurpose functions, such as production of electric energy, navigation control, irrigation supply and flood control (Jansen *et al.* 1979).

Check dams are used also in watercourses capable of transporting and depositing large amounts of sediments (see BEDLOAD). In this case, they retain a certain amount of sediment in the upper part of the basins and reduce solid transport during highwater and minimum flow phases. Since the early 1950s various kinds of open weirs have been planned and constructed; among them, the most used ones are fissure-weirs and filtering weirs. The latter can be comb-like, with windows, network patterned, etc. (Figure 63). These particular check dams accomplish a two-fold purpose: (1) to retain most of the sediment during highwater phases and release it later during a low-flow phase; (2) to retain the coarse debris (including floating tree-trunks etc.) in order to avoid damage to the hydraulic structures downstream. Since the late 1980s check dams with low-angle filtering intake accompanied by a drainage gallery have been constructed along watercourses subject to overconcentrated flows ascribable to DEBRIS FLOWs.

In order to regulate the flow rate distribution during the year, other kinds of works are utilized. Among these, flood attenuation basins should be mentioned. They consist in hydraulic works which connect the watercourse to sufficiently large artificial basins capable of working as retaining reservoirs during river spates. These flood protection structures, also defined as flow-control weirs with energy dissipators, consist of a downstream regulation dam which allows the passage of a flow not superior to a prefixed value,

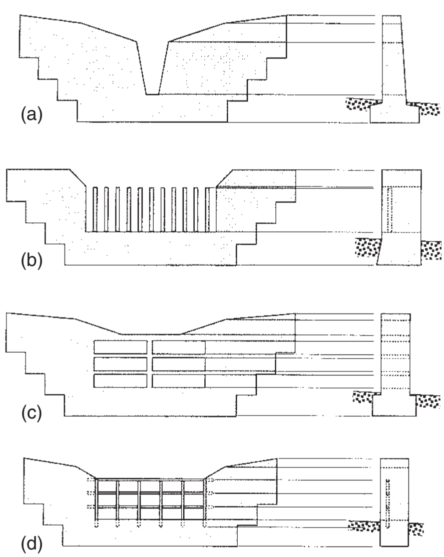

Figure 63 Examples of (a) fissure-weirs; and filtering weirs; (b) comb-like; (c) with windows; (d) network patterned

even in the occurrence of a greater flow. The exceeding volume of water is stored in the basin and returned to the river when the highwater phase is dwindling (Bell and Manson 1998). Flood protection structures can also be equipped with an upstream check dam in order to avoid discharge water from entering the flood attenuation basin when flow rate is below a certain average value. In this way the sedimentation of the bedload usually transported is avoided and, consequently, so is the progressive reduction of the basin's retention capacity. For example, along the central course of the Yangtze River, once this watercourse started to wander in its alluvial plain, a flood control weir was constructed which allows highwater flow to be diverted into a large dissipator basin provided along one of its tributaries (Jingjiang River and Dongting Lake). These hydraulic facilities, built in 1953 and extended in 1990, are made up of a 54-hole barrage, 1,054 m long, and can store up to $8{,}000\,m^3\,s^{-1}$ of water in order to prevent inundations in the stretch located immediately downstream of the catchment works, thus protecting the town of Wuhan and the plain of Janghan from flooding.

Fillways and diversion canals are natural or, more often, artificial watercourses which receive a portion of a river's highwater in order to divert it into another basin or give it back to the same river, downstream of a critical stretch. Diversion canals are characterized by constant flow whereas fillways are utilized only occasionally. In the case of a diversion canal which subtracts water from a river and gives it back downstream of a critical point, usually an inhabited centre, the bypass canal should be long enough to avoid impoundment problems (raised hydrostatic levels due to a return effect starting from the confluence).

Other works which effect a reduction in flow rate levels, consist in modifications of the geometrical features of the river bed (see HYDRAULIC GEOMETRY). These solutions should be implemented with care as they are the result of engineering viewpoints and seldom do they take into sufficient account the geomorphological features of the canal. The usual procedure involves deepening, widening and redressing the canal. In the case of reshaping and widening of a canal's section, a decrease of the stream velocity takes place with consequent sedimentation and rising of the river bed which, therefore, requires periodical dredging. Other cases of course modification consist in canal straightening. For example, the best known cases of straightening are those in the lower course of the Mississippi River, which were carried out during the 1930s and 1940s by means of cutoffs of the meandering course for a total length of 210 km. The consequence of these modifications was an average increase of the riverbed gradient (Winkley 1982) which contributed to an increase of bedload, a rise of the canal's width/depth ratio and a tendency to change from a MEANDERING course to a braided (see BRAIDED RIVER) one. Thus, navigation problems arose due to the decreased depth of the river bed and, as a consequence, expensive dredging and bank protection works were necessary in order to mitigate the problem which still persists in this stretch of the river.

Among the works which regulate flow, artificial embankments should also be mentioned; they are the first fluvial works of a certain entity ever realized by humans. Indeed, it is well known that the construction of considerable embankments in the Po Plain, in northern Italy, started during the Roman age, although traces of partial embankments date back even to the previous Etruscan epoch (Marchetti 2002).

In the alluvial plains of economically advanced regions, flood protection works are connected to dense canal networks with draining and irrigation purposes which allow intensive farming activities in the plain areas and, at the same time, reduce the risk of flooding in urban areas.

Extreme cases of flow regulation systems consist in the implementation of real changes of the hydrographic network of a region by means of artificial diversions of important watercourses. In the ex-Soviet Union, in a vast territory characterized by north- and east-bound large rivers which flow through largely infertile cold regions and extensive drought-stricken areas, impressive diversions were planned and partially implemented. For example, the southward diversions of the rivers Peciora and Irtys, carried out in order to avoid the drying up of Lake Aral following the heavy water exploitation of Syrdarja and Amudarja, has produced, on the one hand, the drying up of vast northern territories covered by taiga and, on the other hand, the swamping of large areas to the south.

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SEE ALSO: flood; fluvial erosion quantification; river restoration

MAURO MARCHETTI

FLOW VISUALIZATION

The human eyes have some difficulty in perceiving the displacement of air, water and of most fluids. However, looking at fluid interfaces (surface boils in river) or particles suspended in a fluid (such as snow flakes or air bubbles) aids in the recognition of structured bodies within the moving fluid. Flow visualization provides a set of tools that have led to significant advances in our understanding of fluid dynamics. Flow visualization tools allow us to track and follow individual turbulent flow structures as they develop (Lagrangian reference frame) or to define flow parameters at one specific location within the flow, such as flow recirculation boundaries (Eulerian reference frame). As a result, these tools have become indispensable in the study of the structured motion of fluids on and within the globe's surface.

Sketches of turbulent flows made by Leonardo da Vinci (1452–1519) demonstrate that flow visualization has long fascinated the imagination of scientists. However, in order to complement quantitative flow monitoring, flow visualization techniques were developed primarily in laboratory studies of fluid mechanics during the last half-century. In the later 1960s, the use of flow visualization led to a major breakthrough in the understanding of turbulent boundary layer structure. Kline *et al.* (1967) (using air bubbles) and Corino and Brodkey (1969) (using neutrally buoyant particles) showed that the flow near a boundary, albeit turbulent in nature, exhibit structured patterns and mechanisms. Today, a wide range of visualization techniques is routinely used in laboratory studies (see the *Atlas of Flow Visualization* and the proceedings from several *International Symposiums on Flow Visualisation* for in-depth reviews of techniques and results from wind tunnels and water flumes experiments).

Most flow visualization techniques rely on the presence of a foreign tracer in the flow. These are often suspended particles or dye/smoke injected at specific locations. The use of tracers relies on the fact that the fluid motion can be inferred from the tracer movement matching that of the fluid. This implies a clear understanding of the mode of introduction or generation of the tracer in the flow, of the relationship between tracer and fluid motion, and of the physical significance of the observed tracer motion. In highly turbulent flows, cameras are used to aid the interpretation of flow patterns from moving particles. The use of dye/smoke present the advantage of providing a quick expression of the flow structure, but can diffuse rapidly away from the source of injection and, thus, reduce the tracing distance.

As well as in laboratory studies, flow visualization techniques can be used in aeolian, fluvial and costal environments where tracers often naturally occur. Matthes (1947), for example, provided an extensive classification of flow structures found in a river based on visual observations of surface boils, waves, turbidity differences and other visual indices. Roy *et al.* (1999) described two flow visualization techniques used in the natural environment. The first uses the turbidity difference to describe the development of flow structures at the shear layer between two merging streams. The other involves the injection of a milky white fluid to visualize shedding motions from the recirculating flow region in the lee of the obstacle.

Two different approaches of *quantitative flow visualization* exist. The first involves sampling velocity over a dense grid using single point velocity meters, such as electromagnetic current meters or acoustic Doppler velocimeters. This grid is then used to create maps of the turbulent parameters of interest (Bennett and Best 1995). As the measurements are not taken simultaneously, this approach provides a frozen picture of the general flow patterns. The second approach it to take velocity measurements using several single probes simultaneously. Such a set-up allows space–time velocity matrices to be created from which space–time velocity coherence and footprints of flow structures can be observed and described within the region covered by the velocity metres (Buffin-Bélanger *et al*. 2000).

The increasing use of multi-point velocity measurement techniques, such as acoustic Doppler profiling (Wewetzer *et al*. 1999) and particle image velocimetry (Bennett *et al*. 2002), is bound to create new breakthroughs in our understanding of flow structure. These techniques rely on the measurement of velocity from embedded particles in the moving fluid and allow the temporal variability of the flow to be described quantitatively at one point as well as the spatial variability of flow patterns in time. Hence, these techniques combine the qualitative realism of flow visualization with quantitative velocity measurements.

Our ability to use computational fluid dynamics (CFD) to simulate complex flows is increasing dramatically. This improvement gives rise to more and more sophisticated *numerical flow visualization* (Lane *et al*. 2002). Traditional flow visualization also complements CFD in allowing us to compare numerical results to natural flow behaviour.

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SEE ALSO: boundary layer

THOMAS BUFFIN-BÉLANGER AND ALISTAIR D. KIRKBRIDE

FLUIDIZATION

A geomaterial becoming a fluid, behaving like a fluid, usually associated with high-speed debris flows such as very mobile rockslides and NUÉE ARDENTE. It may be that fluidization within a rock avalanche mass can account for the properties of catastrophic landslides. The fluidization arises from the interaction of energetic particles and trapped air under pressure. In a fluidized system granular material is supported by air (or any other gas, but usually air). The geomorphological relevance of fluidization is to ground failure in which debris flows occur, usually as long run systems apparently supported by air. Some classic landslides, e.g. Blackhawk, Elm, Frank, Saidmarrah, etc., fall into this category. One of the most striking properties of debris is its relatively high fluidity; debris flows with 80–90 per cent granular solids by weight can move in sheets about 1 m thick over surfaces with slopes of 5–10 degrees. The high fluidity suggests that the debris is 'fluidized' in the sense that this term is used by the chemical engineers. In fluidization, the interstitial fluid moves so rapidly upwards through the granular solids that these solids are suspended.

There seem to be three possibilities for debris flow mechanisms: either the flow moves essentially as a mass, supported by an air cushion, which allows long travel; or the system is fluidized in the classical sense and air and particles are interacting to keep the system mobile; or the mobility is ensured by particles interacting with each other in the high energy system. There are factors which support all three views. The Blackhawk landslide fell about 1,000 m from the mountain of the same name in south California in some prehistoric period. Possibly the slide moved almost as a single unit, gaining a nearly frictionless ride on a cushion of compressed air. This theory appears to be supported by the marginal ridges of debris formed where material was dropped as air leaked from the edges of the slide mass, and by debris cones on the landslip formed by air leakages blasting up through holes in the main mass.

There were no witnesses to the Blackhawk event; the Elm rockslide in Switzerland was closely observed. In September 1881 a large part of the Plattenberg mountain fell about 400 m and landed near the village of Elm. A vast amount of rubble crashed to the valley floor, bounced 100 m up the opposite wall, turned and – in less than a minute – careered down the valley for over 1 km before coming to a sudden stop. It is feasible that the Elm rockslide could have travelled down the valley on a cushion of air but eyewitnesses suggest that some fluidization mechanism was more likely; the surface of the slide was observed 'boiling' in great turbulence, and parts of the slide ran into houses, suggesting a great overall mobility.

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SEE ALSO: liquefaction

IAN SMALLEY

FLUVIAL ARMOUR

'Armour' is one of several terms applied to clastic deposits in which the surface layer is coarser than the substrate (see BOULDER PAVEMENT). The phenomenon is widespread in gravel-bed rivers, where maximum surface and subsurface grain sizes may be similar but the respective median diameters differ by a factor of 2 to 4 because fine material is largely absent from the surface. It can also occur in sand-bed rivers that contain a little gravel, which becomes concentrated on the surface as a stable lag deposit. The traditional explanation is preferential winnowing of finer sediment from the surface during degradation, for example below dams which cut off the gravel flux from upstream so that the river erodes its bed to regain a capacity BEDLOAD. This degradation is self-limiting because as the surface coarsens, the transport capacity of the flow declines and the bed becomes immobile. This 'static' armouring has been investigated in flume experiments with no sediment feed and has been modelled mathematically; see Sutherland (1987) for a good review.

Coarse surface layers also exist in unregulated rivers with an ongoing sediment supply and peak flows which can transport all sizes of bed material. This 'mobile armour' allows the channel to be in equilibrium (neither degrading nor aggrading, neither coarsening nor fining) despite the size-selective nature of bedload transport, since intrinsically less mobile coarse fractions are preferentially available for transport whereas potentially mobile fine fractions are mainly hidden in the subsurface (Parker and Klingeman 1982). The armour forms by vertical winnowing during active bedload transport: entrainment of coarse clasts during floods creates gaps which are filled mainly by finer grains. Extreme floods may wash out the armour, but it re-forms during intermediate flows in most environments. In ephemeral streams there may be no such flows and armouring is generally absent (Laronne *et al*. 1994).

Mobile armour helps reduce bedload flux to match a restricted supply, with static armour as the limiting case when supply is cut off completely (Dietrich *et al*. 1989; Parker and Sutherland 1990). Changes in grain packing, as well as size distribution, are involved in this self-regulation; they include imbrication of coarser clasts and the development of pebble clusters, stone cells and transverse steps. The river bed is thus a degree of freedom in the adjustment of alluvial channels (see CHANNEL, ALLUVIAL) towards grade (see GRADE, CONCEPT OF).

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ROB FERGUSON

FLUVIAL EROSION QUANTIFICATION

The quantification of fluvial erosion, as for any other geomorphic process, is a way to make more precise and objective the assessment of the morphological changes that affect the Earth's relief. Beside this purpose, fluvial erosion quantification has a critical importance in the field of APPLIED GEOMORPHOLOGY because it can help the elaboration of erosion rate prediction models that are useful in the evaluation of GEOMORPHOLOGICAL HAZARDS.

Fluvial erosion consists of the entrainment and transport (by solution, suspension and traction) of particles that make up the stream bed or the stream banks (see EROSION). To quantify fluvial erosion, therefore, means to assess the amount of materials that a stream is capable of wearing away from bedrock or alluvial channels. These materials, however, become part of the total solid load and it is almost impossible to differentiate them from those delivered to the stream and deriving from the denudation processes acting within the whole basin. In other words, although streams are powerful erosional agents, they operate in conjunction with other exogenetic agents and with unconcentrated surface waters in particular; therefore drainage basins are acknowledged as the fundamental geomorphic units and rivers as their main elements along which energy is available. Consequently the quantification of fluvial erosion must be understood as quantification of the overall DENUDATION affecting both the slopes and the channels of the drainage basins.

The estimates of denudation affecting the slopes are often based on the direct determination of the amount of sediments removed from small sample areas or erosion field plots, or on the measurement of ground surface lowering and calculation of the volume of sediment dislodged. The evaluations of SOIL EROSION obtained in this way, however, are strictly dependent on the peculiarities of the studied slopes; therefore they have only local significance and can lead to misleading conclusions when they are extended to larger areas. In spite of this limit, a large number of field data on soil loss are useful in developing erosion prediction equations when they are plotted against several erosion controlling factors: the UNIVERSAL SOIL LOSS EQUATION is the most famous of them.

Channel erosion is usually evaluated by surveying the modifications of channel form and calculating the volume of material removed. The methods used include measurements to reference pegs and periodical controls of both channel cross section and long profile. Long-term variations can also be estimated by comparing aerial photographs of different periods.

As the material removed from slopes and channels of a drainage basin is the source for fluvial transport, the total amount of sediment load (see SEDIMENT LOAD AND YIELD) at the main river mouth can measure the intensity of denudation affecting the whole basin. Actually, most of the attempts directed towards the determination of erosion rate are based on the assessment of stream load quantity. Such assessment can be approached in different ways. One approach consists in the measurement of all the transported materials at the recording stations, that is to say the direct measurement of dissolved load, SUSPENDED LOAD and BEDLOAD. The indirect approaches, instead, lead to the stream load prediction through theoretical formulae or multiple regressions.

The field determination of dissolved load is obtained by portable instruments that measure certain water quality parameters, such as conductivity and pH. More often the dissolved solid content is determined in laboratories by evaporation of known volumes of water and weighing the residue. Suspended load concentration is obtained by measuring the turbidity of water samples collected by specially developed devices, ranging in complexity from simple dip-bottles to sophisticated apparatus; the total suspended load is then obtained multiplying the suspended load concentration by the discharge. The assessment of bedload is extremely difficult; many measuring apparatuses have been developed, like slot traps, collecting basin, basket samplers, etc., but none has been universally accepted as adequate for the determination of bedload.

Direct measurements of total solid load by rivers have encountered many problems; among them there are the high costs of instruments, the running expenses, and the alteration of pattern of flow and transport by the presence of the sampler, which can distort especially bedload data. One more problem is sampling both in time and space. Observations made at given time intervals could miss the extreme events; furthermore it may require many years of record before data are enough to be significant. The choice of sampling site must consider the accessibility of instruments, the lack of interference and the planning of a dense instrumentation network.

The theoretical estimation of solid load implies the derivation of specific formulae based on the characteristic of channel flow and of the transported materials. This procedure is unsuitable to predict suspended load, as it is essentially a non-capacity load, but it has been tentatively followed to predict bedload; however none of the derived formulae would seem to offer a completely satisfactory prediction.

An indirect method largely used to evaluate fluvial erosion takes into account the data available on suspended load (the most systematically measured at recording stations) and leads to significant regressions that relate suspended load to several parameters which express the principal factors influencing the spatial pattern of sediment production (Table 18). Once obtained these equations are used to predict suspended load of rivers lacking a recording station. Although suspended load values are a partial measure of erosion processes in drainage basins, they have been used also to obtain world maps of denudation rate (Fournier 1960).

Table 18 Some examples of multivariate regressions between suspended load and controlling variables

Author	Region	Equation
Fournier (1960)	Temperate alpine areas	$\log \mathbf{E} = 2.65 \log (\mathbf{p}^2\ \mathbf{P}^{-1}) + 0.46\ \mathbf{H}_m \tan \phi - 1.56$ $\mathbf{E}$ = suspended sediment yield (tonnes km^{-2} year^{-1}); $\mathbf{P}$ = precipitation in month of maximum precipitation (mm); $\mathbf{P}$ = mean annual precipitation (mm); $\mathbf{H}_m$ = mean elevation of basin (m); ϕ = mean basin slope (°)
Jansen and Painter (1974)	Humid microthermal climatic areas	$\log \mathbf{S} = -5.073 + 0.514 \log \mathbf{H} + 2.195 \log \mathbf{P} - 3.706 \log \mathbf{V} + 1.449 \log \mathbf{G}$ $\mathbf{S}$ = suspended sediment yield (tonnes km^{-2} year^{-1}); $\mathbf{H}$ = altitude (m.a.s.l.); $\mathbf{P}$ = mean annual precipitation (mm); $\mathbf{V}$ = measure of vegetation cover; $\mathbf{G}$= estimate of proneness to erosion
Jansen and Painter (1974)	Temperate climates	$\log \mathbf{S} = 12.133 - 0.340 \log \mathbf{Q} + 1.590 \log \mathbf{H} + 3.704 \log \mathbf{P} + 0.936 \log \mathbf{T} - 3.495 \log \mathbf{C}$ $\mathbf{S}$ = suspended sediment yield (tonnes km^{-2} year^{-1}); Q = annual discharge (10^3 m^3 km^{-2}); $\mathbf{H}$ = altitude (m a.s.l.); $\mathbf{P}$ = mean annual precipitation (mm); $\mathbf{T}$ = average annual temperature (°C); $\mathbf{C}$ = natural vegetation index
Ciccacci *et al.* (1986)	Italy	$\log \mathbf{TU} = 2.79687 \log \mathbf{D} + 0.13985\ \Delta a + 1.05954$ $r^2 = 0.96128$ $\mathbf{Tu}$ = suspended sediment yield (tonnes km^{-2} year^{-1}); $\mathbf{D}$ = drainage density (km km^{-2}); $\mathbf{\Delta a}$ = hierarchical anomaly index

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ELVIDIO LUPIA-PALMIERI

FLUVIAL GEOMORPHOLOGY

Fluvial geomorphology is strictly the geomorphology of rivers. As rivers have always held a prominent role in the study of landforms, it is not surprising that debates about fluvialism, as to whether rivers could produce their valleys, continued to rage early in the nineteenth century until uniformitarianism prevailed, whence temperate areas were seen as the result of rain and rivers. At the end of the nineteenth century rivers were seen as central to the Davisian normal cycle of erosion which came to exert a dominant influence upon geomorphology for the first half of the twentieth century (Gregory 2000). It took time to appreciate that there was insufficient understanding of fluvial processes and, with hindsight, it has been suggested that the way in which G.K.Gilbert approached rivers, including significant contributions on the transport of debris by flowing water (Gilbert 1914), could have provided at least an additional, if not an alternative, approach to that developed by Davis.

Until the mid-twentieth century fluvial geomorphology was dominated by attempts to interpret landscapes in terms of phases of river evolution with emphasis placed upon terraces as indicating sequences of valley development, and erosion or planation surfaces employed to reconstruct stages of landscape development. Questioning the basis for such reconstructions, and realizing a need for a greater focus upon fluvial processes, was answered by *Fluvial Processes in Geomorphology* (Leopold *et al.* 1964) which came to have a dramatic influence upon the way in which fluvial geomorphology was subsequently pursued. In addition to increasing interest in hydrological processes (see HYDROLOGICAL GEOMORPHOLOGY) and leading to recognition that the drainage basin could be regarded as the fundamental geomorphic unit (Chorley 1969), it provided the basis for an expansion of research on the contemporary fluvial system. Emphasis was effectively placed upon seven different themes (Gregory 1976) which were: drainage network morphometry; drainage basin characteristics particularly in relation to statistical models of water and sediment yield; links between morphology and process in the hydraulic geometry of river channels; and the controls upon river channel patterns; together with theoretical approaches to the fluvial system; investigations of the significance of dynamic contributing areas in runoff generation; and finally, ways of analysing changes in fluvial systems including the PALAEOHYDROLOGY-river metamorphosis approach. At this stage in the rapid development of fluvial geomorphology, textbooks were produced including emphases on dynamics and morphology (Morisawa 1968), form and process (Richards 1982; Morisawa 1985; Knighton (1984), 1998), rivers and landscape (Petts and Foster 1985), and the fluvial system (Schumm 1977a). There was thus a progressive development of fluvial geomorphology; in his conclusion to *The Fluvial System*, Schumm (1977a) contended that landscape, like science itself, proceeds by episodic development. In the second part of the twentieth century the development of fluvial geomorphology proceeded episodically, and the result of research progress made has been to demonstrate, in turn, exactly how episodic fluvial systems can be.

Comparatively few explicit definitions of fluvial geomorphology have been given in papers and books but five (Table 19) indicate core themes and to some extent intimate how the subject has evolved, expanded and progressed since the 1960s.

In the course of the development of fluvial geomorphology since the middle of the twentieth century, three sequential phases can be visualized: one of import and expansion, one of consolidation, and one of innovation. The first saw import and utilization of understanding and techniques from other disciplines including hydraulics, hydrology, sedimentary geology and engineering. Clarification and refinement of field

Table 19 Some definitions of fluvial geomorphology

Definition	Source
Fluvial geomorphology has as its object of study not only individual channels but also the entire drainage system	Kruska and Lamarra (1973) cited by Schumm (1977b)
A primary objective of fluvial geomorphology must be to contribute to *explanation* of relationships among the physical properties of flow in mobile-bed channels, the mechanics of sediment transport driven by the flow, and the alluvial channel forms created by spatially differentiated sediment transport	Richards (1987)
Geomorphology is the study of Earth surface forms and processes; fluvial phenomena – those related to running water	Graf (1988)
The study of changing river channels is the domain of fluvial geomorphology. . . . Fluvial geomorphology is a field science; classification and description are at the heart of this science	Petts (1995)
The science that seeks to investigate the complexity of behaviour of river channels at a range of scales from cross sections to catchments; it also seeks to investigate the range of processes and responses over a very long timescale but usually within the most recent climatic cycle	Newson and Sear (1998), cited by Dollar (2000)

approaches together with modelling methods including stochastic, deterministic (Werritty 1997) and experimental (Schumm *et al*. 1985) approaches enabled the developing foundation to focus upon equilibrium concepts throughout several distinct sub-branches of fluvial geomorphology established as the outcome of this phase. The independence or dependence of variables involved in research investigations was anticipated to vary according to the steady, graded, or cyclic timescale (Schumm and Lichty 1965) of the investigation being considered. Once established with a significant number of practitioners, fluvial geomorphology experienced a second phase, one of internal consolidation, characterized by investigations of changes over time as referred to different timescales, embracing palaeohydrology, and river channel adjustments as instigated by the effects of land use changes impacting upon the channel. These investigations meant that controls upon change of the fluvial system were explored, including thresholds, complex response and sensitivity. The third phase of innovation, aided by new technology of remote sensing and GPS, has seen the development of exciting links between investigations undertaken at several spatial scales, together with equally innovative developments linking studies of process with landscape development. This phase has been one of export whereby results from fluvial geomorphology are contributing to multidisciplinary projects and making a distinctive input to management problems (e.g. Thorne *et al*. 1997).

Against this background of development of the subject, a choice for fluvial geomorphology was suggested to exist by Smith (1993) because he perceived the discipline to be at a crossroads, requiring major changes in ways of thinking and operating, so that he proposed it needed to move forward and to adopt the ways of the more competitive sectors of the Earth and biosciences. However this need may have been overstated (Rhoads 1994) in view of the vitality shown by recent publications in fluvial geomorphology, and by the way in which collaboration and multidisciplinary activity have increased, complemented by attention being devoted to the scientific foundation of the subject. Rhoads (1994: 588) sees the most critical challenge facing fluvial geomorphologists as that of devising effective strategies for integrating a diverse assortment of research, spanning a broad range of spatial and temporal scales.

From the ideas that prompted Smith's challenge and Rhoads's response, it is possible to seek a general definition; the broad range of research approaches that exists; and ways in which collaboration is now possible. Embracing earlier proposals (Table 19) a general definition for fluvial geomorphology could be that it investigates the fluvial system at a range of spatial scales from the basin to specific within channel locations; at timescales ranging from processes during a single flow event to long-term Quaternary change; undertaking studies which involve explanation of the relations among physical flow properties, sediment transport and channel forms; of the changes that occur both within and between rivers; and that it can provide results which contribute in the sustainable solution of river channel management problems.

Although developed at different times and progressed significantly since the seven themes suggested in 1976 (Gregory 1976), there is now a range of research approaches which are the branches of fluvial geomorphology occupying most practitioners at any one time. These can now be envisaged as focused on components of the fluvial systems, process mechanics, temporal change and management applications.

Components of the fluvial system

Studies of components of the fluvial system have been concerned particularly with morphology of elements of that system across the range of spatial scales from in-channel locations to the complete drainage basin. Particularly significant investigations focused on relations between form and process in fluvial systems and upon the controls upon morphology. This has required definitions which can be applied in different basins including those for channel capacity, channel planform, floodplain extent, and drainage density of the channel network; and at each of these levels there have been attempts to establish equilibrium relations between indices of process and measurements of fluvial system form. Some of the earliest developments in fluvial geomorphology were concerned with analysis of *drainage networks* using techniques of drainage basin morphometry and with the relationship between *channel capacity* and the frequency of the bankfull discharge which was thought to exercise a major control upon channel morphology. In these and other components of the fluvial system it has now been appreciated that the links between form and process and the associated explanation is more complex than at first thought. Thus drainage networks could not easily be related to discharge and channel processes, and the relationship between channel capacities and controlling discharge has been the subject of considerable research, particularly the way in which networks generate the Geomorphic unit hydrograph (Rodriguez-Iturbe and Rinaldo 1998). Relations between dimensions of cross-sectional area and width can be used to provide a basis for discharge estimation at ungauged sites (Wharton *et al*. 1989). In addition research has focused upon *river channel patterns*, upon the controls on single thread and multi-thread patterns and what determines the thresholds between them. The floodplain is also controlled by the interaction between recent hydrological and sediment history together with the characteristics of the local area; and the variability of floodplain characteristics has been reflected in the definition of the river corridor as well as of the floodplain itself. Three major floodplain classes, based on stream power and sediment characteristics, have been recognized (Nanson and Croke 1992), further subdivided into a combination of thirteen floodplain orders and suborders, namely:

1 High energy non-cohesive floodplains: disequilibrium landforms which erode either completely or partially as a result of infrequent extreme events.
2 Medium energy non-cohesive floodplains: in dynamic equilibrium with the annual to decadal flow regime of the channel and not usually affected by extreme events. Preferred mechanism of floodplain construction is by lateral point bar accretion or braid channel accretion.
3 Low energy cohesive floodplains: usually associated with laterally stable single-thread or anastomosing channels. Formed primarily by vertical accretion of fine-grained deposits and by infrequent channel avulsion.

As more is known about each of the several spatial scales of investigation of the fluvial system it is appreciated that the question of explanation relies upon the controls that apply to each particular spatial level. Thus it is necessary to see how the flows and sediment transport are significant in relation to each of the spatial levels of the fluvial system, and how they interrelate. Furthermore, when focusing on the integrity of the fluvial system, it is necessary to consider how a hierarchy of interrelated components makes up the river basin channel structure. Any such structure needs to take account of the progress made by biologists and aquatic ecologists in this regard and an original framework (Downs and Gregory 2003: Chapter 3) involves seven nested scales which are drainage *basin*, basin *zones*, valley *segments*, stream *reaches, channel unit, within channel*, and channel *environment at a point*. In addition there are a number of environmental flows that have been defined (Dollar 2000) including those that maintain a channel morphology and which have been specified for the purposes of practical application.

Process mechanics

Process mechanics began with hydrological analyses whereby fluvial processes were examined from the standpoint of analysis of stream hydrographs and their generation, so that investigations of dynamic contributing areas became a major reason for field experiments based in small experimental catchments. On the basis of the considerable progress made, attention then moved to the sediment budget, and to suspended sediment, bedload and solute loads. Analysis previously dominated by simple rating curves, which assumed a linear relationship between suspended sediment concentration and discharge, was refined once it was shown that because of hysteretic effects and sediment supply problems, the relationships were much more complex so that earlier estimates of rates of denudation had to be revised. Bedload transport had been very difficult to measure so that transport equations often tended to assume a capacity load, whereas along many rivers the transport of material was supply limited. Advances in instrumentation enabled continuous recording devices to be used providing the basis for more complete explanatory analysis of transport rates; and continuous measurements of channel bank erosion could be the basis for more precise relationships between erosion rates and the controlling variables. Such studies facilitated more detailed investigations within the channel and these have been concerned with the entrainment of bed material, the patterns of flow and sediment movement at confluences, and the controls upon in-channel change. A particularly fruitful theme has derived from the interrelationships with ecology because aquatic ecologists have investigated river channels in relation to instream habitat conditions and aquatic plant distributions; combination of such results with geomorphological data has promoted biogeomorphological investigations of river channels. Along rivers bordered by riparian trees, or flowing through forested areas, the investigation of CWD (coarse woody debris) has become important because such CWD exerts an influential control upon the channel processes, the morphology and ecological characteristics. Numerous investigations have been undertaken considering the impact of CWD upon channel morphology, demonstrating the extent and significance of wood often as debris dams, together with the impact on channel processes, the reasons for spatial variations, as well as the stability of dams and their persistence together with the management implications.

Temporal change

Study of temporal changes had been a long tradition in fluvial geomorphology through the interpretation of past stages of development based on river terrace sequences, but it was not until 1954 that the idea of palaeohydrology was proposed (Leopold and Miller 1954). This contrasted with earlier approaches because it was retrodictive in approach, utilizing understanding that had been gained of contemporary processes, and it was exemplified by Quaternary palaeohydrology suggested by Schumm (1965) and augmented by ideas of river metamorphosis (Schumm 1977a). Palaeohydrology evolved (Gregory 1983) to utilize knowledge of contemporary processes applied to the past, whereas river metamorphosis similarly employed contemporary relationships between channel form and process as a basis for interpreting river channel resulting from a range of causes including dam and reservoir construction, land use change including urbanization and, particularly, as a consequence of channelization. Such human-induced channel changes were found to be extensive (e.g. Brookes and Gregory 1988) and were superimposed upon the impact of shifts in sequences of climate which in some parts of the world, such as Australia, led to the alternation of periods of drought-dominated and flood-dominated regimes. Analysis of river channel changes was initially confined to particular reaches affected, often downstream from the influencing factor, but they have subsequently been analysed in the context of the entire basin with attention accorded to the spatial distribution of adjusting channels and emphasis given to the extent to which they are potentially able to recover to their former condition (e.g. Fryirs and Brierley 2000). The considerable progress achieved as a result of studies of channel change includes ways in which thresholds can be identified, reaction and relaxation times (Graf 1977), patterns of palaeohydrological change in different parts of the world (Benito and Gregory 2003) all culminating in further, more informed understanding of the ways in which fluvial processes relate to environmental change (Brown and Quine 1999) and how fluvial systems and sedimentary sequences reflect shifts of climate, both short or longer term during the Quaternary (Maddy *et al.* 2001). A particularly effective way in which studies of the past have been successful has been the analysis of palaeofloods (see HYDROLOGICAL GEOMORPHOLOGY), overcoming not only inaccuracies in estimating the ages of floods and in reconstructing flood discharges, but also allowing incorporation of palaeoflood data into flood frequency analysis, in order to analyse the effects of climatic shifts and non-stationarity. The results of palaeoflood hydrology have been of practical application by bringing very specific benefits in the design or retrofitting of dams or other floodplain structures.

Management applications

Palaeoflood analysis is just one example of ways in which fluvial geomorphology provides applications to management. Many other applications have been developed and initially were very problem and reach specific, including estimation of sediment yield and the possibility of gullying and channel change (Schumm 1977a), progressing through consequences of particular impacts such as channelization (Brookes 1988), leading to improved procedures for management (Brookes and Shields 1996) and then to comprehensive statements of ways in which fluvial geomorphology can be applied to river engineering and management (Thorne *et al.* 1997). Particular emphasis has been placed upon river restoration and how fluvial geomorphology is able to contribute significantly to restoration projects (Brookes 1995). One aspect of restoration to be considered is what is natural (Graf 1996) and therefore what should be the objective for a particular restoration project. It is being appreciated that in all cases where human impact affects fluvial systems, current knowledge of the way in which the system has evolved can illuminate the way in which management is undertaken. It is also the case that some aspects of human activity are now being substantially reversed and the implications of dam removal (Heinz III Center 2002) is one topic of current interest.

It may seem from the foregoing outline of approaches in fluvial geomorpology that they have become increasingly reductionist and diverse, but there are a number of ways in which there has been integration and collaborative activity tending to unify fluvial geomorphology – although remembering that fluvial geomorphology is not as independent of other disciplines as it once was. Thus analysis of sediment slugs showing how waves of sediment are transmitted through a fluvial system has implications for management, and for interpretation of temporal change as well as for the understanding of contemporary channels' forms and processes. In addition, the use of fallout radionuclides including Cs-137 and Pb-210 has enabled precise dating of specific fluvial changes including floodplain sedimentation (e.g. Walling and He 1999). Based upon knowledge of a number of specific cases, the limits of explanation and prediction have been emphasized, including ten ways to be wrong (Schumm 1991). This introduces the idea of risk so that the fluvial system can be seen in terms of the incidence of twenty-eight geomorphic hazards which may occur, associated with drainage networks, hillslopes, main channels, piedmont and plain areas (Schumm 1988). From each of the above themes, clear signs are emerging of more integrated investigations, for example linking process and morphology in bank stability and modes of channel adjustment which involve bank erosion. There is great awareness of, and interaction between, the range of spatial scales investigated. Such linking analysis is now being facilitated by enhanced remote sensing techniques, and GPS which enhances the detail of data capture and the speed of analysis. Progress is now being made towards enhanced conceptual models which seek to model aspects of the fluvial system in ways not previously possible (Coulthard *et al.* 1999). The outstanding challenge is for further understanding to be achieved of the way in which information can be linked from one timescale to another.

It is inevitable that the investigation of rivers should have expanded greatly, as one of the most researched fields of geomorphology; it is now becoming more integrated but not necessarily strictly confined to geomorphology, as links with ecology, engineering and hydraulics and sedimentary geology prove to be very worthwhile, and multidisciplinary approaches are increasingly common. The concerns expressed by Smith (1993) are being met, and fluvial geomorphology is now sufficiently well founded to address major questions including what has been suggested to be the greatest challenge: 'to understand the way in which short timescale and small space scale processes operate to result in long timescale and large space scale behaviour' (Lane and Richards 1997: 258).

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KENNETH J. GREGORY

FOLD

Structures that originally were planar, like a sedimentary bed, but which have been bent by horizontal or vertical forces in the Earth's crust. Folds can occur at scales that range from mountain ranges to small crumples only a few centimetres long. They can be gentle or severe, depending on the nature and magnitude of the applied forces and the ability of the beds to be deformed. When beds are upfolded or arched (Figure 64a) they are called *anticlines*, whereas downfolds or troughs are termed *synclines*. A *monocline* is a step-like bend in otherwise horizontal or gently dipping beds. Folds can be asymmetrical in shape and, when the deformation is particularly intense, they can be overturned. They also tend to occur in groups rather than in isolation and some mountains consist of a folded belt. Folds can result from various processes: compression in the crust, uplift of a block beneath a cover of sedimentary rock so that the cover becomes draped over the rising block, and from gravitational sliding and folding where layered rocks slide down the flanks of a rising block and crumple. The process of folding may involve either small-scale shearing along many small fractures or flowage by plastic deformation of the rock.

Large folds can have a substantial influence upon landform development. Folding can also occur rapidly, creating vertical increases in elevation of up to 10 m 1,000 y^{-1}. However, when folds cease to grow, the influence of erosion becomes increasingly more important than the original shape of the fold. Thus, in the case of an anticline, initially it will form an area of upstanding, often donal relief. However, as it is eroded, the uppermost strata may be cut through. If the older rock that forms the core is of low resistance the result is a *breached anticline* in which a series of inward-facing escarpments rise above a central lowland. A classic example of this is the Weald of southern England (Jones 1999) (Figure 64b). Conversely, as in the Paris and London Basins, the rim of a syncline will possess outward facing scarps. Folding has major impacts on river systems. If a fold develops across a stream course the river, if it can cut down quickly enough, can maintain its course transverse to the developing structure. Such a channel is said to be *antecedent*.

Antecedence is, however, only one explanation for drainage that cuts across anticlinal structures. An alternative model (Alvarez 1999) is that fold ridges emerge from the sea in sequence, with the erosional debris from each ridge piling up against the next incipient ridge to emerge, gradually extending the coastal plain seaward. The new coastal plain, adjacent to each incipient anticline, provides a level surface on which a newly elongated river could cross the fold, positioning it to cut a gorge as the fold grew. This mechanism is in effect a combination of antecedence and superimposition. The model has been applied both to the Appalachians of the USA and the Apennines of Italy.

On actively developing anticlinal folds, drainage density varies according to the gradient of the evolving slopes. However, the form of the relationship between gradient and drainage density is process-dependent. Talling and Sowter (1999) suggest that a positive correlation occurs when erosion results from overland flow, while a negative correlation occurs when erosion is dominated by shallow mass-wasting. A traditional description of drainage in folded areas such as the Jura is provided by Tricart (1974).

Where sedimentary rocks are tilted by folding there may be a succession of lithologies exposed that have differing degrees of resistance. River channel incision will tend to be more effective on

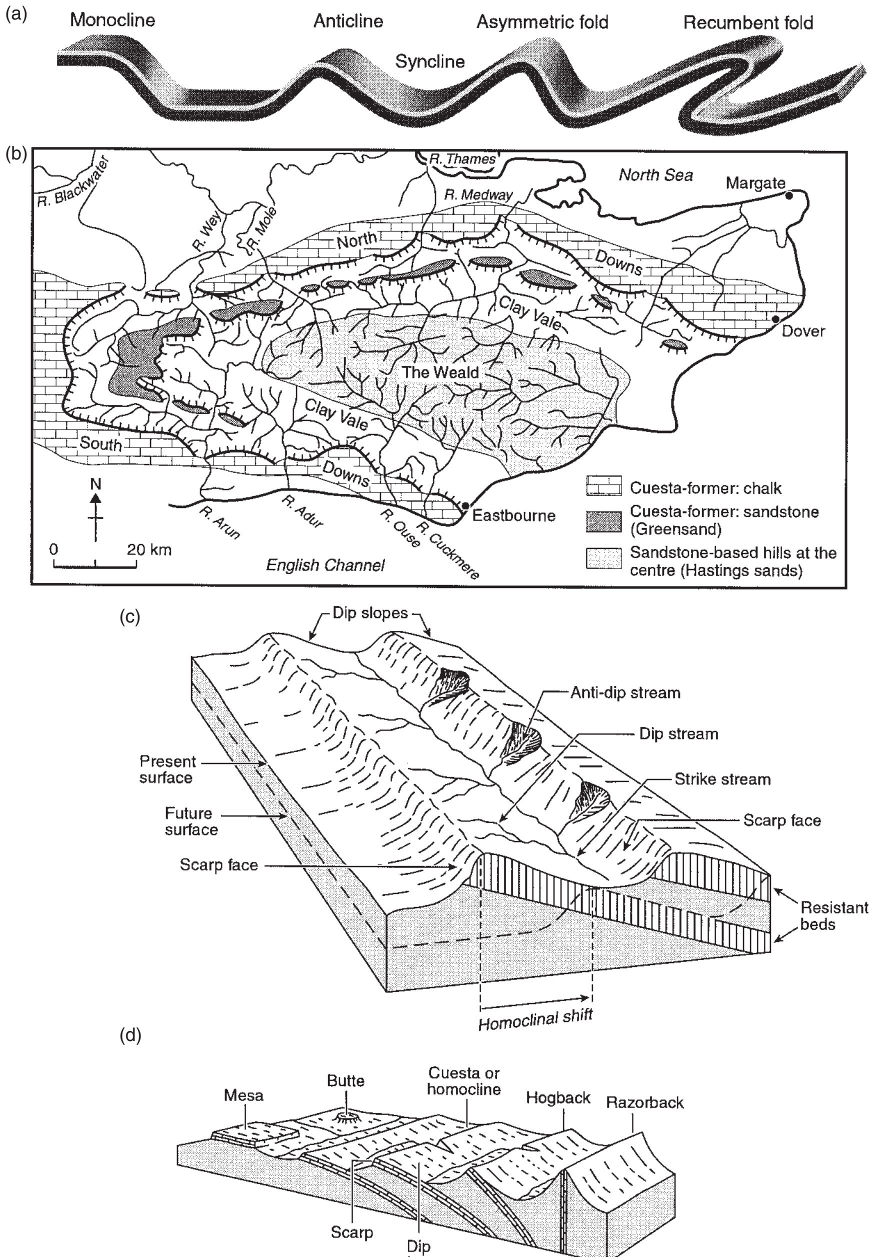

Figure 64 Folds and their relationship to relief: (a) some major types of fold; (b) the Wealden Anticline of south-east England; (c) drainage and slope forms associated with dipping strata; (d) drainage and slope forms associated with strata of progressively steepening dip

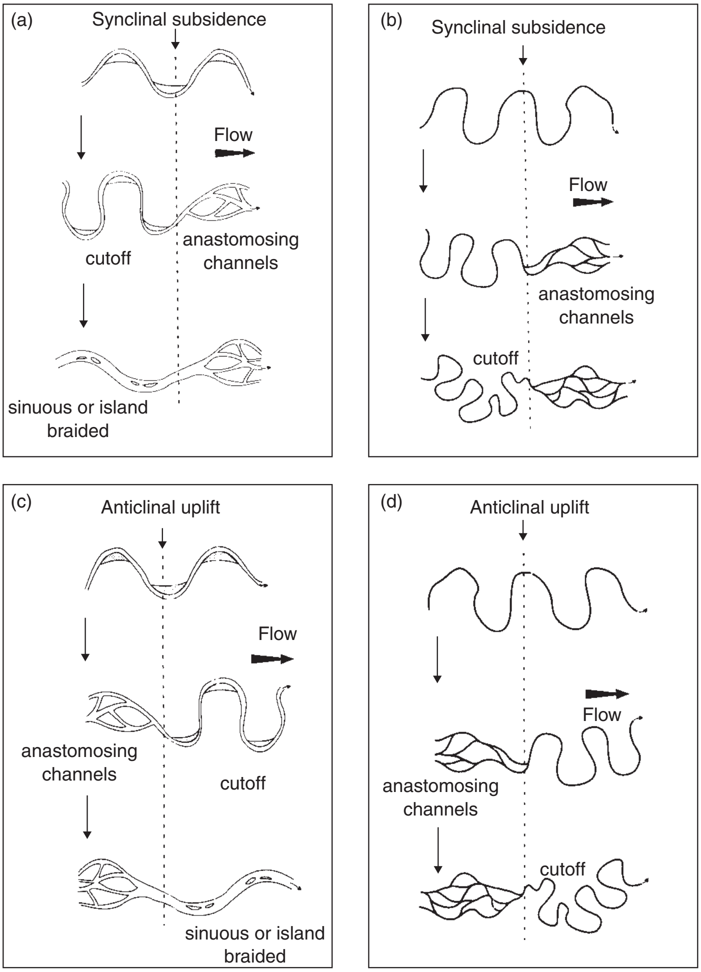

Figure 65 River channel patterns associated with synclinal and anticlinal activity: (a) and (c) = mixed-load meandering channels; (b) and (d) = suspended-load meandering channels (modified from Ouchi in Schumn *et al.* 2000, figures 3.10, 3.11, 3.12 and 3.13)

less resistant beds leading to the development of a *strike valley* (Figure 64c), flanked on the up-dip side by a dip slope and on the down-dip side by an escarpment. The roughly parallel strike streams will be joined at high angles by short *dip streams* and *anti-dip streams*. As downward incision occurs, the rivers will migrate laterally by a process known as *homoclinal shifting*.

The angle of dip also influences topographic form (Figure 64d). Resistant beds in very gently dipping or horizontal beds, form flat-topped plateaux (MESAS). Modestly dipping beds create a CUESTA, whereas steeply dipping strata produce a ridge known as a HOGBACK.

In recent years tectonic geomorphologists (see, for example, Burbank and Anderson 2001) have taken a great interest in how the style and rate of folding affects landform evolution. In particular, Schumm *et al*. (2000) have described how terrace formation, channel form, the locations of degradation and aggradation, valley long profiles, and the spatial distribution of flooding, may be related to folding activity. Figure 65 shows the response of stream channel form to anticlinal uplift and synclinal subsidence for mixed-load and suspended-load streams. Streams flowing across zones of uplift (live anticlines) may show deformed terraces and convex sections of long profile.

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A.S. GOUDIE

FORCE AND RESISTANCE CONCEPT

Materials at the Earth's surface are subjected to a whole range of applied forces (inputs of energy) both physical and chemical. Whereas in Newtonian mechanics applied force (or stress) and reaction (or strain) are thought of as equal and opposite and as essentially simultaneous, the outcome of an application of unit force to Earth materials often cannot be readily predicted. The asymmetry between energy input and response which is almost universally experienced in geomorphology is due partly to the multifaceted nature of resistance and partly to the significance of the specific sequence of energy inputs.

B.W. Sparks was especially concerned to stress that the resistance of rocks to geomorphic processes was always contingent upon the precise manner in which energy was applied. An obvious example is the very different resistance of limestones under chemical or mechanical attack (cf. Sparks 1971). This difficulty is well exemplified by M.J. Selby's elaborate quantification of the ROCK MASS STRENGTH of hillslopes, which nevertheless can only be invoked when the applied force environment is closely defined (Selby 1993: 104). Similarly, the difficulty in writing rational physical equations to describe fluvial flow and sediment transport may be directly traced to the non-uniform expression of channel boundary resistances.

Coupled with this first asymmetry is the widespread observation that energy inputs of equal magnitude do not result in equal amounts of geomorphic work. The non-linearity (see NON-LINEAR DYNAMICS) of process and response is of profound significance in all historical sciences. The most common manifestation of this asymmetry in geomorphology is in hysteresis loops of discharge and sediment plots. The applied force of a given discharge will generally neither entrain nor carry the same volume or calibre of sediment on the rising and falling limbs of a flood. The asymmetry is produced by the variable quantity and quality of sediment available to be moved: that is, on the temporally specific state of the channel boundary. The study of slope failures also (cf. Schumm and Chorley 1964) provides examples of small inputs of energy which propel a system across a resistance threshold (see THRESHOLD, GEOMORPHIC).

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SEE ALSO: Goldich weathering series; magnitude–frequency concept; non-linear dynamics; rock mass strength; threshold, geomorphic

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FOREST GEOMORPHOLOGY

Forest geomorphology is a specialized area of research focused both on the interaction between forest ecosystem dynamics and landform processes, and the effects of forest management activities on the rates and thresholds of geomorphic process events. As a specialty within the field of geomorphology, the study of forest associated process dynamics has broad application to the study of ecosystem dynamics, endangered species and conservation studies, and paleo-landscapes.

Several features distinguish geomorphic processes in forest environments from those in other vegetation types. Forest ecosystems play a prominent role in many aspects of sediment production, transport and storage. Entrainment of soil by windthrow of trees, the binding effect on soils and regolith by tree root strength, and soil displacement by burrowing animals, are among many examples of how sediment movement is influenced by forests. Standing trees and fallen logs on slopes in forested terrain act as temporary sediment storage 'traps', while forest management activities, such as road construction and the removal of trees, transport and disturb sediment production in other ways. Forest composition plays a major role in the interception, evapotranspiration, infiltration, and runoff of precipitated water, and, as in the sediment examples cited, profoundly influences the type, frequency and mechanisms controlling mass wasting and slope stability. Ultimately, such influences of forest vegetation profoundly affect the development and operation of drainage basins, adding a significant organic component to sediment load, hydrologically influencing the discharge response of the channel network, even influencing channel development through bank resistance or flow deflection. Added to these considerations are natural cycles within the forest ecosystem such as wildfire, floods, disease or insect impacts, and human disturbance factors, which can affect the linkage of geomorphic processes, as well as their magnitude and frequency.

The diversity of objectives, approaches and professional disciplines of those who conduct forest-related process studies has resulted in poor communication among scientists and land managers who share many common interests. In recognition of the need to promote interdisciplinary dialogue, and approach complex environmental research within a spatial and temporal scale appropriate to the changes in forest succession affecting process-response, the International Council of Scientific Unions created the International Geosphere–Biosphere Program (IGBP) which recognizes several prominent forest biomes. Geomorphologists are placed prominently among the scientific teams that continue to illuminate the linkages between landforms and ecosystems.

History and development of forest geomorphology

While late nineteenth and early twentieth-century landscape theory focused on denudation and landform development, early geomorphologists were aware of variation in landscape appearance under varying climate and vegetation conditions. Cotton (1942) explains variations in landscape form as climatic 'accidents', including the resultant vegetation influences, as complications of the 'Geographical Cycle' theorized by William Morris Davis (1899). In a similar vein, Birot (1968) further expands upon Davis, and adds vegetation and soil factors to illustrate the influences of climatic variation. Peltier (1950) also refers to Davis's cycle, but places emphasis on the variations in geomorphic process activity under a variety of forest biomes (selva, savanna, boreal). Peltier credits Professor Kirk Bryan with many of these concepts of 'vegetation modified process'.

Hack and Goodlett (1960) illustrated the relationships between process and forest structure to identify relict features within a forested landscape, and heralded the concept of coexisting ecological and geomorphic equilibriums contributing to landform genesis in humid temperate forests, while Douglas (1968) described the effects in humid tropical forests, and added human disturbance factors. Chorley used examples such as these to illustrate the benefits of multidisciplinary approaches in geomorphology using a 'systems' approach to integrate information from widely divergent sources at different temporal and spatial scales. Chorley's 1962 work encouraged scientists to make contributions across disciplinary lines, and US Geological Survey ecologist Sigafoos (Sigafoos and Hendrick, 1961; Sigafoos 1964; Hupp and Sigafoos 1982) dated trees to determine the temporal and spatial activity of glaciers, floods and blockfields (Alestalo 1971). Successive works were absorbed by the resource management community, which funded research concentrated on the effects of timber harvest practices, road and bridge construction and land use change.

Hydrology, sediment budgets and channel stability in forested watersheds

The application of concepts and techniques from geomorphology to ecosystem studies and terrain analysis represents a great opportunity for the discipline, given the need for an interdisciplinary approach to complex environmental problems.

Large-scale forest management, such as practised by government agencies in the United States, has resulted in numerous controversies between the economic, recreation and conservation interests. In the northwestern Unites States, the practice of large-scale total tree removal, called 'clearcutting', is controversial because of ecological, hydrological and erosion concerns. The US Department of Agriculture, accustomed to years of soil erosion studies at its experiment stations, created a network of experimental forests in the 1960s. Paired watershed studies were conducted to evaluate resulting sediment and water yields resulting from various management treatments. Fredriksen (1970) described the effects of traditional 'clearcutting' using roads, tree removal using a unique cable system to completely suspend the trees as they were cut, and a control basin that remained fully forested. This brief research report sparked considerable interest both in the USA and elsewhere, and a number of research investigations have followed the recovery progress of these small watersheds in the H.J. Andrews experimental forest in western Oregon over the past fifty years (Swanson and Jones 2001).

The natural and management effects of large woody debris on channel morphology and sediment transport in forested streams has been of considerable research interest. Natural log debris in stream channels often produces 'log steps' or 'organic knickpoints' that produce a 'step-pool' profile in mountainous forest streams. These pools act as sediment and nutrient traps, provide fish and invertebrate habitats, and may persist for a long time (Swanson and Lienkaemper 1978). The study of channel stability in forested streams has taken on special significance due to the effects that disturbance within such channels has upon the spawning cycle of anadromous fish. Several species of salmon, steelhead and char have been listed as 'endangered' because of loss of spawning gravel habitat, or loss of access to headwaters. The origin of gravel spawning beds, and their preservation, has occupied substantial geomorphic research including the delivery of gravel from regolith by mass wasting, winnowing of fines by floods, the effects of woody debris on in-channel storage, and the effects of surface 'armour layer' development on entrainment and transport. Stream ecologists, working with fluvial geomorphologists, can provide insights into physical habitat processes and sensitivity to disturbance. Several outstanding classification and inventory schemes have been produced regionally to predict habitat sensitivity and stability (Brussock *et al.* 1985).

Forested slopes have been shown to exhibit lower runoff, increased interflow, and greater stability arising from the root strength of the trees. In general, soils are subsequently deeper than on unforested slopes under similar conditions, resulting in conditions that 'trigger' mass wasting events of both natural and man-induced origins (O'Loughlin 1974). Moss and Rosenfeld (1978) have shown that mass wasting events have the potential to alter the composition and character of the forest community structure in predictable ways, thus leading to a model of interrelationships between landform features and vegetation community characteristics.

On a larger scale, Caine and his co-workers (Swanson *et al.* 1988) demonstrated that ecosystem behaviour can be predicted by a better understanding of how landforms affect those processes. They illustrate that ecosystem–terrain interactions often take multiple forms, with patterns imposed by one set of interactions often coexisting in time and space with other sets. The linkage between ecosystem development and landform stability incorporates both geomorphic and biological events of varying magnitude and frequencies, such as wildfire, floods and landslides. Rosenfeld (1998) illustrates that threshold events, such as exceptional storms, can have predictable 'triggering' effects based on morphology, forest composition and management history. Thus, anticipating the effects of global change on forest biomes are realistic objectives.

Recognition of the complex interdisciplinary nature of landform–forest interactions, and the significance of these linkages in the assessment of human impacts and global change, has been included in the principal themes of the Earth Systems Science Committee, established by the National (US) Aeronautics and Space Administration in 1986. These themes have been incorporated in the international Earth Observation System, and in global research designs for the International Space Station. Geomorphologists will continue to be integral members of ground-based research teams quantifying the linkages between terrestrial processes and forest ecosystems. The US National Academy of Sciences has established 'Long Term Ecological Reserves', with a minimum research planning term of two hundred years. Other nations have expressed similar plans, and a global network of sites, focused on major forest biomes, is a major scientific objective.

As forest geomorphology becomes established as a significant sub-field within the discipline, the need for interdisciplinary education has become apparent. Several sessions dedicated to forest geomorphology have been held by the International Association of Geomorphologists (IAG), and at least one formal graduate programme has been established.

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SEE ALSO: applied geomorphology; bar, river; channel, alluvial; climatic geomorphology; landslide; mass movement; sediment routing; step-pool system; threshold, geomorphic

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FORMATIVE EVENT

An important idea in geomorphological science is that morphological stability or stasis can be interrupted by brief, instantaneous, episodes of erosion or deposition when significant morphological change takes place (Erhart 1955; Butler 1959; Ager 1976; Gould 1982; Reading 1982; Dott 1983).

A second important idea is that all geomorphological processes are made up of discrete events of varying frequency, magnitude, duration and sequencing characteristics. If we are to understand landform change it can only be with respect to the characteristics of the events which cause change (Brunsden 1996).

An event is a period of activity of a process, at any place. Events may be classified according to their role in landform evolution. An effective event is one that exceeds the resistance or tolerance of a system and does work. Following Wolman and Gerson (1978) this is measured by the ratio of the event to the mean annual condition of erosion, denudation rate or deposition. Small but frequent events cause morphological change in a cumulative way. All that is required is time. A crucial component is the sequence in which events of different potential effectiveness occur. A very effective event may have considerable feedback effect on succeeding events. If all the available work has been done, later events may perform below their energy potential. If the effective event unlocks potential energy (e.g. by creating steep slopes) it may build in to the system further progressive and diffusive change. If the event prepares a threshold (see THRESHOLD, GEOMORPHIC) condition and is followed by another effective event there may follow unusual or rare forms of change.

It is therefore helpful to use the term formative event. A formative event is an event, of a certain frequency and magnitude, which controls the form of the land. If it does more work than the cumulative everyday event the landform it produces will persist (perhaps for long periods) despite the modifying effects of the more frequent events. It may require another formative event to obliterate the landforms produced or such an event may reinforce the effect. Multiple glaciation of a valley is an example, the 'U' form, once produced by a glacial 'event', may remain for millions of years, surviving all changes in the environmental controls. The word 'persistence' describes the length of time a landform survives as a diagnostic element of the landform assemblage.

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DENYS BRUNSDEN

FRACTAL

Sciences such as geomorphology are concerned with inherently variable phenomena – things that are not exactly predictable or repeatable because of the sensitive dependence on initial conditions that many geomorphological systems exhibit (i.e. CHAOS THEORY and the 'butterfly effect'), and because of the contingencies among the elements that make up the system (see SELF-ORGANIZED CRITICALITY). While detailed long-term predictions of such systems are impossible, we should be able to provide some statistical bounds within which the future (or past) should lie. Moreover, while actual events may be unpredictable, they are obviously not unexplainable. Fractals are fundamental components of the methods that are required when analysing or modelling such systems, methods that are amenable to complex, non-linear dynamical systems (see COMPLEXITY IN GEOMORPHOLOGY).

Consider that our primary evidence of the past lies in the patterns we observe today – be it on hillslopes or in river valleys. If the means by which we attempt to characterize those patterns are not able to capture the true complexities of the systems, then how are we to turn back the hands of time and develop an understanding of the Earth's past (Werner 1995)? Fractal patterns, chaos and self-organization can provide the null hypotheses against which process-based interpretations can be tested.

The field of fractals emerged primarily from the writings of one person – Benoit Mandelbrot (1967, 1982) – to become a mainstream field of research. Along with chaos theory and self-organization, it has dramatically altered our view of nature, and of geomorphology (Turcotte 1992).

Fractals are the unique patterns left behind by the unpredictable movements of the world at work. The branching patterns of rivers and trees, the coastline of Britain, the pebbles in a stream, the spatial distribution of earthquake epicentres – all these can exhibit fractal patterns. Fractal objects show similar details on many different scales. Imagine, for example, the rough bark of a tree viewed through successively more powerful magnifications. Each magnification reveals more details of the bark's rugosity. In many geomorphological phenomena, such as river networks and coastlines, this fractal self-similarity has long been observed (e.g. Burrough 1981). This means that, as we peer deeper into a fractal image, the shapes seen at one scale are similar to the shapes seen in the detail at another scale. Fractals are formally defined as objects that are self-similar (Baas 2002).

The measure which most people use to quantify fractal scaling and self-similarity is the fractal dimension or D. The fractal dimension is a number that reflects the way in which the phenomenon fills the surrounding space. The fractal dimension of an object is a measure of its degree of irregularity considered at (theoretically) all scales, and it can be a fractional amount greater than the classical geometrical dimension of the object. The fractal dimension is related to how fast the estimated measurement of the object increases as the measurement device becomes smaller. A higher fractal dimension means the object is more irregular, and the estimated measurement increases more rapidly. For objects of classical geometry, such as lines or curves, the geometric or topological dimension and the fractal dimension are the same. The quantification of fractal patterns led to the discovery that many phenomena, when plotted using appropriate transformations, can be described using a power law ($1/f$ systems).

An important concept tied to the fractal dimension is that of spatial autocorrelation. If nearby conditions on a surface are very similar to each other, then we call that positive spatial autocorrelation. If nearby conditions on the surface are the opposite, then we call that negative spatial autocorrelation. Spatial autocorrelation is zero when there is no apparent relation between nearby conditions. A low fractal dimension for a surface (e.g. 2.1) indicates that the self-similarity exhibits high positive spatial autocorrelation. A high fractal dimension (e.g. 2.9) indicates that the self-similarity exhibits high negative spatial autocorrelation. A fractal dimension in the middle of the range (2.5) indicates that no spatial autocorrelation exists. Brownian motion is a classic example of a fractal at the middle range – it is a process with zero memory of where it came from and no knowledge of where it will go next.

Although, theoretically, labelling something a fractal implies that it exhibits self-similarity across all scales, in fact most natural objects possess a limited form of self-similarity – between certain limits or resolutions, the object behaves in a fractal-like manner. These are often called fractal elements, and it is possible that an object may possess multiple fractal elements. Many scientists now consider the boundaries at which fractal behaviour is observed to be important, for those boundaries clearly distinguish process limits. However, does knowledge about the limits to the form of a phenomena necessarily allow us to make statements about the limits of the process which is responsible for creating that form? The answer to that question remains unanswered, although it is at the heart of most fractal research.

One of the main reasons for the increased interest in the fractal dimension (D) is the awareness that dissipative dynamical systems and fractal spaces (and time) are linked – that we now have a theoretical basis with which to link form (e.g. D) and process (e.g. self-organized criticality). The lack of such a link has long been one of the criticisms levelled at fractal studies (e.g. Mark and Aronson 1984), so the discovery that a link can be made is an important step forward in fractal research. However, while self-organized critical models developed in a computer have been very successful at mimicking many varied systems, the unequivocal existence of self-organized criticality in real systems has yet to be confirmed. Geomorphic concepts, such as negative feedback, static equilibrium, and the concept of the graded stream, are all similar to the concept of self-organized criticality. These existing concepts provide an explanation for many geomorphic phenomena without the need to invoke a mechanism such as self-organized criticality. Many geomorphometric measures also are not statistically related to the fractal dimension – that alone indicates that fractals are not capturing all the aspects of a landscape that geomorphologists consider important (Klinkenberg 1992).

Earthquakes and avalanches are two of the more visible manifestations of self-organized critical systems. Their statistical properties, such as size distributions, generally obey power-scaling laws – they follow a fractal distribution (Bak 1996). If a form is found to be a fractal form then certain statistical properties follow. A fractal form has no one scale dominant – it is scale invariant – and its second moment is theoretically infinite. Conversely, a form which is not scale-invariant, a 'Gaussian' form, can be completely described by a few statistical moments. A fractal form will be characterized by rare intermittent events that, from a process point of view, dominate the statistical record. Thus, one of the challenges that fractal studies are attempting to meet is the characterization of such statistically intractable events or forms (e.g. Xu *et al*. 1993). Furthermore, such statistical properties mean that obtaining enough data with which to compare the predictions of models against reality is a not an easy process (Baas 2002).

Fractal concepts have been applied extensively in fluvial geomorphology (e.g. Rodriguez-Iturbe and Rinaldo 1997); there are several different aspects which can be studied. The most obvious is: what is the true length of a river? One could also, while considering the entire basin, examine the form of the river network within the basin. At a higher resolution, the actual planform of the river can be considered (i.e. quantifying sinuosity). At these scales one must not only consider the river itself, but also the river valley form and its effects on the geometry of the river. Going even further down the scaling hierarchy, studies of the fractal characteristics of river bedforms can also be made.

It has been found that many allometric relations observed in nature are not dimensionally consistent (Church and Mark 1980). For example, dimensional analysis would conclude that the length of the mainstream channel of a river should be proportional to the square root of the area of the basin. Most studies have observed that, in fact, the mainstream channel length is proportional to the 0.6th root of the area. Mandelbrot interpreted that as a fractal finding: if a river meanders such that it has a fractal dimension of 1.2, then the length–area relation (known in the literature as Hack's relation) should be to the 0.6th power (1.2 divided by 2).

Power laws, which are the signature of fractals, have been experimentally observed over a wide range of scales in probability distributions describing river basin morphology. Some of the observed fractal distributions have been:

- Horton's power laws of bifurcation and length.
- Stream lengths follow a power-law distribution.
- The cumulative total drainage area contributing to any link follows a power-law distribution.
- The mean of the local slope of the links of a drainage network scales in a fractal manner as a function of the cumulative area.

The fact that 'fractal' rivers exist in so many regions implies that fractal growth processes occur in every environment.

If we accept that river networks and topography can sometimes be characterized as fractals, then we must question why that occurs and what processes are responsible. The simplistic explanation is that scale-invariant form is the result of scale-invariant processes (e.g. Burrough 1981). Does this necessarily mean that a scale-invariant process operates over all scales, as the scale-invariant spatial form appears similar over all scales? We know that this can't be the case – consider the processes such as chemical weathering, frost action and soil creep that operate only at the microscale level. Obviously, the assumption of a one-to-one correspondence between the scale of the form and the scale of the process cannot hold. Self-organization provides the means of getting around this assumption. Large- and small-scale spatial structures emerge through the operation of small-scale processes. Simple rules at one level can lead to complex behaviour at a higher level, behaviour which is referred to as emergent behaviour. We do not have to program in the complex behaviour; it just appears as a consequence of the actions of the agents at the smaller scale.

Fractals provide an out from the constraints of Euclidean geometry, and capture the patterns of nature in an intuitive way. Experiments have shown how our perceptions of roughness agree very well with the measured fractal dimension of the object. Fractal geometry has shifted research agendas: while strict quantitative measurement, measurement that values quantifiable features like distance and degrees of angles, is still important, it is now recognized that measures also need to embrace the qualities of things – their texture complexity and holistic patterning. Chaos, self-organization and fractals have allowed us to step away from simple linear deterministic models and step towards models which capture the essence of predictably unpredictable natural systems.

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BRIAN KLINKENBERG

FRAGIPAN

A natural subsurface horizon found deep in the soil profile, which has been altered by pedogenic processes responsible for restricting the entry of water and roots into the soil matrix. Fragipans possess a higher bulk density than the horizons above, contain very little organic matter, are brittle when moist and exhibit slaking properties when immersed in water. Thickness ranges from 15–200 cm, enough to allow sufficient impact upon plant growth so that roots and water are unable to penetrate 60 per cent of the horizon. Fragipans develop mostly in mid-latitude, medium texture, acid materials overlying albic or argillic soil horizons, and with udic or aquic moisture regimes. Fragipans occur mainly beneath forest vegetation, in cultivated or virgin soils within various parent materials including glacial drift, loess, colluvium, lacustrine deposits and alluvium, though they are not found in calcareous deposits. Fragipans consistently possess an abrupt upper boundary at a depth of 30–100 cm beneath the ground surface (Witty and Knox 1989), and often exhibit evidence of soil formation.

The origins of fragipans are poorly understood, though three main formation mechanisms exist. These are: physical ripening during desiccation of initially slurried material; clay bridging; and bonding by an amorphous component (including Si, Al, and Fe). Unfortunately, fragipan is a generally poorly defined term, with many examples of fragipans worldwide unidentified due to their vague definition in the field.

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STEVE WARD

FREEZE–THAW CYCLE

A freeze–thaw cycle is a cycle in which temperature fluctuates both above and below 0 °C. Field measurements indicate that most freeze–thaw cycles per year occur in the climates of low annual temperature range, which are dominated by diurnal or cyclonic fluctuations (French 1996: 26). These conditions are met in subpolar oceanic locations (e.g. Jan Mayen in the northern hemisphere, Kerguelen Islands or South Georgia in the southern hemisphere) and in intertropical high mountain environments (e.g. Andes, East Africa mountains). Among all cold environments, the least number of freezing and thawing days occur at high latitude and in continental climates, which are dominated by seasonal temperature regimes. In all areas, most cycles occur in the upper 0–5 cm of the ground and only the annual cycle occurs at depths in excess of 20 cm.

Freeze–thaw cycles have important effects on soils, like FROST HEAVE, frost sorting or frost creep. It is important to distinguish between seasonally frozen ground and PERMAFROST (Washburn 1973: 15). In non-permafrost environments, the depth of seasonal freezing increases with increasing latitude, the range being from a few millimetres to more or less 3 metres. In permafrost regions, the active layer is the upper part of the ground that undergoes seasonal freezing and thawing.

With respect to rock frost weathering (see FROST AND FROST WEATHERING), alternate freezing and thawing is much more damaging than continued cold, and the effectiveness of frost is dependent on the frequency of temperature fluctuations about the freezing point in the presence of water (Ollier 1984: 125). Nevertheless, the number of freeze–thaw cycles undergone by materials unfortunately cannot be used as a direct measure of frost action effectiveness for several reasons. First, the use of air temperatures to define cycles is not satisfactory at all, since significant differences exist between air and ground temperature. This can be caused for example by the insulating effect of snow or by insolation on dark rock surfaces (Washburn 1973: 58). Second, the exact freezing temperature across which the oscillations should be measured is difficult to define, as all the water contained in soils and rocks does not freeze instantaneously, nor always at 0 °C but at negative temperatures, because, for example, of the capillary forces existing in the porous media or the supercooling phenomenon. Freezing temperature can also be lowered in presence of salts or clay. Freezing has been reported to begin at temperatures lower than −10 °C in the case of rocks characterized by very small pores.

Finally, what constitutes a freeze–thaw cycle is debatable, as some authors define specific minimum negative temperatures that have to be reached for most of the rock-absorbed water to freeze, or minimum durations for the periods at negative and positive temperatures between successive cycles. For example, according to different studies, one cycle is completed when the hourly rock temperature changes from $\geq +1$ °C to ≤ -1 °C and then back to $\geq +1$ °C (Lewkowicz 2001: 359), or when a fall below −2 °C is followed by a rise above +2 °C (Matsuoka 1991: 276). Although these thresholds have been defined in order to take into account the actual stresses undergone by the rock as accurately as possible, they make any comparison of cycle frequencies reported in different studies very difficult.

Other important components of freeze–thaw cycles with respect to frost weathering are the duration of freezing (the time period during which negative temperatures persist), the intensity of freezing (the extent of temperature decrease below 0 °C) and the rate of freezing (the rapidity or slowness with which temperature decreases below 0 °C) (McGreevy and Whalley 1982: 158). The influence of these three parameters is quite controversial.

As far as the intensity of freezing is concerned, since the greatest part of pore water freezes between 0 °C and −5 °C, volume expansion causing frost weathering of rock occurs mostly in this temperature range (McGreevy and Whalley 1982: 159; Matsuoka 1991: 272). This explains why frost decay rates do not change significantly between freeze–thaw cycles reaching minimum temperatures of −8 or −30 °C.

The impact of freezing duration has to be viewed in relation to the intensity of freezing. It is the pore sizes that determine the freezing point of water within rocks. Thus freezing occurs over a range of gradually decreasing temperatures and rocks undergo some stress only if the required critical temperatures have been reached, and for a period long enough so that the temperature change propagates from the rock surface into the centre of a block or into a rockwall. There must indeed be time for the transfer of the necessary latent heat to cause the freezing or thawing of the water in the rock (Ollier 1984: 125). The duration of the period at minimum temperature has been considered by laboratory work as completely insignificant (under constant temperature and if the freezing front stopped progressing, no breaking strain can be built up) or quite important (in an open system with a constant unfrozen water supply, segregation ice lenses may keep growing by unfrozen water migrations under constant temperature conditions). On the other hand, in field studies carried out in alpine environments where wedging (see FROST AND FROST WEATHERING) of a massive rock mass is the predominant decay process, freezing intensity and duration have been considered as fundamental parameters as they are responsible for the depth reached by the freezing front. Only long freezing periods, with stagnation of the freezing front at depths between 10 and 50 cm, are able to furnish large slabs in addition to small blocks (Coutard and Francou 1989: 415).

Various rates of freezing can favour various weathering mechanisms and lead to different degrees and types of rock decay in the same rock type. Quick cooling favours bursting and wedging effects, as more pressure is built up in pores and cracks when no time is left for water migration to occur and to relieve some of this pressure. On the contrary, slow cooling offers optimal conditions for the formation of segregation ice lenses and for scaling effects. Numerous works report higher degrees of decay after quicker frosts although some studies argue that freezing rate is not a particularly critical parameter (McGreevy and Whalley 1982: 158), or stress on the quite complex impact of freezing rates, making the evaluation of its importance difficult (Matsuoka 1991: 272). According to Matsuoka, slow freezing in an open system results in prolonged water migration toward the freezing front and, hence, in rising ice force. In contrast, in a closed system, rapid freezing favours a large ice growth strain, because pore ice contracts with time.

Rates of freezing measured in natural environments generally range between 0.2 and 4 °C per hour. However, laboratory simulation usually favours quick cooling rates (in order to accelerate COMMINUTION and the achievement of decay results) and values higher than 10 °C per hour are not uncommon. Results obtained by such experimentation may not reflect natural environmental processes.

Freeze–thaw cycles have been the subject of data collection in the field and of laboratory investigations, testing the impact of different temperature regimes on frost susceptibility. A large variety of cycle characteristics have been used, but the two main types reflect a daily moderate freezing regime (down to −8 °C) characteristic of polar maritime regions and a more intense and prolonged freezing regime (down to −30 °C) characteristic of polar continental areas.

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SEE ALSO: experimental geomorphology; frost and frost weathering; mechanical weathering; periglacial geomorphology; weathering

ANGÉLIQUE PRICK

FRINGING REEF

The morphology and genesis of CORAL REEFS varies significantly. They may be divided into ATOLL reefs, barrier reefs and fringing reefs (Nunn 1994). The youngest and most ephemeral of the three forms are fringing reefs, which also often lack the breadth, continuity and species diversity of atoll and barrier reefs. In addition, because they are located nearest the land – and indeed cannot exist distant from it – fringing reefs are those which are usually most affected by humans.

Development of fringing reefs

Unlike atoll reefs and barrier reefs, most fringing reefs formed as discrete units only during the most recent period of postglacial sea-level rise. Most began growing from shallow depths on the flanks of a tropical coastline when ocean-water temperatures (and other factors) at the end of the glacial period became suitable for reef growth. Encouraged by sea-level rise, the nascent fringing reefs began growing upwards and exist as living entities today only if they were able to ‘keep up’ or 'catch up' with sea level during the transgression (Neumann and MacIntyre 1985).

Once sea level reached its maximum level during the Holocene (about 5,000 cal. yr BP), 'keep-up' fringing reefs would have stopped growing mainly upwards and would have begun growing laterally, an ecological transformation involving a change in coral species distribution. Branching corals in particular would slowly have been replaced by other species adapted to outward rather than upward growth. A classic study is that of Hanauma Reef which began growing about 7,000 years ago on the inner flanks of an ancient volcanic crater on the Hawaiian island Oahu (Easton and Olson 1976).

On the other hand, 'catch-up' fringing reefs would not by definition have been able to keep pace with rising sea level and may have 'caught up' only when sea level was falling during the late Holocene. In such cases the change from upward to outward growth may have occurred more recently.

The outward growth of a fringing reef is constrained by the slope angle of the coastline from which it rises. On steeply sloping coasts, like those of the central Pacific island Niue, it is no surprise that fringing reefs are barely noticeable (and have little role in shoreline protection), often no more than a few metres in width. On coasts which slope more gently, fringing reefs may reach several hundred metres in breadth and have well-defined morphological zones (see below).

Some writers like Davis (1928) believed that a fringing reef was part of a genetic continuity and would eventually become a barrier reef and finally, when the land from the flanks of which the reef rose was submerged, an atoll reef. This is valid in only a general sense but did not take into account the effects of sea-level changes and the fact that, at the end of each Quaternary glacial period, fringing reefs re-grew. Such writers often equated the presence of fringing reefs with a coastline that had just begun sinking and, where a barrier reef was found farther offshore, would often cite a complex series of tectonic (rather than sea-level) movements to explain the association.

Morphology of fringing reefs

Fringing reefs have morphological characteristics that are shared with atoll and barrier reefs and others which are not. Along their outer, submarine slopes, fringing reefs have slopes of talus derived from the mechanical erosion of the reef edifice. Owing to the youth of fringing reefs and the comparative shallowness of the adjoining seafloor (usually a lagoon floor), these talus slopes are generally less voluminous than the equivalent features off barrier or atoll reefs. Similarly, owing to the wave energy being generally less along the fronts of fringing reefs (because waves hitting fringing reefs are commonly generated within a lagoon or are residual waves reduced in amplitude from crossing a barrier reef), reef growth and coral diversity on the outer reef crest is generally less than on barrier or atoll reefs. Yet, where a fringing reef faces directly into the ocean, these features and others are of the same size as on barrier or atoll reefs. A good example is the south coast of Tongatapu Island in the South Pacific where the south-east trade winds drive swells straight onto the narrow fringing reef which has well-developed spur-and-groove morphology along its front and an impressively high algal (*Porolithon*) ridge (Nunn and Finau 1995).

Behind the outer reef crest of fringing reefs is generally found a reef flat several tens of metres broad in which there are comparatively few living corals but an abundance of fossil reef, often planed down from a higher level. A good example is from New Caledonia (Cabioch *et al*. 1995). Particularly if the fringing reef has been significantly affected by humans (see below) the back reef area may be covered with seagrass beds or the alga *Halimeda*, sometimes terrigenous sand, all of which inhibit reef growth and may in consequence reduce the supply of calcareous sand to adjacent shorelines.

At the back of many fringing reefs is a 'boat channel' eroded in the reef surface at the point where freshwater comes out of the adjacent land. Freshwater springs are common in such places.

Emerged fringing reefs

Along those coasts where coral-reef upgrowth was able to keep pace with postglacial sea-level rise, and the sea level exceeded its present level during the middle Holocene, it is expected that fringing reefs would have grown above their present levels and that remnants of such 'emerged' fringing reefs would now be visible to testify to this. The morphology of emerged fringing reefs is often comparable to that of their modern counterparts although many are much reduced by erosion.

In the Hawaiian Islands, for example, many years of searching for emerged fringing reefs bore fruit only quite recently (Grigg and Jones 1997).

Human impact on fringing reefs

Fringing reefs are those most vulnerable to deleterious human impact. Many bear the brunt of indirect impacts like pollution and sedimentation from adjacent land areas. Direct impacts, particularly along coasts where fringing reefs are central to subsistence economies or to recreational activities, include overexploitation of edible reef organisms, trampling by humans, physical damage from boat anchors, and even poisoning or dynamiting for easy kills of large numbers of reef fish.

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SEE ALSO: coral reef

PATRICK D. NUNN

FROST AND FROST WEATHERING

Frost action is a collective term describing a number of distinct processes which result mainly from alternate freezing and thawing of water in pores and cracks of soil, rock and other material, usually at the ground surface. It is widely believed that frost action is the fundamental characteristic of present-day periglacial environments. Frost-action processes probably achieve their greatest intensity and importance in such areas.

In soils, FROST HEAVE, NEEDLE-ICE formation, frost creep and thermal contraction cracking are very common frost-related processes. The term *cryoturbation* refers to all soil movements due to frost action (French 1996).

Frost weathering (also called *frost shattering, congelifraction, gelifraction* or *gelivation*) contributes to the *in situ* mechanical breakdown of rocks by various processes. The conventional view is that rock decay is due to the fact that when water freezes it expands by about 9 per cent. This creates pressures, calculated to be around 2,100 kg cm^{-2} at -22 °C, that are higher than the tensile strength of rock (generally less than 250 kg cm^{-2}). However, this process rarely induces critical pressures, reached only when freezing occurs in a closed system with a very high rock moisture content (about 90 per cent). Such conditions are not common in natural environments, but when occurring, the volume expansion effect may cause rock *bursting*.

A more realistic model, also applicable to soils, is the segregation ice model (Hallet *et al*. 1991), which treats freezing in rock as closely analogous to slow freezing in fine-grained soils. When water freezes in rock or soil, the ice nuclei attract unfrozen water from the adjoining pores and capillaries. Tensions are primarily the result of these water migrations to growing ice lenses (Prick 1997). Frost weathering is induced by the progressive growth of microcracks and relatively large pores wedged open by ice growth. In the segregation ice model, low saturation in hydraulically connected pores (open system) does not preclude water migration and crack growth. The detachment of thin rock pieces by the growth of ice lenses is called *scaling*.

Frost wedging refers to rock fracturing associated with the freezing of water in existing planes of weakness, i.e. cracks and joints. Wedging can be caused by the volume expansion of water turning into ice in cracks, or by hydraulic pressure. According to this second process, the freezing front penetration in a rockwall induces a freezing of the most external part of the crack first, creating a solid plug of ice. In depth, where the saturated crack is thinner (and the freezing point thus lower), some water can be trapped under pressure by the ice growing further in from the rock surface and so contribute to crack growth outwards and downwards. In both cases, the thinner the crack, the quicker and the more severe the frost has to be in order to cause a wedging effect.

The rate at which frost shattering occurs depends on climatic factors and rock characteristics. Among climatic factors, the most important ones are the number of FREEZE–THAW CYCLES and the availability of moisture. Some thermal characteristics of the freeze–thaw cycle can also have some importance, like the freezing rate or the duration of the freezing period.

The water availability in the environment and the rock moisture content are certainly the most critical elements for defining the susceptibility of this environment to frost action (Matsuoka 1990). Laboratory experiments have shown that the amount of disintegration in rocks supplied with abundant moisture is greater than that in similar rocks containing less moisture. For this reason, dry tundra areas and cold deserts may undergo less extreme frost weathering than moister environments.

If some particular locations are characterized by a continuous and abundant water supply (for example a block sitting next to a lake shore or to a melting snow patch), a large majority of blocks exposed in cold-climate environments experience neither close to saturation conditions (because of insufficient water supply), nor a dry state (intense drying is rare).

A critical degree of saturation can be defined as a threshold moisture level for each rock type (Prick 1997): only when moisture exceeds this level will the material be damaged by frost. This parameter reflects the influence of rock characteristics on frost susceptibility and defines the part of the porous medium that has to be free of water in order not to build up a breaking strain.

The nature and characteristics of the rock are indeed a crucial factor for frost susceptibility. Rocks such as tough quartzites and igneous rocks tend to be most resistant, while porous and well-bedded sedimentary rocks, such as shales, sandstones and chalk, tend to be least resistant. Among the rock characteristics influencing frost weathering, the most determinant ones are: the rock specific surface area, permeability, porosity, pore size distribution, and mechanical strength.

A large specific surface area (i.e. internal surface of the porous media) induces a larger contact area between rock and water and therefore enhances a higher susceptibility to frost decay. A high rock permeability, by allowing easy and quick water migration, prevents critical pressures to build up (Lautridou and Ozouf 1982). Rocks with a very poor porosity are not frost susceptible: experimentation showed that rocks with a porosity of less than 6 per cent are little damaged after several hundreds of freeze–thaw cycles (Lautridou and Ozouf 1982); further research showed that this threshold value can be considered as a valuable, but rough estimate.

Pore size distribution (also called *porosimetry*) can influence frost susceptibility in various ways. Rock porous media characterized mainly by large pores (macroporosity) will tend to be frost resistant, as macroporosity favours a good permeability. Unimodal porous media (i.e. characterized by one predominant pore size) offer ideal conditions for segregation ice formation; rocks with such a pore size distribution tend to be susceptible to any type of freezing (even with a moisture content far below saturation and with a slow cooling rate) and will undergo an increased decay as freezing/thawing goes on. Multimodal porous media (i.e. characterized by pores of various sizes) are not favourable to the set up of large-scale water migrations; rocks with such a pore size distribution tend to be frost susceptible only with high moisture content, preferably in the case of a quick freezing.

Among ROCK MASS STRENGTH parameters, tensile strength has a considerable influence on rock frost decay (Matsuoka 1990). Crack density and width often influence water penetration in the bedrock and allow wedging to take place.

Frost action is one component of *cryogenic weathering*, i.e. the combination of weathering processes, both physical and chemical that operate in cold environments either independently or in combination. Many aspects of cryogenic weathering are not fully understood, but the processes other than frost that may be efficient decay agents are: HYDRATION (see WETTING AND DRYING WEATHERING), thermal fatigue (see INSOLATION WEATHERING), SALT WEATHERING, CHEMICAL WEATHERING, ORGANIC WEATHERING and PRESSURE RELEASE (particularly in recently deglaciated areas). Solutional effects are present in limestone and KARST terrain exists in PERMAFROST regions. The dominance of frost action among these processes is considered as doubtful, but the definition of the exact role of each of these processes in the different cold environments and in the different periods of the year is problematic.

Frost weathering characteristically produces angular fragments of various sizes. In periglacial areas, cryogenic weathering determines the formation of some extensive features like blockfields (see BLOCKFIELD AND BLOCKSTREAM), GRÈZE

LITÉES, SCREES, TALUS slopes or ROCK GLACIERS. Its action is also often crucial for MASS MOVEMENT processes like rock avalanches and rock falls.

The predominant size to which rocks can be ultimately reduced by frost action is generally thought to be silty particles with grain sizes between 0.01 and 0.05 mm in diameter. Experimentation on mineral particles indicated that frost weathering occurs within the layer of unfrozen water adsorbed on the surface of these particles. The minerals' susceptibility to weathering depends not so much on their mechanical strength as on the thickness and properties of this unfrozen water film. Decay occurs when this water film becomes thinner than the dimensions of the microcracks and defects that characterize the surface of mineral particles. The protective role of the stable film of unfrozen water is highest with silicates, such as biotite and muscovite, and lowest with quartz. Experimentation results indicate that under cold conditions the ultimate size reduction of quartz (0.01–0.05 mm) is smaller than for feldspar (0.1–0.5 mm), a reversal of what is assumed for temperate or warm environments (Konishev and Rogov 1993).

Frost weathering is studied both in the field and in the laboratory. The most commonly used techniques are: visual observation and photographic documentation of the decay evolution, weight loss, frost-shattered debris characterization, assessment of mechanical properties (like tensile or compressive strength) or elasticity properties (Young's modulus), ultrasonic testing, evolution of porosity and pore size distribution, dilatometry, and crack opening assessment.

Some field studies have been undertaken with the aim of increasing the availability of data upon rock temperature and moisture content. The lack of such data has up to now been a considerable impediment to a definition of the exact role of frost action in cryogenic weathering and to the realization of laboratory simulation using thermal and moisture regimes likely to occur in natural environments. Other studies focus on the rate of bedrock weathering by frost action (Lautridou and Ozouf 1982) and on the definition of predictive models (Matsuoka 1990). Laboratory simulation and modelling identify the climatic conditions and rock characteristics that emphasize frost action efficiency and so define the exact role of the various weathering mechanisms (e.g. Hallet *et al*. 1991; Prick 1997).

A major gap remains between field and laboratory research (Matsuoka 2001). This is due to a difference in the size of the study object (small blocks in the laboratory, but rockwalls sometimes in the field), in the type of rock material (intact soft rocks with medium or high porosity are overrepresented in laboratory simulations, but jointed massive rocks with low porosity are very common in cold environments) and thus in the type of frost weathering process taken into account (mostly bursting or scaling in the laboratory simulations; mostly wedging in the field). Wedging may sometimes be the only frost weathering process acting on fractured rock characterized by a low porosity and a high mechanical strength for the individual blocks. This may lead to *macrogelivation*, i.e. frost weathering at a large scale, acting mainly through the crack system, as opposed to *microgelivation*, which refers to frost decay acting in the porous media of individual small-sized blocks. This further illustrates the inadequacy of a simplistic view of frost weathering.

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SEE ALSO: experimental geomorphology; mechanical weathering; periglacial geomorphology; weathering

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FROST HEAVE

Frost heave is best known from the wintertime uplift of the ground surface, familiar to dwellers in cold climates, which is evidenced by jammed gateways, uneven roads, cracked foundations and the breaking-up of road surfaces in the spring thaw. These effects are not ascribable to the expansion of water that occurs on its freezing (9 per cent). They are due to the movement of water into the soil that is freezing, with the formation of accumulations of ice – increasing the soil volume, giving displacement (the 'heave'). These ice structures are called 'lenses', 'schlieren' or ice 'masses' and known collectively as segregation ice (because each is larger than pore size and has been segregated from the soil pore structure). Segregation ice is not ice from entrapped snow, buried glacial ice or buried lake or marine ice, although it may reach cubic metres in size. Its nature and extent depends on the nature of the granular soil material and a variety of local factors (drainage conditions, temperature regime, depth in the ground, etc.). Thus frost heave is commonly uneven, giving rise to irregularities of the ground surface (bumpiness) usually recurring year after year or, in PERMAFROST, persisting for many years. The forces generated by the heaving material can be very large.

Segregation ice and thus frost heave is an expression of the fundamental thermodynamic behaviour of a porous medium on freezing; this thermodynamic behaviour is ultimately responsible for the main properties and characteristics associated with soils in cold climates. As a consequence frost heave has enormous economic (geotechnical) significance; overcoming its effects is the essential problem for construction of buildings, roads, airports and pipelines in the cold regions.

The processes associated with frost heave largely explain the origin of most terrain forms occurring naturally and characteristically in cold climates – so-called 'periglacial' (see PERIGLACIAL GEOMORPHOLOGY) features. Boulders ('growing stones') are heaved to the ground surface by annual cycles of freezing and rearrangement of soil particles at thawing. Incremental frost heave is an important process in formation of PATTERNED GROUND, such as stone circles and stone polygons, where stones and boulders are heaved in particular directions (as a function of temperature and other factors) to give rise to conspicuously ordered surface arrangements. PINGOs, features occurring locally in regions with permafrost, look like volcanic cones and are elevated by the large, hidden central core of ice. They are the product of a particular thermal regime, commonly involving the gradual freezing of previously unfrozen ground below a receding water body. The frost heave process is largely responsible for lifting the above-surface material in pingos to elevations of tens of metres, so the forces developed must be large.

The instability of slopes, and the development of certain forms of SOLIFLUCTION, mudflows and landslides are ascribable to the excess water released on thawing of frost-heaved soils with their ice segregations, and the associated high PORE-WATER PRESSURES. Not infrequently the volume of segregated ice exceeds the volume of water the soil can hold in the thawed state by a factor of two or more. This accounts for a greatly weakened state of the newly thawed soil.

Fundamentally, the water moves toward a zone of freezing in the soil because of thermodynamic potentials arising with the growth of ice crystals in small spaces. Although the thermodynamic principles have been recognized for more than a century (and also describe, for example, crystallization phenomena in solutions, the formation of ice crystals or of water droplets in the atmosphere, or the nucleation of bubbles in liquids) the significance for soils has been realized fully only in recent decades. The thermodynamic potential may be regarded, with some simplification, as a pressure, and is referred to by different terms in different branches of science and technology. The pressure of the water falls with temperature in freezing soil, so that there is a gradient from warm (unfrozen) to cold (frozen). However, the pressure of the ice in the ice segregations rises as the temperature falls, and it is demonstrably this pressure which causes frost heave. Furthermore, the thermodynamic relations require ice to form in larger spaces and pores first. As a consequence, there is an unfrozen water content of frozen soils, decreasing with temperature as progressively smaller pores are filled by ice, and which is, therefore, a function of pore size distribution.

The significance of the soil water accumulating (the process of frost heave) and then freezing in this way over a range of temperatures down to several degrees below 0 °C, is that the thermal and mechanical properties of the frozen soils are highly temperature dependent. Frozen, heaved soils are prone to creep in a manner rather similar to glaciers but with lower rates of deformation; this is probably the cause of certain large vegetation-covered solifluction terraces on slight slopes. The grain-size and pore size distribution of a soil are crucial to its behaviour when frozen because they control the (unfrozen) water content of the specific soil. The release of latent heat of freezing of the water effectively controls the heat capacity of the soil; thermal conductivity is also modified (though less so) because of the difference in thermal conductivity between ice and water. The thermal diffusivity, which is the ratio of thermal conductivity to heat capacity, is consequently highly temperature dependent and controlled by the pore size distribution – that is, by the nature of the soil (clay, silt or sand, or combinations of particle sizes), and the amount of frost heave.

The thermal diffusivity controls such phenomena as the depth to which winter freezing occurs, and the depth of summer thawing (the ACTIVE LAYER) above permafrost. Indeed the distribution of permafrost itself (ground remaining frozen year in, year out), in depth and in time (and in response to climate or microclimate change), depends substantially on its thermal diffusivity. Terrain features, ascribable to frost heave and associated with the comings and goings of permafrost, include ALASes, palsa and THERMOKARST.

Counteracting effects of frost heave and subsequent thawing added billions of dollars to the cost of the transAlaska oil pipeline. The forces generated (CRYOSTATIC PRESSURES) by frost heave around gas pipelines in permafrost regions threaten their stability and thus their financial viability. Avoidance of frost heave is the main reason for added costs of infrastructure in general in the cold regions; these added costs are greatest in the 25 per cent of the Earth's land surface underlain by permafrost but are also a major factor in construction (especially of highways and airports) in the further 20 per cent or so which has significant winter frost penetration, and consists largely of highly populated temperate lands.

When Taber (1918) first clearly demonstrated that the geotechnical problem of frost heave was due to the migration of water with accumulation of excess ice in the frozen ground, he paved the way for Beskow's classic work (1935) on frost heave and its significance in relation to the local environment (soil type, groundwater conditions, confining pressures, etc.). In 1943 the remarkable study by Edlefsen and Andersen (resulting from the wartime collaboration of two scientists in different fields) established the thermodynamic interpretation, which substantiates the largely empirical approach that has been used by geotechnical specialists concerned with engineering (Andersland and Ladanyi 1995) for cold regions development in the broadest sense. Agronomists too, have an important involvement. Today, geocryology, the study of the ground surface regions in freezing climates (Williams and Smith 1989) notably developed in Russia (Yershov 1998), is concerned mainly with the effects of frost heave, a phenomenon first recognized some two hundred and fifty years ago (see Beskow 1935).

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PETER J. WILLIAMS

G

GENDARME

A needle-shaped rock pinnacle located along a mountain ridge or arête. The term gendarme is universal, yet employed predominantly in alpine geomorphology and mountaineering. Gendarme shares its name with a French policeman, as both may block one's passage and hinder progress. They are commonly found in the Alps, such as Pic de Roc gendarme in Chamonix, French Alps. However, similar forms exist in other mountainous regions, such as Bryce Canyon, USA. Gendarmes are also referred to as rock pinnacles and aiguilles, yet are generally more pointed and larger than an aiguille.

SEE ALSO: rock and earth pinnacle and pillar

STEVE WARD

GEOCRYOLOGY

The study of Earth materials having a temperature below 0 °C. Washburn (1979) recognized that it was sometimes taken to include glaciers, but argued that it was more specifically used as a term for PERIGLACIAL GEOMORPHOLOGY and PERMAFROST phenomena. Indeed, the subtitle of his magisterial volume *Geocryology* was 'A survey of periglacial processes and environments'. In that volume he studied such phenomena as frozen ground, FROST AND FROST WEATHERING, PATTERNED GROUND, avalanches (see AVALANCHE, SNOW), SOLIFLUCTION, NIVATION and THERMOKARST.

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A.S. GOUDIE

GEODIVERSITY

The geodiversity concept first appeared in Australia (especially Tasmania), and received wider recognition, even if always not proper understanding, in the mid-1990s. This robust geodiversity concept has been poorly developed yet in methodological terms. The most popular definition of geodiversity was put forward by the Australian Natural Heritage Charter (AHC 2002):

> Geodiversity means the natural range (diversity) of geological (bedrock), geomorphological (landform) and soil features, assemblages, systems and processes. Geodiversity includes evidence of the past life, ecosystems and environments in the history of the earth as well as a range of atmospheric, hydrological and biological processes currently acting on rocks, landforms and soils.

Geodiversity is now being used in a very holistic way to emphasize the links between geosciences, wildlife and people in one environment or system. The above definition can be supplemented with the statement that geodiversity also embraces quantitative and qualitative topics or indicators at any timescale which make it possible to distinguish marked peculiarities of a georegion, a spatial unit of an unspecified taxonomic rank. This means that bedrock, landforms and soils can be classified by at least two important categories: uniqueness and representativeness. From the geomorphological diversity perspective, an outstanding landform is a feature which is rare, unique, an exceptionally well-expressed example of its kind, or otherwise of special importance within a georegion. A representative landform, in turn, may be either rare or common, but is considered significant as a well-developed or well-exposed example of its kind. A landform type or system can be characterized by an isotropic entity in terms of topographic shape, physical contents, morphogenetic controls and processes, as well as time of formation.

The term geodiversity is commonly used in two meanings, simpler and broader. The first refers to the total range, or diversity, of geological, geomorphic and soil phenomena, and treats geodiversity as an objective, value-neutral property of a real geosystem. In this case a statement of the diversity is made, but the geosystem is not assessed in terms of what kind of geodiversity it is: low or high? The other usage conveys the idea that geodiversity refers specifically to particular geosystems that are in themselves diverse or complex, and thus does not apply to systems which are uniform or have low internal diversity. An example can be the valley of a river flowing through mountains, uplands and lowlands, filled with a wealth of valley, channel and bedforms, and therefore showing high geodiversity, whereas an area of lowland without any streams, basins and/or hummocks has low geodiversity. Questions about the measure of geodiversity are troublesome. Which area displays higher geodiversity: one in which there are 15 mogotes, or another featuring 5 volcanoes, 5 glaciers and 5 river valleys? Or another: has geodiversity increased or diminished in an area transformed by numerous and extensive man-made changes? Landforms are defined by their surface contours and that is why some people claim that the disturbance of significant landform contours (e.g. by excavation) will by definition degrade their geodiversity values, while others see this morphological disturbance as enrichment of geodiversity. Obviously, this situation calls for some clear-cut criteria of geodiversity. One of the possible solutions is a hierarchical classification of landforms: morphoclimatic zone (polar), morphogenetic zone (mountain), morphosystem (glacial system), type of relief (depositional relief), set of landforms (morainic landforms), and single form (terminal moraine). This classification is a function of complexity (see COMPLEXITY IN GEOMORPHOLOGY) reduction. One might argue that an increase in complexity entails an increase in geodiversity, and variations in this relationship are a matter of two functions: asymptotic and exponential.

Because geodiversity is valuable from a variety of perspectives (intrinsic, ecological, geoheritage, as well as scientific, educational, social, cultural, tourist, etc.) it should undergo geoconservation as a result of which it is possible to create GEOSITES for present and future generations.

It should be added that the term geodiversity is analogous to the term biodiversity, which is used to denote species, genetic and ecosystem diversity. It is important to note that the only analogy is that both involve a diversity of phenomena and beyond this self-evident similarity, no further analogies between the nature of ecological and geomorphic processes are expressed or implied. For example, both processes contrast strongly in their timescales. Ecosystems with plant or animal life cycles of tens to hundreds of years do not closely parallel the much longer term active or relict geosystems with weathering, erosion and sedimentation, or Earth internal processes such as seismic or volcanic activity and plate tectonics controlled by processes acting over many thousands or millions of years.

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ZBIGNIEW ZWOLINSKI

GEOINDICATOR

The concept of geoindicators was put forward by the International Union of Geological Sciences in 1992. The task of the IUGS working group was to draw up an inventory of indicators to be measured and evaluated under any programme of abiotic environment monitoring. The inventory is not supposed to be a universal standard, but rather to provide a list for the selection by environment monitoring teams of those indicators that can be usefully employed with reference to their study area and time period. Thus, while the list of twenty-seven geoindicators is finite, their choice for the description of environmental change is free. Each geoindicator was evaluated relative to a set of checklist parameters: name, description, significance, human/natural cause, applicable environment, types of monitoring sites, spatial scale, method of measurement, frequency of measurement, limitations of data, applications to past and future, possible thresholds, key references, other information sources, related issues and overall assessment (see Table 20).

Table 20 Geoindicators: natural* vs. human influences**, and utility for reconstructing past environments***

Geoindicator	N*	H**	P***
Coral chemistry and growth patterns	High	High	High
Desert surface crusts and fissures	High	Moderate	Low
Dune formation and reactivation	High	Moderate	Moderate
Dust storm magnitude, duration and frequency	High	Moderate	Moderate
Frozen ground activity	High	Moderate	High
Glacier fluctuations	High	Low	High
Groundwater quality	Moderate	High	Low
Groundwater chemistry in the unsaturated zone	High	High	High
Groundwater level	Moderate	High	Low
Karst activity	High	Moderate	High
Lake levels and salinity	High	High	Moderate
Relative sea level	High	Moderate	High
Sediment sequence and composition	High	High	High
Seismicity	High	Moderate	Low
Shoreline position	High	High	High
Slope failure (landslides)	High	High	Moderate
Soil and sediment erosion	High	High	Moderate
Soil quality	Moderate	High	High
Streamflow	High	High	Low
Stream channel morphology	High	High	Low
Stream sediment storage and load	High	High	Moderate
Subsurface temperature regime	High	Moderate	High
Surface displacement	High	Moderate	Moderate
Surface water quality	High	High	Low
Volcanic unrest	High	Low	High
Wetlands extent, structure and hydrology	High	High	High
Wind erosion	High	Moderate	Moderate

Source: After ITC (1995)

From the point of view of geomorphology, especially dynamic geomorphology, the geoindicator concept seems to be particularly well-suited to determine changes in morphogenetic and sedimentary environments or, broadly speaking, in geosystems. Just like systems theory or allometric analysis, the geoindicator concept has also been adapted from biological sciences. Geoindicators are measures of surface and near-surface geological processes and phenomena that tend to change significantly in less than a hundred years, and which supply crucial information for estimating the state of the environment. This definition specifies the time interval concerned as under a hundred years, which means that geoindicators embrace those processes and phenomena that are highly variable at a short timescale. Hence geoindication will not cover processes involving slow change, like metamorphism or large-scale sedimentation. Geoindicators should answer such questions as, e.g.:

- How often does a process occur?
- What is the rate of river load transport?
- How stable is an individual landform?
- Is the given landform still active, or is it a remnant of an earlier developmental stage?

This way of question formulation determines the specific character of geoindicators: they can express the magnitude, frequency, and rate and/or behaviour trend of an event, process or phenomenon. This means that geoindicators can have widespread application in present-day geomorphological research and, when backed up by paleoenvironmental research, they can provide an excellent basis for forecasting studies. It is especially important when one considers the last decades with their climate change and the consequences it has for the operation of most geosystems throughout the globe. This characterization of geoindicators can be extended to include interactions between the abiotic and biotic environments as well as the fact that it is possible to use geoindicators for different-sized areas to measure extreme, secular and predominant events and to observe natural and man-made processes. Altogether, geomorphologists will find that they have acquired a research tool which is bound to bring about methodological changes in their field.

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ZBIGNIEW ZWOLINSKI

GEOMORPHIC EVOLUTION

Geomorphic evolution at its simplest means the mode of change of landform or geomorphic system over time. Qualitative theories continue to dominate geomorphology but a quantitative theory of landform evolution is becoming a central challenge. Traditional qualitative models of landform evolution include the geographical cycle (Davis 1899), Penckian morphological analysis (Penck 1924), the semi-arid erosion cycle (King 1962) and climatogenetic geomorphology (Büdel 1977). These four models represent the framework and options within which landscape evolution were considered from about 1890 until the 1960s. Each of them (except for Penckian morphological analysis) is still in vogue among those who are interested in landscape evolution at the regional scale. The geographical cycle (Davis 1899) is still widely celebrated as a uniquely effective pedagogic device. The orderly evolution of landscape through the stages of youth, maturity and old age, and its interruption at widely separate points in time by massive tectonic uplift, is intuitively appealing. Davis claimed that his model embraced the five factors of structure, process, stage, relief and texture of dissection, but much of the literature says that he considered only the first three. The single major problem with this model was the complete absence of field measurements to confirm or reject assumptions in the model. Nevertheless, there are few better models available to interpret, in a qualitative way, the massive erosional unconformities of, for example, the Grand Canyon of the Colorado River. The geographical cycle has never been proven wrong, but it has been bypassed rather than replaced.

Lester King's subaerial cycle of erosion (King 1962) is perhaps the only serious competitor with the geographical cycle as an interpreter of large-scale, low gradient erosion surfaces. King, strongly influenced by his observations on African escarpments and plateau surfaces, framed his model around the notion of the parallel retreat of scarps. He also attempted, with debatable success, to link his model with the global plate tectonics framework that was evolving during his productive career. His concept of CYMATOGENY (arching of extensive land surfaces with little rock deformation) was a necessary addition to the traditional concepts of orogeny and epeirogeny, and flies in the face of conventional plate tectonics, where massive horizontal movements are favoured. His attempts (King 1962) to correlate pre-Tertiary erosion surfaces globally have met with little debate (except for the critique of cymatogeny) because so few geomorphologists are working at this scale.

A third interesting model of geomorphic evolution is provided by Julius Büdel (1977) and known as the climatogenetic model. The major elements of this model have been interpreted for English readers by Hanna Bremer. The underlying premise is that landscapes are composed of several RELIEF GENERATIONS and the challenge is to recognize, order and distinguish these relief generations It is unfortunate that the major references in the English literature have been sceptical of the model and have failed to give a balanced review (Bremer 1984). Twidale (1976) provides a refreshing summary of the relevance of etchplains (see ETCHING, ETCHPLAIN AND ETCHPLANATION) in Australian landscape evolution.

The Penckian model (Penck 1924) was called morphological analysis. The underlying premise of his analysis is that the rate of uplift, and variations in that rate of uplift over time, dictate landform evolution. His ideas were not taken seriously in

Germany, but they were widely promulgated in Anglo-America because of Davis's interest and opposition to the model. Details of the slope processes discussed are hard to verify and understand because of the lack of field data. But in its championing of endogenic processes and its time independent emphasis, this model was strongly differentiated from the first three. Time-independent models (in which the idea of evolution sits uncomfortably) have been promoted by G.K. Gilbert (1877) and J.T. Hack (1960).

The dichotomy between historical evolutionary studies and functional geomorphology implies that these two approaches do not fit easily together. Indeed, Bremer has said that geomorphology is developing along two lines: the origin of landforms is primarily being studied in continental Europe with climatogenic or tectogenic causes in the foreground. In the English-speaking world the study of geomorphic processes prevails.

Discussions by Schumm (1973), Twidale (1976), Brunsden (1980; 1993) and Ollier (1991) have attempted to reconcile these apparently contradictory positions within a largely qualitative dialogue. The essential contributions to this more recent discussion are the concept of geomorphic thresholds and complex geomorphic response (Schumm), formative events, relaxation time and landform persistence (Brunsden), the understanding that pre-Tertiary landscapes are still decipherable (Twidale), the importance of reconciling plate tectonic theory and morphological evidence (Ollier) and the disequilibrium of all landscapes influenced by Quaternary glaciation (Church and Slaymaker 1989).

A quantitative theory of landform evolution, by contrast with the theories discussed above, requires that the storage and flux rates of water, its flow paths and pressure fields be quantitatively related to their controls and that the boundary conditions of climate, rock properties, topography and stratigraphy be known. But by far the bulk of research on geomorphic evolution has taken place at meso- and micro-scales. And this is where the basic disjuncture in geomorphic thinking has been most evident. Systems modelling and mathematical modelling has tended to drive geomorphic discussion towards the smaller scale landforms, and geomorphic evolution has become, for example, slope evolution, or channel evolution or shoreline evolution.

The work of Ahnert (1967 et seq.) and Kirkby (1971 et seq.) is instructive in that they have been able to satisfy the requirements of quantitative theory by limiting the scale of their models and establishing precise boundary conditions to simulate real world slopes and basins. From 1967 to 1977, Frank Ahnert developed a series of models that used empirical equations to deal with possible ways of relating waste production, delivery and removal at a point on a slope. His final model was a three-dimensional process-response model of landform development. From 1971 to the present, Kirkby has developed increasingly integrated models of slope and drainage basin development, many of them using differential equations that constrain mass balance and thereby maintain continuity These models have had difficulty in dealing realistically with such phenomena as landslides (too rapid) and storage accumulation (too slow), but they represent the cutting edge of modelling in geomorphology from a slope process perspective.

Hydrogeomorphologists, such as Dunne, Dietrich, Montgomery and Church, have led the movement from micro-scale modelling of fluvial process towards a meso-scale modelling of drainage basins, in which they couple slope and channel processes and exploit the drainage network properties to produce more realistic dynamic drainage basin models. Howard (1994) poses a series of critical questions around the landscape modelling project. What is the simplest mathematical model that will simulate morphologically realistic landscapes? What are the effects of initial conditions and inheritance on basin form and evolution? What are the relative roles of deterministic and random processes in basin evolution? Do processes and forms in the drainage basin embody principles of optimization and, if so, why? Is there some basin characteristic form that is invariant in time even under a change in the relative role of the chief land-forming processes. The development of drainage basins requires at least two superimposed processes. He called them soil creep and water flow; in the language of the modellers, one must be a diffusional creep-like mass wasting process capable of eroding the land surface even for vanishingly small contributing areas. Such a process requires an increase in gradient downslope because of its loss of efficiency with increasing area.

The other is an advective fluvial process that increases in efficiency with increasing contributing area. The interplay of these processes produces a combination of convex and concave landforms. By enforcing continuity of flow and continuity of sediment through a coupled system of partial differential equations, the rate of change of elevation can be made dependent on the net flux of sediments as forced by linear increase in discharge. This fundamental step in the understanding of the self-organization of landscape depends on the coupling of the developing landscape with flow rate.

Willgoose *et al.* (1992) presented a catchment evolution model that was essentially a process-response model sensitive to the erosional development of river basins and their channel networks. The model describes the long-term changes in elevation with time that occur in a drainage basin as a result of large-scale mass transport processes. The mass transport processes modelled are tectonic uplift, fluvial erosion, creep, rain splash and landslides. Individual landslides are not modelled but the aggregate effect of many landslides is. The model explicitly differentiates between the part of the basin that is channel and the part that is hillslope. A channel initiation function provided by Dietrich (Dietrich *et al.* 1992) defines a threshold beyond which a channel is formed.

Both dynamic equilibrium and transient states can be modelled in this way. Howard (1994) has noted that the erosion, transport and depositional processes, especially in the river channels, have been greatly oversimplified in the Willgoose *et al.* model and he has generated both alluvial and non-alluvial channel versions of his own model. A more fundamental criticism is that the model does not clarify the linkages of fundamental aspects of the dynamics and the existence of general scaling relations in the network and the landscape itself. Hence the search for improved understanding through analysis of the fractal characteristics of river basins, particularly scale invariance, self-similarity and self-affinity. Multifractality has become a valuable property to identify changing domains of specific process sets (Montgomery and Dietrich 1994).

Understanding of the variety of modes of geomorphic evolution at a variety of spatial and temporal scales is the best evidence of progress in the field. For a number of years at the beginning of this century, researchers were expected to adopt a single model and to stick with it. As a result, the field stagnated under the influence of a single paradigm. In the contemporary state of geomorphology, one of the large issues within models of evolution at the site and basin scale relates to the relation between deterministic and probabilistic modelling.

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SEE ALSO: dynamic geomorphology; fractal; geomorphology

OLAV SLAYMAKER

GEOMORPHOLOGICAL HAZARD

A significant practical contribution of geomorphology is the identification of stable landforms and sites with a low probability of catastrophic or progressive involvement with natural or man-induced processes adverse to human occupance or use. Hazards exist when landscape developing processes conflict with human activity, often with catastrophic results. People are killed and property is destroyed or damaged by extreme geomorphic events, and the toll has become greater as human activity has stretched to areas that were avoided in the past. As the population of the Earth has more than doubled from the three billion of 1960, annual losses due to disasters have grown more than ten fold (Bruce 1993).

Tragic examples abound: in 1970 a cyclonic storm surge pushed three to five metres of water into the low deltas entering the Bay of Bengal. The surge, and riverine flooding caused by discharge blocking, resulted in the deaths of an estimated 300,000 to 500,000 people in Bangladesh; an earthquake-induced debris flow descending the flanks of Mt. Huarascan that same year buried over 25,000 people in Peru; a 1991 storm-induced mudflow overwhelmed a concrete drainage channel, killing an estimated 7,000 people in the Philippines; and despite more than fifty years of comprehensive flood control, the 1993 floods along the Mississippi River were the costliest in American history.

Geomorphologists are increasingly engaged in the mapping and modelling of geophysical, hydrological and surficial material characteristics which expose areas to rupture, failure, fire, inundation, drought, erosion or submergence. Coupled with land use and human infrastructure analyses they examine the location, value, exposure and vulnerability of the human environment to hazard damage (see ENVIRONMENTAL GEOMORPHOLOGY). When the population density and demographics are added, potential casualties and emergency services needs may be forecast. This requires the integration of social scientists who focus upon social, technical, administrative, political, legal and economic forces which structure a society's strategies and policies for risk management (i.e. prevention, mitigation, preparedness, prediction and warning, and recovery), public awareness, emergency training, regulation and social insurance. Such a comprehensive approach would have been nearly inconceivable in the past, but with the advent of computerized geographic information systems (GIS) such mapping, modelling, and decision support systems are becoming more commonplace (Carrara and Guzzetti 1995).

Perplexing questions centre upon the apparent increase in frequency of catastrophic geomorphic hazards. Accurate statistical analyses of such infrequent occurrences require an observation period of well in excess of a century (Berz 1993), while consistent reporting of most types of disasters have a much shorter history. Monitoring techniques and measurement scales (e.g. Richter, Beaufort), remote sensing, and communications have only recently allowed the global reporting of events in comparable terms. A study of volcanic eruptions by Simpkin *et al.* (1981) concludes that the reported increase in volcanic activity over the past 120 years is almost certainly due to improved reporting and communications technology; they even report a reduction in 'apparent' activity during the two world wars. Despite the growing influence of global databases, scientific consortia, and the news media, additional factors may be influencing the growing number of reported disasters.

While considerable scientific debate lingers around the issue of global climatic change, geomorphologists are well aware of other indirect effects of human activity (Rosenfeld 1994b). Certainly the deforestation of large areas has caused landslides and increased both the frequency and peak flows of flood events in many areas, while overgrazing has accelerated drought effects and erosion. Groundwater withdrawal and irrigation diversions have affected natural vegetation and micro-climates of some regions, and have even induced earthquakes in some cases.

As climate models become increasingly realistic, their mathematical results consistently point to a more hazardous world in the future. That increasing concentrations of greenhouse gases in the atmosphere, primarily resulting from the burning of fossil fuels, is changing the radiation balance and perhaps the climate is consistent with recent disaster experience. There is general agreement among atmospheric scientists that a 'warmer' world would be a 'wetter' world, with no increase in the number of days with rain, but with more intense rainfall. Combined with the hydrologic effects of land use changes, the frequency and severity of floods would surely increase, especially in the monsoon climates of south Asia where flooding already reaches catastrophic proportions. Drought effects in sub-Saharan Africa, South America and Australasia could occur more frequently and be more severe as a result of intensified El Niño–Southern Oscillation events. Resultant sea-level rise could pose additional storm surge or tsunami risk to heavily populated low-lying coastal regions such as lower Egypt, Bangladesh and many Pacific islands, along with the loss of most freshwater resources in the latter case. At higher latitudes, global warming may induce profound effects upon the human use and occupancy of land underlain by permafrost. Regardless of the causes, the impacts of anticipated changes in extreme weather hazards as a result of global climatic change, and their implications for human activity, demand the attention of geomorphologists.

Observational framework: natural hazards paradigms

Early academic research into natural hazards was characterized by an emphasis on human response to natural events. American geographer Gilbert White (1974) proposed the following research paradigm:

1 estimate the extent of human occupancy in areas subject to natural hazards;
2 determine the range of possible adjustment by social groups to those extreme events;
3 examine how people perceive the extreme events and resultant hazards;
4 examine the process of choosing damage-reducing adjustments;
5 estimate the effects of varying public policy upon that choice process.

This view emphasizes human response to specific catastrophic events, focusing on only extreme events, and implying that rational decisions are made based on cultural perceptions. Subsequent studies have increased the importance of risk assessment and the vulnerability of a population based upon the probability of an event. Although these views appear to reduce the role of the geomorphologist, the evaluation of a site with respect to specific risk lies at the very heart of hazard research.

Burton *et al.* (1978) suggest ranking the significance of potential hazards by evaluating the physical parameters of an event in terms that are obvious to geomorphologists:

1 magnitude: high to low
2 frequency: often to rare
3 duration: long to short
4 areal extent: widespread to limited
5 speed of onset: rapid to slow
6 spatial dispersion: diffuse to concentrated
7 temporal interval: regular to random.

Although qualitative, this view recognizes that events can range from intensive (such as a storm surge briefly affecting a stretch of coastline) to pervasive (such as the erosional effects of global sea-level rise).

Causal linkages are inherent in the notion of *geomorphic* hazards, where an extreme event may initiate other exceptional events of another type. Thus geomorphic hazards that are associated with landform response may be 'triggered' by climatic, hydrological, geophysical or man-induced events. Landslides may be causally linked with earthquakes, volcanic eruptions, heavy precipitation or construction activity. Pervasive linkages can result from land use change within a watershed affecting the magnitude and frequency of discharge within the stream.

Planners and developers often focus on a particular site or region, where mapping of hazard areas evaluates the potential risks for all potential hazards in such locations. Most physical scientists shun this 'hazardousness of place' concept, not wanting to venture beyond their own areas of expertise. Many social scientists characterize the actual hazard event only by its immediate physical effects, concentrating only on the societal response.

Geomorphologists recognize that the high-magnitude, low-frequency catastrophic events (large earthquakes, hurricanes) capture the attention due to the immediacy of large casualty and financial losses, but that events of moderate frequency (landslides, floods) often do as much or more collective damage. Geomorphic hazards tend to be more at the pervasive end of the hazards continuum, have slower speed of onset, longer duration, more widespread areal extent, more diffuse spatial dispersion, and more regular temporal interval. Exceptions such as slope failure exist, but in general landform change occurs over the long term at slow rates. Nevertheless geomorphologists should adopt a hazard paradigm in an effort to promote compatibility within this area of complex, and essential, interdisciplinary research.

Geomorphic hazards research

Gares *et al.* (1994) use the paradigm suggested by Burton *et al.* (1978) to discuss the role of geomorphic hazards research with respect to specific hazards. They illustrate the great variety of processes involved and suggest geomorphic evaluation in terms of the following aspects:

1 the dynamics of the physical process;
2 the prediction of the rate or occurrence;
3 the determination of the spatial and temporal characteristics;
4 an understanding of people's perception of the impact of the occurrence;
5 knowledge of how the physical aspects can be used to formulate adjustments to the event.

Geomorphologists vary widely in their definition of geomorphic hazards. Gares *et al.* (1994) limit their inclusion only to those process actions that gradually shape landforms, not agents of catastrophic change that arise from the consequences of geophysical, hydrological or atmospheric hazards, although many of these hazards result in geomorphic events. Wolman and Miller (1960) recognized that low-frequency, high-magnitude events often produce spectacular damage and geomorphic change, but events of moderate magnitude often do as much work (damage, change) cumulatively over the long run. As an encyclopedic entry, our definition will be necessarily inclusive of all agents of surficial change, pervasive to episodic.

Pervasive processes, such as soil erosion, are minimal in natural environments, but accelerate greatly with human disturbance such as forest clearing or agricultural tilling. Perception of soil erosion as a hazard involves farmers who lose crops to sheet wash and gullies, water managers and engineers who suffer siltation in reservoirs or canals, fishermen whose catch is reduced by silt and turbidity, and all who suffer from reduced crop and water yields. Soil erosion rarely results in direct loss of life, but it has a widespread distribution, high remediation costs, and long-term effects on water and food production. Despite more than seventy years of soil erosion mitigation research, countless thousands suffer malnutrition due to lost soil productivity.

Numerous geomorphic processes have causal links to volcanic eruptions. Geophysicists and volcanologists monitor eruptive precursors, prior eruptive history and distrubution of past eruptive products to assist disaster managers with warnings about the type and magnitude of imminent risk. Geomorphologists contribute to post-eruption mitigation efforts, as impacts such as pyroclastic flows and ash fall frequently result in unstable slopes, lahars clog stream channels, and overall sediment yield is greatly increased. Siltation and debris loading of streams radiating out from affected areas result in reduced channel capacity, increasing the frequency of overbank flooding, causing flow deflections and bank erosion. Applications of geomorphology include erosion control and engineering impact analysis, along with mitigation and recovery planning. Since the 1980s geomorphologists have made significant contributions following the eruptions of Mt. St Helens, USA, Mt. Pinatubo, Philippines and Nevada del Ruiz, Colombia.

Heavy rainfall can saturate soils causing rapid debris flows and mudflows. In 1938, such events in Japan were triggered by typhoon rains, resulting in the loss of more than 130,000 homes and over 2,000 lives. The magnitude of this loss prompted government attention focused on landslide control, and similar rains in 1976 affected less than 2,000 homes and cost 125 lives. Similar reductions have occurred with other catastrophic events. The horrific death toll experienced in Bangladesh due to storm surge and flooding in 1970 was reduced by thirty to fifty times during a cyclonic storm surge of similar magnitude in 1985 because a satellite-based early warning system prompted the evacuation of island and coastal dwellers. These two examples come from opposite ends of Asia's economic spectrum. Rosenfeld (1994a) points out that economically developed countries often suffer the greatest economic losses, while their lesser developed counterparts endure the highest loss of life. In developing nations, there is often a conscious decision to allocate resources toward economic development, at the risk of underfunding disaster mitigation, often with the effect of greater loss of both infrastructure and human lives.

In some instances, disaster mitigation strategies and international relief efforts may actually be partially responsible for rising losses. In many developed countries, state-backed hazard insurance programmes are designed to encourage the use of hazard zoning and the implementation of damage-resistant building codes to reduce the demand for structural control measures. However this may have actually encouraged the development of hazard-prone areas through the combined effect of lower land costs and cheap indemnity. Thus, the 'insured' transfers the risk to the 'insurer' and may dismiss the concern for loss prevention measures.

Geomorphologists have the opportunity to demonstrate the nature of geomorphic hazards (Figure 66), map landform or surface material conditions that have hazard potential, and recognize the effects of human modification of natural conditions which could result in increased hazard potential. As scientists, we are reluctant to translate this knowledge into arguments for the adoption of specific mitigation or management strategies, and thus are less than proficient at applying our information base. Most land use managers, planners, developers and government decision-makers rely on 'on the job' training to develop expertise in the interpretation of technical information for risk assessment and disaster reduction. Often this is prompted only in response to significant losses.

Automated monitoring networks and advanced computer modelling techniques are giving us new tools to test alternative hazard mitigation strategies. Geomorphologists must be willing to embrace new technologies which will permit them to exercise their specialized talents globally, interface more readily with professionals outside the Earth science community (for example social scientists and engineers), and associate more closely with monitoring networks and scientific unions to ensure that major events are anticipated by identifying their physical precursors. Natural hazards research is obviously an interdisciplinary field involving a range of physical scientists with social scientists assessing the human dimension of the problems. Given the limited observed record of hazard events in most regions, a geomorphological approach, where the areal extent, and perhaps the frequency, of events can be determined from the landscape, is essential. The geomorphological approach may also encourage the 'nature knows best' path to designing hazard mitigation strategies in balance with the dynamics of processes within the region. In the final analysis, occupance of hazard-prone areas is both physically and economically self-regulating. It is the function of science, as a servant of society, to identify those limitations and point the way toward minimizing the disastrous consequences of 'learning our lessons' in nature's way.

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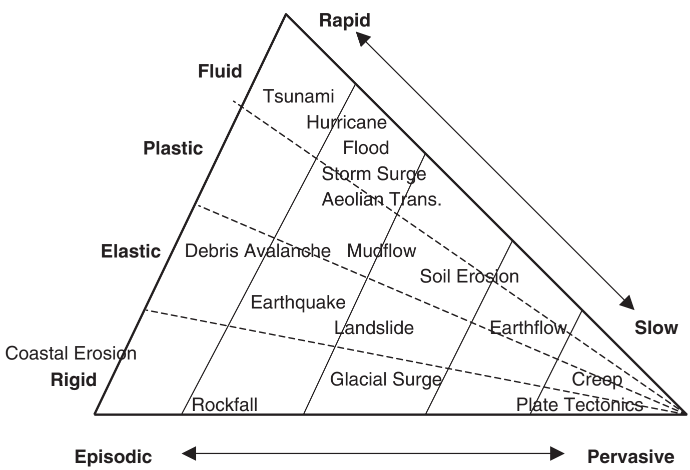

Figure 66 Geomorphic hazards include pervasive processes, which may be imperceptible to individuals, to episodic events which may have frequencies of occurrence below thresholds deemed significant by planners, but which may have significant magnitude. These processes involve the full spectrum of stress/strain modulus, and involve virtually every speciality within the discipline

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CHARLES L. ROSENFELD

GEOMORPHOLOGICAL MAPPING

Geomorphological mapping encompasses one of a group of techniques under the general category of TERRAIN EVALUATION employed to record systematically the shape (or morphology), landforms, landscape-forming processes and materials that constitute the surface of the Earth. Lee (2001) identifies three forms of geomorphological map:

1 Regional surveys of terrain conditions, either for land use planning or in baseline studies for environmental impact assessment (e.g. the 1:25,000 scale maps of Torbay, Doornkamp (1988)).
2 General assessments of resources or geohazards at scales between 1:50,000 and 1:10,000 (e.g. Bahrain Surface Materials Resources Survey, Doornkamp *et al.* (1980); ground problems in the Suez City area, Egypt, Jones (2001)).
3 Specific-purpose large-scale surveys to delineate and characterize particular landforms (e.g. the 1:500 scale investigations around the Channel Tunnel portal, Folkestone, Griffiths *et al.* (1995)).

The initial stage of geomorphological mapping involves factually recording ground shape through a process of morphological mapping. This requires the production of a map on which the land surface is subdivided into planar facets separated by gradual changes or sharp breaks in slope. On the map the changes and breaks in slope are identified as either concave or convex in nature and recorded using decorated lines, a system first established by Savigear (1965). Arrows with a numeric value in degrees indicate the slope angle and downslope direction of the planar facets. Once the morphology has been recorded a geomorphological interpretation is undertaken whereby details of the contemporary and relict landforms and geomorphological processes are added to the map. In addition, data on the nature of materials and hydrology of the area are noted. Geomorphological interpretation can allow a suite of derivative maps to be produced, e.g. resource maps and landscape genesis maps. Standard symbols to be used on all these maps are contained in Cooke and Doornkamp (1990), although, Demek and Embleton (1978) provide a more comprehensive collection of symbols that allow subtle differences in the landscape to be highlighted. However, in many situations the geomorphological maps are produced as unique products with a bespoke legend.

The techniques used to compile the data involve both field survey and, where possible, examination of remote sensing information. The main form of remote sensing analysis has traditionally been through the interpretation of vertical pairs of aerial photographs viewed stereoscopically. An initial preliminary morphological map and geomorphological interpretation is produced using aerial photographs but this should normally be subject to 'ground-truth' mapping in the field. With the advent of higher resolution satellite images this preliminary mapping stage increasingly is being carried out using data from the array of new satellite-based scanners.

A two-person team normally undertakes the field mapping. The main requirement for the production of effective geomorphological maps is an accurate base map at a suitable scale. The base map may be a standard survey map depicting man-made and natural features including ground topography, or a spatially corrected ortho-photo. The field data should be compiled directly on the base map. Spatial data and slope information can be obtained through a simple tape, compass and clinometer survey, using more sophisticated land survey techniques, use of global positioning systems, or a suitable combination of these methods. The geomorphological, materials and hydrological data are noted on maps and recorded in field notebooks where appropriate.

Whilst geomorphological mapping has been used for general landscape investigations, it has been employed most successfully by applied geomorphologists, particularly for engineering studies. Brunsden *et al.* 1975, articulated the aims of geomorphological mapping for highway engineering:

1. Identification of the general terrain characteristics of the route corridor, including suggestion of alternative routes and location of hazards.
2. Defining the 'situation' of the route corridor, for example identifying influences from beyond the boundary of the corridor.
3. Provision of a synopsis of geomorphological development of the site, including location of materials for use in construction and location of processes affecting safety during and after construction.
4. Definition of specific hazards, e.g. landsliding, flooding, etc.
5. Description of drainage characteristics, location and pattern of surface and subsurface drainage, nature of drainage measures required.
6. Slope classification, according to steepness, genesis and stability.
7. Characterization of nature and extent of weathering, also susceptibility to mining subsidence and erosion.
8. Definition of geomorphological units, to act as a framework for a borehole sampling plan and to extend the derived data away from the sample points.

Although these aims were developed specifically for highway projects they represent an appropriate checklist for all geomorphological mapping programmes undertaken for civil engineering projects.

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GEOMORPHOLOGY

Definition and scope

Geomorphology is the area of study leading to an understanding of and appreciation for landforms and landscapes, including those on continents and islands, those beneath oceans, lakes, rivers, glaciers and other water bodies, as well as those on the terrestrial planets and moons of our Solar System. Contemporary geomorphologic investigations are most commonly conducted within a scientific framework (see Rhoads and Thorn 1996) although academic, applied or engineering interests may motivate them. A broad range of alternative research methodologies have been employed by geomorphologists, and past attempts to impose a systematic structure on the discipline have yielded stifling tendencies and overt resistance. Geomorphologists frequently profess to innate aesthetic appreciation for the complex diversity of Earth-surface forms, and, in this regard, a fitting definition of geomorphology is simply 'the science of scenery' (Fairbridge 1968).

Past and present concerns have focused on the description and classification of landforms (including their geometric shape, topologic attributes and internal structure), on the dynamical processes characterizing their evolution and existence, and on their relationship to and association with other forms and processes (geomorphic, hydro-climatic, tectonic, biotic, anthropogenic, extraterrestrial, or otherwise). Geomorphology is an empirical science that attempts to formulate answers to the following fundamental questions. What makes one landform distinct from another? How are different landforms associated? How did a particular landform or complex landscape evolve? How might it evolve in the future? What are the ramifications for humans and human society?

Modern geomorphology is currently subdivided and practised along the lines of specialized domains. *Fluvial geomorphology*, for example, is concerned with flowing water (primarily in the form of rivers, streams and channels) and the work it accomplishes during its journey through the terrestrial phase of the hydrologic cycle. A very broad spectrum of interests are subsumed within fluvial geomorphology, ranging from the influence of turbulence on the entrainment, transport and deposition of sediment particles at the finest scale, to the mechanics of MEANDERING, POINT BAR formation and FLOODPLAIN development at middle scales, to the nature and character of DRAINAGE BASIN evolution at the coarsest scales. Within the other substantive areas of geomorphology are: *hillslope geomorphologists*, who boast expertise on the geotechnical properties of soil and rock, the mechanics of LANDSLIDEs, and the movement of water within the ground; *tectonic geomorphologists*, who study neotectonic (see NEOTECTONICS) stress fields, continental-scale sedimentary basins and active/passive margin landscapes; *glacial and periglacial geomorphologists*, who are interested in alpine and continental glaciers, PERMAFROST and other cold-climate forms or processes that involve ice, snow and frost; *karst geomorphologists*, who deal with soluble rocks (e.g. limestone) and chemical processes of DISSOLUTION that lead to landforms such as gorges, caverns and underground streams; *coastal geomorphologists*, who study nearshore, lacustrine and marine systems where oscillatory, rather than unidirectional, flow processes dominate; and *aeolian geomorphologists*, who study the transport of sand and dust by wind, mostly in desert or semi-arid environments, but also along beaches, over agricultural fields and on the moon and Mars. Other subspecialities include: soils geomorphology, biogeomorphology (zoogeomorphology), climatic geomorphology, tropical geomorphology, desert geomorphology, mountain geomorphology, extraterrestrial (planetary) geomorphology, remote-sensing geomorphology, experimental geomorphology, environmental geomorphology, forest geomorphology, applied geomorphology, engineering geomorphology and anthropogeomorphology.

Major themes and concepts

Landforms are dynamic entities that evolve through time as a consequence of characteristic suites of processes acting upon Earth-surface materials. Geomorphologists are concerned with documenting and unravelling the mysteries of this process-form interaction. Relevant knowledge includes not only the manner and direction of landform evolution (progressive or cyclic, slow or rapid), but also the processes that dominate or direct the evolution (type, intensity) as well as the mutual adjustments and feedbacks that occur between the forms and processes as energy and matter are cycled through the landscape. To better understand these complex interrelationships, geomorphologists have proposed various conceptual themes or templates to aid in organizing their thinking. Among these are:

1. *Endogenic–exogenic forces* Geomorphic systems are governed by dynamic controls that may be internally produced (endogenic) or externally imposed (exogenic) upon the system. Tectonic, volcanic and isostatic activities are manifestations of endogenic forces within Earth, whereas rainfall and meteorite showers are exogenic forces. The spatial and temporal scales of the geomorphic system influence the types of endogenic–exogenic forces that are relevant. The control–volume, force–balance approach in fluid mechanics is an analogue to this concept.
2. *Destructive–constructive action* Some geomorphic processes create landforms (e.g. volcanic cones, meteorite impact craters, termite mounds) whereas other processes (e.g. chemical weathering, rainwash, human activity) destroy landforms or cause widespread denudation. More typically, most geomorphic

processes both create and destroy landforms simultaneously. For example, flowing water in a river will erode the outer bank of a meander bend while depositing sediment on the inner bank in the form of a point bar. Similarly, a glacier can erode, excavate and sculpt the surface upon which it moves while also depositing sediment in the form of eskers and moraines.

3 *Erosional–depositional forms* Some landforms are sculpted by erosion of pre-existing materials (e.g. bedrock canyons, roches moutonnées) whereas others are built via deposition of new material on existing substrate (e.g. deltas, lava flows). Yet others are hybrid features formed through both erosion and deposition locally (e.g. impact craters) or maintained by an intricate balance between erosion and deposition at different positions on the same form (e.g. migrating sand dunes).

4 *Stress–strength relationships* Most geomorphic processes induce landscape change by stressing the system, as with flowing fluids, chemical reactions, tectonic motion or the prolonged action of gravity. The materials upon which these processes act have the ability to resist change because of inherent properties that provide strength (e.g. mineralogy, cohesion, structure, relative placement). Geomorphologists have generally devoted more effort to measuring processes than to investigating how material strength is a complementary and counterbalancing factor (exceptions are many, and they include the efforts of the Japanese school to elucidate systematically the nature of rock control on geomorphic evolution, of the many coastal geomorphologists interested in rocky coasts, of several geomorphological geologists concerned with relict landforms and ancient landscapes, and various engineering geomorphologists who study slope failures).

5 *Polygenesis and inheritance* Landscapes consist of landform assemblages that are rarely simple. Complex suites of polygenetic forms may coexist in the same location if, for example, multiple processes are active contemporaneously or when particularly resistant relict forms are inherited from prior eras. The latter are progressively modified by contemporary processes that also create new forms to produce palimpsest landscapes.

The integrated sum of exogenic–endogenic forces and destructive–constructive actions working in concert to create erosional–depositional forms according to dominant stress–strength relationships will dictate whether landscape RELIEF will be enhanced or reduced over a given time interval. At one end of the spectrum are the steep mountain and valley systems of the globe (e.g. Himalayas), and at the other are the extensive abyssal plains of the deep ocean as well as the PENEPLAIN of William Morris Davis. Geomorphologists have identified several additional themes and concepts that serve to strengthen the theoretical foundation of their science. These include scale, causality, equilibrium, equifinality, thresholds, magnitude–frequency, landscape memory and relaxation. Readers should consult the references and other entries in this encyclopedia for detailed discussions.

Early historical development

The subject matter of geomorphology has occupied human thinking for thousands of years, and early writings on landforms can be traced to the time of the ancient Greek, Roman, Arab and Chinese philosophers. Aristotle (384–322 BC) and Strabo (54 BC–AD 25), for example, had keen insights into the origin of springs, the work of rivers and the importance of earthquakes and volcanoes. Nevertheless, the history of geomorphology (see Chorley *et al.* 1964; Tinkler 1985, 1989) is typically traced back only as far as the European Renaissance because few written documents about geomorphic knowledge remain from the period prior to the sixteenth century. During the Renaissance, most studies of Earth were conducted from a naturalist, philosophical perspective because specialized academic disciplines had not evolved and scientific methods were not widely known. Leonardo da Vinci, Bernard Palissy, Nathanael Carpenter, Bernhard Varenius, Thomas Burnet and Nicolaus Steno are among the key figures from this period, and unwittingly they began to lay the foundation for the science of geomorphology. Unfortunately, this was also a time when the Church exerted powerful control over academic thinking, and the predominant objective of learned men was to reconcile their day-to-day observations of natural processes with strict religious orthodoxy and bibliolatry. The biblical scholar, James Ussher, Archbishop of Armagh, decreed that Earth was created on Sunday, 23 October 4004 BC and that the Flood of the Old Testament began in 2349 BC, and in so doing, he may well have imposed the most stifling proclamation on the developmental history of the Earth sciences. All evolutionary processes, by definition, had now to be contemplated within the constraints of a 6,000-year Earth history, and to think otherwise was heresy. Unsurprisingly, the dominant interpretation of Earth-surface processes invariably involved catastrophes, cataclysms and disasters such as global deluges and seismic convulsions.

The period following the Renaissance and into the early nineteenth century was one of scepticism, controversy and debate. It was also one that witnessed several changes that bear directly on the development of geomorphology as an academic discipline. The first was the evolution of specialized areas of study such as biology, physics, astronomy, mathematics, hydraulics and geology, and this set the stage for various subdisciplines, such as petrology, mineralogy, paleontology, stratigraphy and geomorphology, to be spawned. Second was a slow transformation in academic discourse away from the unassailable validity of belief systems and authoritarianism toward a standard of proof based on empiricism and observable evidence. Third was the development of increasingly sophisticated instrumentation and measurement technologies and protocols. Fourth was the enhanced mobility of people and information, thereby facilitating greater exposure to new and interesting environments and ideas. And fifth was the gradual acceptance of gradualism (see UNIFORMITARIANISM) in favour of CATASTROPHISM. Two dominant factions emerged during this period. The Neptunists (or Wernerians) followed the ideas of a German mineralogist, Abraham Gottleb Werner, who contended that rocks on Earth originated from mechanical and chemical processes in a universal ocean. The Plutonists (or Vulcanists) stressed the importance of intrusive and extrusive volcanic processes in rock formation. Key figures during this period include members of the 'French School', such as Jean Étienne Guettard, Nicolas Desmarest and Jean-Baptiste Lamarck, as well as the Swiss geologist, Horace Benedict de Saussure.

James Hutton, credited by some as the founder of modern geomorphology, was a Plutonist who argued vehemently for the importance of gradual subaerial denudation across millennia. His uniformitarian ideas, expressed in well-known phrases such as 'the present is the key to the past' and 'no vestige of a beginning, no prospect of an end', were revolutionary because they shifted the focus of attention away from catastrophic events of 'creation' toward continuous, everyday agents of erosion. Unfortunately, Hutton's teachings were not warmly received by the conservative cognoscenti of that time. After Hutton's death, his friend and colleague, John Playfair, published a book that explained and expanded Hutton's writings, and by the beginning of the nineteenth century, a slow conversion to gradualism was taking hold. Cyclic and timeless theories of landscape evolution were coming into vogue. Three schools of thought regarding landform evolution emerged. DILUVIALISM represented a transformed extension of the catastrophist lineage, and diluvialists such as Reverend William Buckland and Reverend Adam Sedgwick believed that huge floods carved many surface features. Structuralists, such as Henry Thomas de la Beche and John Phillips contended that structural controls were paramount to understanding landscape genesis (see STRUCTURAL LANDFORM), while also acknowledging that both catastrophic and gradual processes could yield substantial erosion. Fluvialists, in contrast, argued for the dominance of rivers and streams in wearing away the landscape through slow, but continuous action.

A chief proponent of fluvialism and UNIFORMITARIANISM was Sir Charles Lyell, whose *Principles of Geology* went into twelve editions after original publication in 1830. Lyell based his arguments on careful observations and measurements, and effectively attacked the notions of theological reconciliation, catastrophism and diluvialism. His writings on uniformitarianism incorporated four distinct notions: (1) uniformity of law (the laws of nature are immutable); (2) uniformity of process (processes operative today were also operative in the past, and exotic causes need only be invoked unusually); (3) uniformity of rate (gradualism); and (4) uniformity of state (change is endlessly cyclical and directionless). The publication of Lyell's *Principles* engendered considerable debate, and the period through to the middle of the nineteenth century witnessed both conflict and compromise regarding the importance of fluvial action, pluvial denudation, marine dissection, iceberg drift and glaciation (see GLACIAL THEORY) as agents of erosion. Indeed, even Lyell began to expound the virtues of marine dissection above fluvial degradation. In part, this was due to the existence of various unexplainable observations such as major unconformities in the stratigraphic record and huge ERRATICs in unexpected places. The powerful action of the sea presented an expedient solution because submarine processes could not be observed or measured directly, and theorization could proceed unbridled. Nevertheless, most of these seemingly contradictory theories incorporated at least some common elements and themes and, invariably, they were cast within a framework of uniformitarianism rather than catastrophism.

By the mid-1870s, some consensus was beginning to emerge about the multifaceted and complex nature of landscape evolution. The marine planation theory of Sir Andrew Crombie Ramsay, for example, proposed that the action of waves and currents in the ocean was not to dissect the sea bottom, but to level off bathymetric protuberances thereby producing marine plains. Upon emergence through tectonic activity, subaerial forces become active and fluvial erosion proceeds to carve out valleys and denude landscapes. Support for this theory came from the many accordant summit heights in the highlands of Wales and England, as well as from the marine abrasion studies of Baron Ferdinand von Richthofen in China. Concurrently, the glacial theories (see GLACIAL THEORY) of Ignace Venetz, Jean de Charpentier and Louis Agassiz were receiving widespread acceptance decades after their introduction, albeit with climatic and glaciofluvial amendments. This was a significant development in geomorphology because environmental dynamism (see DYNAMIC GEOMORPHOLOGY) was implicit to these theories. Gradualist and neo-catastrophist perspectives could both be accommodated under this new framework because uniformity of process (the nature of past and present processes are the same) did not necessarily imply that the intensities and rates of process action could not vary.

At the conclusion of the nineteenth century, geomorphology was poised to begin its emergence as a modern scientific discipline. The word 'geomorphology' had already been coined in the mid-1800s (Tinkler 1985: 4), and several textbooks on exclusively geomorphic matters had been written. As an area of academic study, geomorphology was experiencing legitimate interest under the guise of 'physiography' or 'physiographical geology'. Centres of expertise were arising in many different countries within and outside Europe, all with subtly different identities and separate agendas. British geomorphologists, for example, spent considerable effort on compiling complex denudation chronologies linked to marine processes and periods of tectonic stability/instability and sea-level fluctuation. German geomorphologists (e.g. Hettner, A. Penck, Walther) became interested in the influence of climate as a consequence of conducting research in the Alps as well as in the subhumid tropics. The North American school, in contrast, was dominated by fluvialism buoyed by indisputable evidence derived from the great explorations of the largely unvegetated, semi-arid West. John Wesley Powell's trips into the Grand Canyon and his reports on the Colorado Plateau and Uinta Mountains provided powerful testimony to the efficacy of rivers to erode landscapes. Grove Karl Gilbert's studies on the mechanics of fluvial erosion, sediment transport and turbulence are exemplars of the elegant application of the scientific method. He also investigated the origin of pediments and lateral planation, and in recognition of his many contributions, Gilbert is often identified as the first truly process-oriented American geomorphologist. Indeed, it is largely due to the efforts of Powell, Gilbert, Dana and Dutton and various other United States Geological Survey employees that the North America school became the dominant force in the development of geomorphology at the turn of the century.

Twentieth-century developments

Geomorphology in the twentieth century experienced rapid evolution and growth, and six overlapping phases of development can be identified. These are little more than crude caricatures, and the reader is referred to Chorley *et al*. (1973) and Beckinsale and Chorley (1991) for detailed discussions of the key figures and their substantive contributions. The *historical* phase, roughly from 1890–1930, was dominated by William Morris Davis and his many disciples. Davis's deductively derived model, 'The Geographical Cycle', envisioned serial evolution of landscapes beginning with rapid tectonic uplift followed by progressive denudation in characteristically distinct stages of 'youth', 'maturity' and 'old age'. It served as the genetic template upon which reconstructive narratives of landscape evolution were hung, with relatively little concern for the mechanical and chemical processes responsible for erosion and deposition. Nevertheless, these denudation chronologies spawned keen interest in tectonic geomorphology as well as an appreciation for the importance of unravelling the historical sequence of steps that ultimately manifest themselves as a contemporary landscape.

The *regionalist* phase (1920–1950) was characterized by detailed and thorough investigations of regional landscapes, both in the conventional mid-latitudes of North America and Europe as well as in globally remote areas (e.g. tropics, deserts, high latitudes). Increasingly, these regionally based studies yielded data about landforms and landform assemblages that could not be easily explained within the framework of Davis's geographical cycle, especially his contentions about the 'normalcy' of the humid, mid-latitudes. Although Davis found support within Britain and France, many European schools remained unconvinced by Davis's teachings. Walther Penck, for example, proposed an alternative model of landscape evolution that highlighted the importance of the relative rates of uplift and denudation in controlling landform geometry. Another German geomorphologist, J. Büdel, stressed the dominance of climatic controls and proposed the concept of etchplanation (see ETCHING, ETCHPLAIN AND ETCHPLANATION) and MORPHOGENETIC REGIONS. Climatic geomorphology was also practised by Louis Peltier in North America and it was later championed by J. Tricart and A. Caillieux in France. In this way, Davis's unifying ideas gradually fell into disfavour, and geomorphology became an empirically driven scientific confederacy of polyglot regionalist schools.

The *quantitative* phase (1940–1970) reflected a broader trend within many of the Earth sciences toward enhanced use of sophisticated technologies (often derived from the war effort) to measure, describe and analyse the surface features of Earth. R.E. Horton's publications on stream networks and drainage basin processes are classically identified as the precursor to this quantitative movement, but the foundational works of Bagnold, Gilbert, Hjulstrom, Leighly, Rubey and Shields, among many others, are rightfully acknowledged. These early 'quantifiers' were concerned to understand landforms and geomorphic processes in deterministic or probabilistic, but testable, ways rather than on the basis of deductively derived heuristic models that ultimately yielded little predictive power. Logical positivism was the dominant philosophy and reductionism was the overriding methodological approach. As a consequence, geomorphology became increasingly fragmented and specialized, with fewer and fewer connections between the sub-specializations as well as pronounced distancing from its mother disciplines of geography and geology. Fortunately, connections to other allied disciplines such as fluid mechanics, engineering hydrology, statistics, thermodynamics, meteorology, pedology and agricultural physics were being cultivated, and these provided a theoretical and conceptual richness upon which geomorphologists could draw, if so inclined.

The *systems* phase (1960–1980) in the development of geomorphology was inaugurated by the introduction of general systems theory into the conceptual toolkit of geomorphology by Richard J. Chorley, which was a logical outgrowth of the quantitative phase. The quantification of prior decades was basically of two genres: (a) statistical 'black-box' description (e.g. Horton's Law of Stream Numbers); and (b) detailed measurement and interpretation of dynamical processes (e.g. Strahler 1952). The former proved unrewarding in terms of providing insight into geomorphic behaviour, whereas the latter were typically conducted at a scale that was too small to be relevant to landscape evolution. The systems approach alleviated the 'black-box' quandary by describing geomorphic behaviour in terms of energy and mass flows, equilibrium tendencies, relaxation times and thresholds (see THRESHOLD, GEOMORPHIC) of response. A large number of concepts, such as ALLOMETRY, entropy and ergodicity (see ERGODIC HYPOTHESIS), were borrowed from other disciplines as theoretical templates. These were applied to a broad range of geomorphic systems with varying degrees of success, but the large number of journal articles and textbooks containing box-and-arrow plots attests to the popularity of this approach during the systems phase. Unfortunately, there was an irresistible tendency to equate system behaviour with geomorphic process, much to the detriment of dynamical process investigations.

Since about the 1980s, geomorphology has entered a phase of increasing *reconciliation* and *unification* that signals its arrival as a mature modern science. Introspective debates about catastrophic versus uniformitarian ideas, quantitative-deterministic/stochastic versus qualitative- historical methodologies, and geographical versus geological disciplinary roots are taking place not for purposes of disciplinary leadership or hegemonic posturing, but rather in consequence of the pragmatic need for geomorphology to assert an identity distinct from that of other Earth sciences (e.g. geology, geography, sedimentology, stratigraphy, paleontology) as well as to understand the complex spectrum of conceptual ideas upon which geomorphology is founded (e.g. Rhoads and Thorn 1996). Many extremist ideas of prior eras have been reintroduced into the literature as softened compromises (e.g. NEOCATASTROPHISM, neo-historicism, neo-regionalism) to provide balance to the uniformitarian-style fluvialism that dominated the quantitative and system phases. Invariably, these conceptual ideas were discussed in the context of factual evidence and with a view toward generating insight into unusual geomorphic features or terrain that belie conventional explanation (e.g. Baker 1981). The modern-day geomorphologist has a deep appreciation for the importance of slowly acting processes in concert with large-magnitude, low-frequency events in leaving imprints on the landscape, for the utility of detailed process-mechanical studies as well as historical reconstructions of landform assemblages in unravelling the complexities of the present-day surface, for the interconnectivity between the various sub-specializations of geomorphology and allied Earth and engineering sciences, and for the complementarities among twenty-first century technological capacities when combined with a field geomorphologist's keen sense of the lie of the land.

Future directions

Geomorphology in the twenty-first century will continue to mature as a science and assert its importance among the Earth-science disciplines. The issue of scale will remain a dominant topic of investigation and discourse, and it will be richly informed by expanding concerns about tectonic and structural controls on geomorphic systems over long time periods (i.e. megageomorphology), the evolution of lunar and Martian surfaces (i.e. planetary geomorphology), the intricate linkages between geomorphic and biogeochemical systems, and the hierarchically nested versus scale-invariant nature of geomorphic systems. The term 'neogeomorphology' was recently coined (Haff 2002) to suggest that a new or modern form of geomorphology may be evolving – one which, of necessity, takes into account the sobering fact that humans now displace more soil and rock per year than rivers, glaciers and wind combined (Hooke 2000). The pace of anthropogenically driven landscape alteration, whether direct or indirect (e.g. via global warming), is likely only to increase in the future. And, because there are no analogues for such pronounced surface modification in the stratigraphic record, the relevance and utility of geomorphology (with its traditional focus on process–form interaction) to the planning and environmental management communities seems assured. Geomorphologists already play central roles in mandated environmental impact assessments involving construction, mining and forestry, and increasingly their expertise (in conjunction with biologists and botanists) is utilized in landscape reclamation, rehabilitation and restoration efforts involving streams, wetlands and coastal dunes.

In the quest for a deeper understanding of the past (retrodictive) and future (predictive) evolution of Earth's surface, geomorphologists are becoming increasingly reliant on sophisticated technologies. These include: new dating methods (e.g. cosmogenic radionuclides, optical- and thermo-luminescence, rock varnish, lichenometry) that yield the relative ages of landform elements and thereby unfold the historical sequence of events that produced the landscape; novel remote-sensing techniques (e.g. interferometric synthetic-aperature radar, lidar, ground-penetrating radar, time-domain reflectrometry) to measure and monitor a broad range of surface and subsurface attributes; advanced computational methods involving more powerful hardware, more efficient software codes, and more easily integrated and interoperable data platforms (e.g. Geographical Information Systems, Digital Elevation Models); and enhanced satellite coverage to provide synoptic information about inaccessible regions and across large distances with ever-increasing accuracy regarding absolute location and relative movement via the Global Positioning Systems. In addition, the means to communicate information and ideas virtually instantaneously to the entire community of geomorphologists has been greatly facilitated by the World Wide Web and by various national and international organizations such as the International Association of Geomorphologists (IAG) that maintain electronic bulletin boards and membership/address lists. For the first time in its long developmental history, geomorphology has the potential to become a truly global enterprise in terms of both coverage and participation.

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BERNARD O. BAUER

GEOMORPHOMETRY

Dealing with quantitative analysis of the land surface, geomorphometry is a central theme in both theoretical and applied geomorphology (Pike and Dikau 1995). It is also diverse, dealing both with landforms and with the land surface as a one-sided rough surface, with vertical position a unique function of horizontal location. It could also be referred to as the combination of 'landform morphometry' and 'land surface morphometry'. It does not include surveying, photogrammetry and profiling, which provide the raw data for geomorphometry, but some knowledge of these is essential in considering error margins (Richards 1990: 36–41). Morphometry itself is a broader field, important not only in various aspects of Earth science, but also in engineering, biology and medicine. Each field of application has things to teach the others (Pike 2000), so long as the specifics of the original application are remembered. Geomorphometry inspired the idea of statistical FRACTALS, following difficulties in specifying 'how long is a coastline?'

Where individual landforms can be defined and distinguished from their surroundings, a series of MORPHOMETRIC PROPERTIES can be measured to provide a multivariate characterization of the landform. Such analysis is labelled 'specific geomorphometry'. This is distinct from 'general geomorphometry' of the land surface as in spectral or fractal analysis, or the study of surface derivatives and their interrelations (Evans 1980). General geomorphometry was extremely difficult before the introduction of computers: today it may involve the processing of very large DIGITAL ELEVATION MODELS (DEMs), and has many applications in digital terrain modelling (Pike 2000). Specific geomorphometry has a much longer history, starting with measurements of lunar craters and of coastal sinuosity in the nineteenth century.

The two aspects of geomorphology are not completely distinct, first because some specific landforms such as slopes or hillslopes and drainage networks are so widespread on Earth that their specific geomorphometry acquires a general importance: and second, because some techniques of general geomorphometry can be applied to specific landforms (Evans 1987). This permits analysis of variation within a landform (distributional analysis), not just generalization of its overall characteristics, and is more useful in the context of modelling.

General geomorphometry; surface derivatives

General geomorphometry starts with the altitude (elevation) of the surface – its height above sea level. This has major effects on climate and thus on surface processes. The frequency distribution of altitude (hypsometry) tells us quite a lot about the land surface. In the pre-computer era, this was summarized by its range (relief) and by the hypsometric integral – the relation of mean altitude above minimum, to this range. Relief varies with the size of area considered, and reaches several km (ridge to valley) in high mountain areas: the total range for the Earth is 8,852 + 11,033 m, i.e. 19.9 km. Hypsometric integral is around 0.5 for topography with sharp ridges and valleys, approaching but not reaching 1.0 for a plateau with few deep valleys, or 0.0 for a lowland with a few high hills. Evans (1972) suggested that instead of ranges and extremes, the use of standard statistical concepts – standard deviation and skewness – was both more economic and provided more stable statistics, influenced by the whole body of the distribution rather than by the extremes.

Ohmori (1993) found that hypsometric curves of mountainous areas such as Japan tend to be S-shaped or concave, giving integrals between 0.15 and 0.50. They can be simulated from empirically based relations between uplift, altitude, altitude dispersion and denudation rates. Hypsometric curves vary considerably with extent of area considered, and whether headwaters, large erosional basins or areas including depositional plains are analysed. Fuller understanding of landscape development is obtained by considering dimensional indices (mean and standard deviation of altitude) and not just dimensionless indices.

Gradient (slope angle) is the second local value of a surface that is very important in geomorphology and hydrology. It provides the stress to generate mass movements, and gives energy to surface flows. Engineers prefer percentages, i.e. 100 × (tangent of angle), but geomorphologists prefer to measure gradient in degrees. Mean gradient is of primary interest, but standard deviation and skewness of the point-by-point distribution of gradients tell us much about the regional topography. Fluvially dissected hill areas with slopes near some threshold tend to have low standard deviations of gradient, while glaciated mountains with cliffs, valley floors, terraced areas and often plateau remnants and benches have high standard deviations.

Lowland areas tend to have a few steep slopes and many gentle ones; their gradients are positively skewed. In mountain areas, the opposite applies as slopes approach gradients limited by slope stability and ROCK MASS STRENGTH. For example, in the Japanese mountains, on igneous and sedimentary rocks, the mode of gradients becomes sharper as altitude increases. The mode is at 33 to 37 degrees in all three ranges of the Japan Alps (central Honshu) above 1,000 m, up to 2,800 m (Figure 67; Katsube and Oguchi 1999). Mean gradients increase with altitude, to maxima of 32 to 35 degrees above 2,000 m. In the high relief of the north-west Himalaya, on crystalline rocks, gradients range from 0 to 60 degrees, with modes of 33–37 and means of 30–34 degrees (Burbank *et al.* 1996) despite varying uplift and denudation rates. These distributions may reflect a dynamic equilibrium with landsliding removing fractured rock and river gradients increasing to transport this. The similarity to Japan may be deceptive in that averaging over several hundred metres reduces the Himalayan measurements.

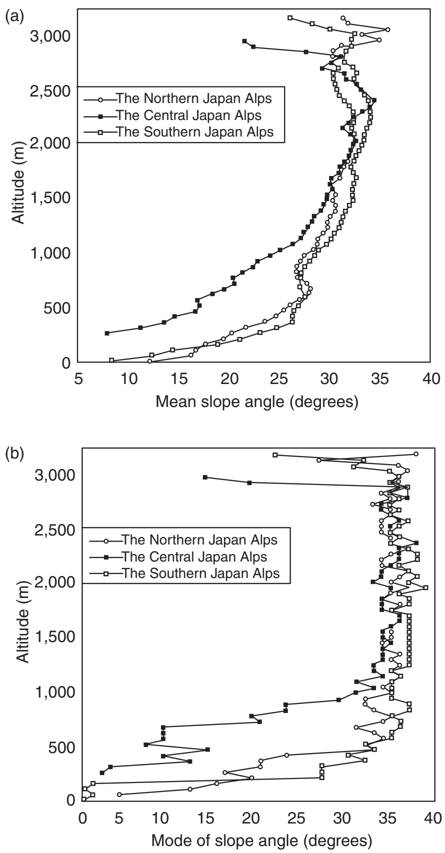

Figure 67 Altitudinal change in (a) mean and (b) modal slope angle (gradient) in three divisions of the Japan Alps

Source: Reproduced from Katsube and Oguchi (1999) with permission from the Association of Japanese Geographers

Gradient is defined as the rate of change of altitude in the direction where that rate is maximized (this is 'true gradient' as opposed to 'apparent gradient' in an arbitrary direction along a profile). This immediately implies a related variable, aspect – the direction or azimuth of this true gradient. Gradient and aspect form a closely related pair, the vector defining surface slope. Aspect, modulated by gradient, has considerable influence on slope climate (mesoclimate), especially solar radiation and exposure to wind. Although slope vectors can be analysed as poles to planes tangential to the surface, that approach ignores the different ways in which gradient and aspect affect surface processes. Aspect is a circular variable ($0 \equiv 360$ degrees) and it is easy to produce misleading results by applying ordinary linear statistics; it should be summarized by vector, directional or circular statistics, and related to other variables through its sine and cosine, in Fourier Series Analysis.

Rates of change of gradient and aspect in turn define components of curvature, the second derivative of the surface. Evans (1980) defined profile convexity as rate of change of gradient (with negative values representing concavity) and followed earlier geomorphologists such as Young (1972) in expressing this in degrees per 100 m. Plan (contour) convexity is thus the rate of change of aspect. Both variables are encountered in models of surface runoff (see RUNOFF GENERATION). Tangential curvature and other definitions have also been used: mathematically, curvature has three independent components. Of the many ways in which surface curvature can be defined, here we are concerned with those related to the gravity field, which is of central importance in geomorphology. As standard deviation of plan convexity measures the intricacy of contours, it expresses DRAINAGE DENSITY: this relationship requires further investigation.

Surface roughness or ruggedness is a broad concept, covering mean and variability of gradient, and variability of curvature in both profile and plan.

Altitude and its first and second derivatives provide local variables, conceptually related to points although in practice small neighbourhoods are used in their measurement. Context or position on the surface is also important, especially in relation to runoff. Contributing area upslope (per unit width of contour) controls the potential runoff that can be generated, and is used in models and applications (Lane *et al.* 1998; Wilson and Gallant 2000).

Other point aspects of the surface are those which are topologically special, in terms of position: these are summits, saddles and pits, and can be further subdivided in relation to the pattern of higher and lower land in the vicinity. Ridges, valleys and breaks of slope provide linear features at which slope either reverses or changes abruptly. Topological and other linear aspects of the surface are considered under DRAINAGE BASINS. At special points or lines, some derivatives may be indeterminate as gradient passes through zero: notably, aspect and plan convexity/curvature. Plains are areas of zero gradient and again aspect is indeterminate: their extent, however, varies with the vertical resolution of the data (e.g. altitude in metre units, or in tenth-metres, etc.).

General geomorphometry gives an appearance of objectivity, but it involves choice of data source, of horizontal and vertical resolution, and of algorithms for interpolation, smoothing and derivative calculation. Most important of all is definition of areas for which statistical summaries are to be provided. Map sheets or tiles of data are easiest to use, but natural regions may be more appropriate. Islands are the most obvious, but there are two complementary ways in which the land surface may be subdivided into exhaustive, non-overlapping areas. These are drainage basins, and 'mountains' bounded by valleys and low passes.

Spatial series, and complexity

Altitude is a positively autocorrelated variable, that is it defines a generally smooth surface. The rate of decline of autocorrelation with separation is thus an important property, and forms a basis for spectral analysis (Pike and Rozema 1975). This relates to the use of geostatistics and FRACTALS. They provide highly simplified models poorly suited to subaerial topography.

The land surface is complex and its morphometry varies from area to area with rock type and structure, climatic variables and their history, and tectonic history. Attempts to compress its variability into two or three statistical dimensions meet with difficulties. Multivariate studies show that at least nine dimensions (Table 21) are largely independent of each other.

Specific geomorphometry

Taking measurements of landforms requires their precise definition (what is/is not . . . ?) and

Table 21 Statistical dimensions of (a) the Wessex land surface, England for 53 areas, and (b) the French land surface, for 72 areas

Property	Statistical descriptor (key variable)	Dimension
	(a) Wessex (Evans)	*(b) France (Depraetere)*
Gradient	Mean gradient	1. Relief
Massiveness	Skewness of altitude	4. Skewness of altitude (and 5.)
Level	Mean altitude	* in 1.
Profile convexity	Skewness of profile convexity	2. Convexity, cols and depressions
Orientation	Weighted vector strength (modulo 180°)	–
Plan convexity	Standard deviation of plan convexity	* in 1.
Altitude-convexity	Correlation of altitude with profile convexity	3. Convexity, crests and slopes
(Profile) variability	Standard deviation of gradient	*in 1.
Directedness	Weighted vector strength (modulo 360°)	–
		5. Skewness of gradient

Notes: All areas 10 × 10 km and analysed from 50 m grids. Numbers in (b) give the rank order of factors
Source: From Evans, in Hergarten and Neugebauer 1999

complete delimitation by a closed outline; here it may be difficult to achieve consistency between researchers. Although specific geomorphometry has been more subjective than general geomorphometry, work has now started on recognition and delimitation of landforms on DEMs by objective criteria. In specific geomorphometry, variables are defined specifically for each landform type. Commonly these include size (length, width, height, area, volume), gradient and shape (often ratios between size variables). The number of possible indices is increased where landforms are subdivided into several parts, e.g. volcano (or impact feature) outer slopes, craters and central peaks. Position (often a surrogate for climate) and geology are sometimes included as potential controlling variables. The more definable landforms include those listed at the end. Each of these has a body of geomorphometric literature. Landform shape and spatial pattern (position relative to others of the same type) were discussed by Jarvis and Clifford (in Richards 1990).

Evans (1987) distinguished eight stages in a specific morphometric study: conceptualization; definition; delimitation; measurement; calculation of indices; analysis of statistical frequency distributions; mapping and spatial analysis; interrelation of attributes; and assessing meaning. Analysis can be in terms of distributions of point variables (altitude, slope and curvature, as discussed above), sets of indices or measurements characterizing each landform (the most common approach), or fitting equations to the whole form or a selected part, outline or profile. In that residuals from such equations usually exhibit spatial pattern, simple equations are rarely good models for landforms.

General concepts in specific geomorphometry include symmetry (radial or axial), scale and the relation of size to shape. The latter can be isometric (shape does not vary with size, expected values of all ratios remain the same) or allometric (see ALLOMETRY; shape changes systematically, often as a power function of size). Scale is fundamental to both general and specific geomorphometry (Dietrich and Montgomery 1998; Wood 1996). Most landforms are defined with specific scales in mind, usually with something like a tenfold range in linear size. Within a particular landform type, different attributes (size, gradient) scale smoothly with each other. Sometimes scale breaks are discovered, and these reveal process thresholds – as for the central features of impact craters (Figure 68; Pike 1980).

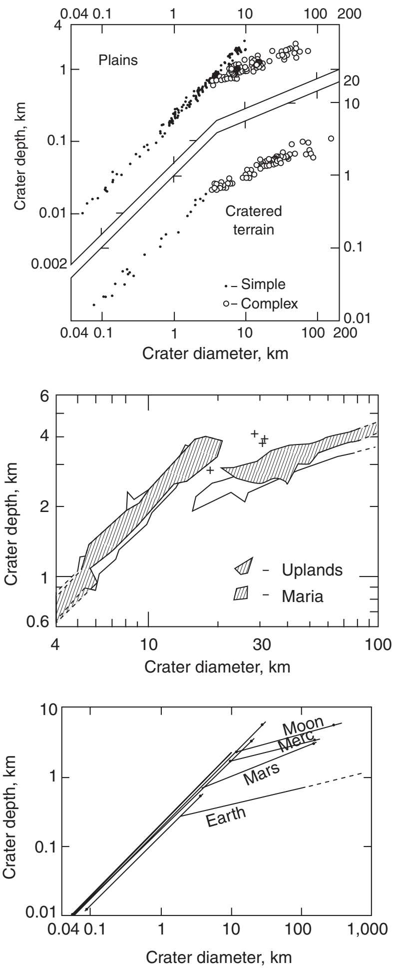

Figure 68 Breaks in the crater depth:diameter scaling relation, illustrating the morphologic transition from simple to complex craters (a) 230 craters on Mars, showing larger simple craters on plains than on 'cratered terrain'; (b) based on 203 mare craters and 136 upland craters on the moon. Simple craters follow a similar relation for maria and for uplands (as for the two divisions of Mars), but complex craters average 12 per cent deeper in uplands; (c) summary of the relationships on three planets and the moon. The transition size increases as gravity decreases

Source: Reproduced from Pike (1980: figures 6, 9 and 2) with permission

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SEE ALSO: hillslope, form; hillslope, process; slope, evolution; and the landforms: alluvial fan; atoll; cave; channel, alluvial (hydraulic geometry); cirque, glacial; crater; doline; drumlin; dune, aeolian; fjord; inselberg; karren; lake; landslide; palsa; pingo; river delta; tafoni; tor; volcano; yardang

IAN S. EVANS

GEOSITE

Geosites (synonyms: geotopes, Earth science sites, geoscience sites) are portions of the geosphere that present a particular importance for the comprehension of Earth history. They are spatially delimited and from a scientific point of view clearly distinguishable from their surroundings. More precisely, geosites are defined as geological or geomorphological objects that have acquired a scientific (e.g. sedimentological stratotype, relict moraine representative of a glacier extension), cultural/historical (e.g. religious or mystical value), aesthetic (e.g. some mountainous or coastal landscapes) and/or social/economic (e.g. aesthetic landscapes as tourist destinations) value due to human perception or exploitation. Various groups of geosites are generally specified in the reference literature: structural, petrological, geochemical, mineralogical, palaeontological, hydrogeological, sedimentological, pedological and geomorphological geosites. In the last case, they are also called geomorphological sites or geomorphosites. Some anthropic objects (e.g. mines) are also considered as geohistorical sites. Geosites can be single objects (e.g. springs, lava streams) and larger systems (e.g. river systems, glacier forefields, coastal landscapes). Active geosites allow the visualization of geo(morpho)logical processes in action (e.g. river systems, active volcanoes), whereas passive geosites testify to past processes; in this case, they have a particular patrimonial value as Earth memory (landscape evolution, life history and climate variations).

Geosites may be modified, damaged, and even destroyed, by natural processes and anthropogenic actions. In order to avoid damage and destruction, geosites need conservation. Conservation strategies are generally based on inventories of geosites requiring the development of assessment methods. Assessment is based on criteria such as integrity (whether the object is complete), exemplarity (to what extent the geosite is representative of the geology or geomorphology of a region or country), rarity (in the space of reference or in scientific terms), legibility (whether it is easily visible scientifically), accessibility (for pedagogic activities), vulnerability, paleogeographical value (its contribution to the history of the Earth), aesthetic value, and cultural/historical value. Geodiversity is a criterion used for assessing groups of geosites: geodiversity is higher where there is a concentration of different objects in a given space facilitating visits and protection. Several quantitative or qualitative procedures for evaluation exist in the literature.

Some countries have adopted specific legislation for geosite conservation (e.g. Great Britain has individuated Regionally Important Geological/Geomorphological Sites – RIGS). Generally, geosite conservation is relatively high in developed countries but low in developing countries.

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SEE ALSO: geodiversity; landscape sensitivity

EMMANUEL REYNARD

GILGAI

A form of micro-relief consisting of mounds and depressions arranged in random to ordered patterns (Verger 1964). There is a great variety of forms and they occur on a range of swelling clay and texture-contrast soils that have thick subsoil clay horizons. They tend to occur on level or gently sloping plains in areas subject to cycles of intense wetting and drying. Gilgai is an Australian aboriginal word meaning 'small waterhole' (Hubble *et al*. 1983) and some seasonal ponding of water does occur in some of the closed depressions of the larger forms.

The mechanisms of gilgai development involves swelling and shrinking of clay subsoils under a severe seasonal climate. A widely adopted hypothesis for their formation is as follows (Hubble *et al*. 1983: 31):

> when the soil is dry, material from the surface and the sides of the upper part of major cracks

falls into or is washed into the deeper cracks, so reducing the volume available for expansion on rewetting of the subsoil. This creates pressures which are revealed by heaving of the soil between the major cracks which, once established, tend to be maintained on subsequent drying. This process is repeated, with the result that the subsoil is progressively displaced, a mound develops between the cracks, and the soil surface adjacent to the cracks is lowered to form depressions.

However, some gilgai are linear forms, known colloquially as 'Adams furrows', 'black-men's furrows', 'stripy country' and 'wavy country' (Hallsworth *et al.* 1955). Beckmann *et al.* (1973: 365) see surface runoff and soil heaving as working together to produce such features, particularly on pediment slopes.

In the Kimberley there are individual linear gilgai up to 2 km long and it is possible that in their case aeolian processes have contributed to their development (Goudie *et al.* 1992).

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A.S. GOUDIE

GIS

A Geographic Information System (GIS) can be defined as a system of hardware and software used for the capture, storage, management, retrieval, display and analysis of geographic data. GIS's have been around since the 1960s, when, independently, initiatives in Canada (the Canadian GIS (CGIS), developed under the direction of Roger Tomlinson within the Canadian Federal Department of Agriculture) and in the United States (the Laboratory for Computer Graphics at Harvard University, established under the direction of Howard Fisher) resulted in the development of computer-based geographic information systems as we now know them (Foresman 1998).

A GIS uses two fundamental data types: spatial data, consisting of points (e.g. a sample location, a spot height), lines (e.g. the bank of a river, a break in a slope), areas (e.g. a drumlin, a drainage basin) and cells or rasters (e.g. a pixel from a satellite image), and attribute or descriptor data, consisting of characteristics associated with the spatial data (e.g. the elevation of the spot height, the area of the drainage basin).

While initial efforts during the 1970s and early 1980s saw the development of many one-off systems, the emergence in the 1980s of dominant players such as Environmental Systems Research Institute (ESRI) and Intergraph produced a shift from building systems to application development. While geomorphologists were among the first to appreciate the power that computers could bring to scientific analyses (e.g. Chorley 1972), a lack of spatial data and appropriate analytical procedures meant that, initially, such systems were not widely used within the scientific community. As well, the widespread acceptance of GIS required time for scientists to learn, apply and review the new technology in light of contemporary research problems. However, the rise in the computing power of personal computers, coupled with an increased availability of spatial data, meant that by the late 1980s GIS had become a tool used by many geomorphologists.

The earliest links between GIS and geomorphological research can be traced back to applications involving digital elevation models. DIGITAL ELEVATION MODELS (DEMs) are the digital representation of elevation, and while contours have traditionally been used to graphically represent topography on printed maps, in a GIS either a regular tessellation such as gridded cells (i.e. a raster representation, typically referred to as a DEM) or an irregular tessellation such as a Triangulated Irregular Network (TIN) (i.e. a vector representation consisting of elevation points and connecting lines forming triangular planar regions) are the preferred means for representing topography (Weibel and Heller 1991). The main concerns surrounding any digital representation of topography relate to issues around fidelity, accuracy and resolution (Moore *et al.* 1991). While TINs, with their variable sampling structure and their ability to represent important landform features such as peaks, ridges, cliffs and valleys, are capable of more accurately capturing the complexity of topography than are DEMs, DEMs remain the favoured means of representing topography for geomorphologists because of the ease with which analytical procedures can be applied to them (e.g. moving windows of 3 by 3 cells within which attributes such as slope and aspect can be quickly calculated).

The resolution of the data available to the analyst determines the areal extent of the study area that is appropriately analysed, and the type of features that can be identified. Initially, much digital data was of coarse resolution that limited the nature of the analysis (e.g. to regional or macro analyses). As higher resolution data become more commonplace, geomorphologists are better able to study processes at finer spatial and temporal scales, and to examine the role that scale plays in physical processes (Walsh *et al.* 1998).

Extraction of drainage features from any digital representation of topography remains a complex process, however. While with TINs the extraction of the drainage network can be an easy process, determination of the direction and accumulation of surface waters remains a geometrically complex task. For DEMs, the presence of artifactual sinks – either through errors in the elevation values or as a result of the discretization of the elevation values – complicates the process, and much effort has been directed at automatically recognizing and removing such features from DEMs (e.g. Maidment 1993).

Geomorphology is concerned with the form, the materials and the processes from which landforms are created. Some geomorphologists claim that a full understanding of materials and process can be obtained by focusing studies on the form of the landscape (Speight 1974) – a view shared by others in fields such as fractals. GIS are the ideal tools with which to study form, and it is not surprising to find that computer-based morphometric analysis of Digital Elevation Models (DEMs) has long been a very active area of research (e.g. Dikau 1989; Pike 1988; see GEOMORPHOMETRY). Early work focused on the derivation of topographic properties of watersheds, including the derivation of slope, curvature, channel links and drainage areas, all properties that can be mathematically derived from a DEM. This early work focused on examining morphologic patterns, rather than the physical processes that control them. While the link between form and process remains tentative, even at present, we are seeing an increasing sophistication in the application of quantitative approaches to the study of landform, along with the integration of more complex process models within Geographic Information Systems.

The use of GIS in geomorphology can be conceptually classified into four general types of analyses (Vitek *et al.* 1996; see also Walsh *et al.* 1998). These include: (1) landform measurement, (2) landform mapping, (3) process monitoring, and (4) landscape and process modelling. Often an application will involve several levels of GIS analysis. Process monitoring and landscape modelling, for example, will routinely require landform measurement and mapping in order to define initial parameters and establish the spatial distribution of controlling factors of interest. Although the utilization of GIS is not always required for carrying out these types of geomorphological analyses, given suitable digital data and processing capabilities, a GIS can provide a platform for automating these functions. This automation can greatly enhance the spatial/temporal scope and resolution of many geomorphological investigations. A general description and research example for each of the four different types of GIS utilization in geomorphology is provided below.

At the most basic level, a GIS may be used to perform fundamental measurements of landform features. This includes the enumeration of landform features, and making measurements of landform length, area and volume. Often it is some relation between fundamental measures that is of interest. Some common geomorphological examples include counting the number of landslides per region area (event frequency), measuring the length of channels per unit area (drainage density), taking the ratio of horizontal to vertical hillslope or channel lengths (slope), and calculating the areal extent of glaciers in an alpine catchment (glacial coverage). Carrying out these types of landform measurements is usually easily and efficiently accomplished using most standard GIS applications. By automating such measurements in a GIS environment, the spatial scope and resolution of the analysis can be expanded beyond what could be accomplished using traditional manual techniques. For example, Fontana and Marchi (1998) used this type of GIS analysis in order to evaluate the intensity of localized erosional processes in two alpine drainage basins of the Dolomite Mountains in northeastern Italy. In their work the combination of two landscape measurements of erosion potential was considered – the contributing drainage area (indication of flow concentration occurrence) and local slope (index of flow erosivity). By using the automated measurement capabilities of the GIS, these hydrologic parameters were calculated over entire drainage basin areas of many square kilometres using a high-resolution ten by ten square metre grid base. This type of analysis provided highly localized information on sediment erosion potential within the alpine catchments, information that would have been impossible to obtain using manual methods.

The second way in which GIS is utilized in geomorphology is in the development and analysis of landform maps (see TERRAIN EVALUATION; GEOMORPHOLOGICAL MAPPING). Landform mapping is commonly used in geomorphology in order to characterize landscapes and relate landform distribution to spatial patterns of physical, chemical and biological geomorphic processes. Through the use of basic GIS-based mapping tools, users can rapidly produce and modify landform maps. The typical map manipulation and analysis functions that are included in most GIS applications further enhance landform map production and interpretation. Such functions include map generalization and simplification, map overlay, spatial query and browsing, and various algorithms for the analysis of spatial patterns and relationships. Computer automation is becoming increasingly necessary in landform mapping because of the large amount of digital geographic data that is available and the rapidly increasing rate of digital geographic data acquisition. Bishop *et al.* (1998) used GIS-based mapping for studying large-scale geomorphic processes acting on the Nanga Parbat Himalaya massif of northern Pakistan. A digital elevation model and multispectral remote sensing data were used to study the structural geology and surface geomorphology of this extensive and remote mountain environment. Landform maps, a major component of the investigation, were developed using GIS software and the integration of some massive digital spatial data sets comprised of both surface and subsurface remote sensing data. Assessments of denudation rates and sediment storage were quantified and glacial, fluvial and mass movement processes were reconstructed for the massif by analysing the three-dimensional form and spatial surface characteristics of the mapped landscape in this study. Walsh *et al.* (1998) and Bishop *et al.* (1998) both stress the growing importance of the integration of remote sensing and GIS spatial analysis in order to solve complex, large-scale geomorphic problems.

The analytic capabilities of GIS are well suited to studies of river channel dynamics. At a fundamental level, GIS is an efficient tool for mapping channel features including, for example, sand and gravel bars, vegetated islands, channel banks, woody debris jams and historic (abandoned) channels. Commonly, channel features are digitized from existing maps, orthophotos, aerial photographs or satellite images and coded in the GIS as line or polygon features. Data collected in the field may also be included. If a scale or co-ordinate system has been defined, GIS is an efficient tool for quickly examining spatial relations between, and for measuring the length, width and area of digitized features. If maps or imagery are available for different dates, more advanced spatial and temporal overlay analysis can be performed such as lateral migration, loss or gain of riparian surfaces, aggradation or degradation of sediment, and changes in channel planform. Changes can be examined visually, or summarized and tabulated in a database, while rates of change may be additionally determined if exact dates are known. Similarly, if sufficient historical data are available, GIS may further be used to show trends in morphologic development over time, and even predict landform evolution. The capability of most GIS to collate data sources with different scales and co-ordinate systems is key to these types of analysis. In recent years, GIS has been used increasingly to study the relation between (morphologic) form and (hydraulic) process in river channels using fully distributed topographic models of the channel bed and banks for different dates (cf. Lane *et al.* 1994). Topographic information may be derived from conventional cross-section surveys, tacheometry, photogrammetry or depth soundings. The data are imported to the GIS in order to produce either a TIN or DEM, and then are overlaid in order to produce maps and volumetric summaries of channel scour (erosion) and fill (deposition). The net difference between channel scour and fill may then be used to infer rates of sediment transport within the framework of a sediment budget.

The seamless integration of environmental models with GIS is the ultimate goal of many researchers (Raper and Livingstone 1996). Since landscape and process modelling requires the storing, retrieving and analysing of spatio-temporal data sets, it is not surprising that GIS can play a significant role in this area. GIS enhances the process as it assists in the derivation, manipulation, processing and visualization of such geo-referenced data. Boggs *et al*. (2000) used an integrated approach to look at landform evolution in a catchment that could be subject to impacts from mining activities. Landform evolution models typically require extensive parameterization, often involving both hydrology and sediment transport models, and using an integrated environment allows for the rapid production of modified input scenarios. Therefore, a much wider array of impact scenarios can be made, and the environmental implications of decisions made today can be modelled over the long term, which should lead to better management decisions.

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BRIAN KLINKENBERG, ERIK SCHIEFER AND DARREN HAM

GLACIAEOLIAN (GLACIOAEOLIAN)

The association between glaciation (past and present) and aeolian processes and forms. Glaciation comminutes rock fragments by grinding and so produces material, including silt and sand, that can then be transported by wind to create dunes and LOESS deposits. Such materials are also available to cause wind abrasion in glacial and near-glacial environments, thereby producing wind-moulded pebbles and cobbles (VENTIFACTS) and streamlined ridges (YARDANGS) and grooves. Deflation from glacial deposits can create STONE PAVEMENTS (Derbyshire and Owen 1996).

The fine materials blown across proglacial plains and beyond create dunes and sandsheet deposits (coversands). These are widespread in Canada and Central Europe. The facies progression from coversands through to sandy loess and then loess is well documented from the proglacial forelands around the northern hemisphere. Above all, however, the thick loess deposits of Central Europe, Central Asia, China, New Zealand, the Argentinian Pamas and the mid-USA, may at least in part be indirect products of glacial sediment supply.

Ice sheets and glaciers may themselves affect air pressure conditions and generate high velocity winds that contribute to the power of aeolian processes in their vicinity.

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A.S. GOUDIE

GLACIAL DEPOSITION

Glacial deposition occurs when debris is released from glacial transport at the margin or the base of a GLACIER. Narrow definitions include only primary sedimentation directly from the ice into the position of rest, but broader definitions include secondary processes such as deposition through water and resedimentation of glacigenic materials by flowage. In addition, release of material onto the glacier surface is sometimes referred to as supraglacial deposition, but if the glacier is still moving this is only a temporary stage in the sediment transport process.

Glacial deposition produces characteristic sediments and landforms, and landscapes of glacial deposition exist across large areas of the mid-latitudes formerly covered by ice sheets. The term till is generally used to describe sediments deposited by glaciers, and replaces the term boulder clay, which was commonly used in the past. Landforms created by glacial deposition are called MORAINES. Glacial sediments are often unstable in non-glacial environments and are subject to reactivation under PARAGLACIAL conditions. Glacially deposited materials are therefore an important debris source in geomorphic processes in postglacial and proglacial environments.

The characteristics of glacial sediments reflect both the processes of their deposition and also the processes of GLACIAL EROSION and entrainment by which the material was originally produced. A substantial literature exists on the classification of glacial sediments and the processes by which they form (e.g. Schlüchter 1979; van der Meer 1987; Goldthwait and Matsch 1988) and several convenient summaries of depositional processes have been produced (e.g. Whiteman 1995). Following a broad definition, the main mechanisms of deposition include:

1. release of debris by melting or sublimation of the surrounding ice
2. lodgement of debris by friction against a substrate
3. deposition of material from meltwater (glacifluvial deposition)
4. chemical precipitation
5. flow and resedimentation of deposited material
6. glacitectonic processes (see GLACITECTONICS).

Different processes of sedimentation are dominant in different parts of a glacier. In supraglacial locations, material can be released by ablation of the ice surface. This occurs most commonly by melting, and is referred to as melt-out, but sublimation can make a significant contribution in cold arid environments (Shaw 1988). Englacial debris, and debris in basal ice penetrating to the surface, is then exposed on the ice surface and can contribute to a supraglacial sediment layer. This supraglacial sediment is liable to redistribution by flow, wash and mass movements on the ice surface as ablation continues. Resedimented material derived from flow of supraglacial debris, sometimes referred to as ‘flow-till’, can make up a substantial proportion of glacial deposits in environments where supraglacial sedimentation occurs (e.g. Boulton 1968). The extent of supraglacial sedimentation depends on the debris content of the ice, the ablation rate and the rate of removal of sediment from the surface by processes such as wash, deflation and mass movement. Supraglacial sediment can be deposited at the ground surface when the glacier retreats or disintegrates.

At glacier margins material can be released by ablation and dumped directly into the proglacial environment. Sediment can be released directly from within the ice, carried to the margin supraglacially and dropped over the margin as if over the end of a conveyor belt, or brought to the margin by water flowing from the interior or surface of the glacier. Sediment can also be transferred to the margin by deformation of subglacial material (e.g. Boulton *et al.* 1995). The amount of sediment that is transported to the margin is primarily a function of the size and speed of the glacier, the glacier’s erosive capability, the erodibility of the substrate and the input of sediment from extra-glacial sources such as rockfalls or tephra. The characteristics of the sediments and landforms produced by deposition at the margin depend on whether the margin occurs on land or in water, and the processes and environments of their formation are reflected in their morphology and sedimentological structure. Reactivation and resedimentation of material at the margin by movement of the ice or by fluvial and mass movement processes is common.

In subglacial environments, material can be released by ablation either in cavities or in contact with the bed, or can be lodged against the bed by moving ice. Release of basal sediment by melt-out or sublimation beneath moving ice can produce conditions conducive to lodgement, to the development of thick basal till, and to subglacial deformation of the released material. Lodgement of sediment against a rigid bed occurs when the friction of the clast against the bed outweighs the tractive power of the ice and material is released from the ice by either pressure melting or plastic deformation of ice around clasts. Lodgement can also occur against the upper surface of a deforming bed if the overlying ice is moving faster than the deforming layer. Deposition within a deforming subglacial layer occurs when deformation cannot remove all of the material that is being supplied to the layer and deformation ceases either throughout the layer or for a certain thickness at its lower margin.

Subglacial deposits can also include chemical precipitates, although these are not always considered in the context of glacial sediments. Chemical precipitates such as calcite can be deposited when solute-rich waters freeze. In carbonate environments, such as where limestone bedrock is present, comminuted carbonate rocks contribute to a highly reactive rock flour. Meltwater produced at the bed can take carbonate into solution, and if the water subsequently refreezes the carbonate can be released as a precipitate onto bedrock or basal debris (e.g. Souchez and Lemmens 1985).

Many glaciers release sediment into water, and GLACIMARINE and GLACILACUSTRINE sediments form an important part of the glacial sediment record. Release of sediment into water can produce different effects from terrestrial sedimentation. Sediment can be released directly from the ice, by discharge of sediment-rich meltwater, and by the melting or breakup of icebergs. The characteristics of subaqueous sediments and landforms reflect aqueous as well as glacial processes and conditions. Where a glacier is grounded on its bed beneath the water, glacial deposition can occur beneath the margin by lodgement and melt-out. However, material that is dumped from the front of the glacier into water, or from the base of the glacier where the glacier is floating, forms a deposit that is not strictly a glacial deposit, as its character upon settling will be controlled largely by aqueous sedimentation processes. The principal features of ice margins in water include subaqueous moraines caused both by pushing of proglacial sediments and release of sediment from the glacier; subaqueous grounding line fans formed from material emerging from beneath the glacier into the water at the grounding line; ice contact fan deltas that form when grounding line fans grow and emerge at the water surface; and a distal proglacial zone in which sediment settles out from suspension in the water and rains out from icebergs drifting away from the ice margin. Useful reviews of sedimentation at marine and lacustrine glacier margins include those provided by Dowdeswell and Scourse (1990) and Powell and Molnia (1989).

The mechanisms of glacial deposition impart specific characteristics to the deposited material. Fabric, particle size and shape characteristics, and consolidation have all been used to infer depositional processes and glacier characteristics from glacial sediments. The distribution of structures in deformed sediments can be used to reconstruct former ice sheets, and Boulton and Dobbie (1993) suggested that consolidation characteristics of formerly subglacial sediments can be used to infer basal melting rates, subglacial groundwater flow patterns, ice overburden, basal shear stress, ice-surface profiles and the amount of sediment removed by erosion. Glacial deposits can thus provide valuable information about glacial and climatic history.

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SEE ALSO: glacier; moraine

PETER G. KNIGHT

GLACIAL EROSION

Glaciers cause erosion in a variety of ways (Bennett and Glasser 1996). First of all, glaciers can be likened to conveyor belts. If a rockfall puts coarse debris on to a glacier surface, for example, or if frost-shattering sends down a mass of angular rock fragments on to the glacier surface, it can then be transported, almost whatever its size, down valley. Second, beneath glaciers there is often a very considerable flow of meltwater. This may flow under pressure through tunnels in the ice at great speed, and may be charged with coarse debris. Such subglacial streams are highly effective at eroding the bedrock beneath a glacier. This can contribute to the excavation of TUNNEL VALLEYS. Meltwater may also cause chemical erosion. Third, although glacier ice itself might not cause marked erosion of a rock surface by *abrasion*, when it carries coarse debris at its base some abrasion can occur. This grinding process has been observed directly by digging tunnels into glaciers, but there is other evidence for it: rock beneath glaciers may be *striated* or scratched, and much of the debris in glaciers is ground down to a fine mixture of silt and clay called *rock flour*.

Glaciers also cause erosion by means of *plucking*. If the bedrock beneath the glacier has been weathered in preglacial times, or if the rock is full of joints, the glacier can detach large particles of rock. As this process goes on, moreover, some of the underlying joints in the rock may open up still more as the overburden of dense rock above them is removed by the glacier. This is a process called *pressure release*.

As debris-laden ice grinds and plucks away the surface over which it moves, characteristic landforms are produced which give a distinctive character to glacial landscapes. Of the features resulting from glacial quarrying, one of the most impressive is the CIRQUE. This is a horseshoe-shaped, steep-walled, glaciated valley head. As cirques evolve they eat back in the hill mass in which they have developed. When several cirques lie close to one another, the divide separating them may become progressively narrowed until it is reduced to a thin, precipitous ridge called an *arête*. Should the glaciers continue to whittle away at the mountain from all sides, the result is the formation of a pyramidal *horn*.

With valley glaciation the lower ends of spurs and ridges are blunted or truncated; the valleys assume a U-shaped configuration; they become more linear; and hollows or troughs are excavated in their floors. Many high-latitude coasts, such as those of Norway, New Zealand and western Scotland, are flanked by narrow troughs, called FJORDS, which differ from land-based glacial valleys in that they are submerged by the sea.

Fjards are related to fjords. They are coastal inlets associated with the glaciation of a lowland coast, and therefore lacking the steep walls characteristic of glacial troughs. A good example of a fjard coast is that of Maine, USA.

A further erosional effect of valley glacier is the breaching of watersheds, for when ice cannot get away down a valley fast enough – perhaps because its valley is blocked lower down by other ice or because there is a constriction – it will overflow at the lowest available point, a process known as *glacial diffluence*. The result of this erosion is the creation of a *col*, or a gap in the watershed.

Tributary valleys to a main glacial trough have their lower ends cut clean away as the spurs between them are ground back and truncated. Furthermore, the floor of a trunk glacier is deepened more effectively than those of feeders from the side or at the head, so that after a period of prolonged glaciation such valleys are left hanging above the main trough. Such *hanging valleys* have often become the sites of waterfalls.

The development of an ice sheet tends to scour the landscape. In Canada there are vast expanses of territory where the Pleistocene glaciers scoured and cleaned the land surface, removing almost all the soil and superficial deposits and exposing the joint and fracture patterns of the ancient crystalline rocks beneath. Streamlined and moulded rock ridges develop, including *roches moutonnées*. In parts of Scandinavia and New England these may be several kilometres long and have steepened faces of more than 100 m in height. They are interspersed with scoured hollows which may be occupied by small lakes when the ice sheet retreats. In western Scotland relief that is dominated by this mixture of rock ridges and small basins is called *knock and lochan* topography. Areas of calcareous rocks that have been scoured by ice often display LIMESTONE PAVEMENTS.

Considerable debate has been attached to the question of rates of glacial erosion (Summerfield and Kirkbride 1992) and some have argued (see GLACIAL PROTECTIONISM) that glaciers can protect the underlying surface from erosion. Much depends on the context in which a glacier or ice sheet occurs, but even for an area like the Laurentide Shield there is controversy (see Braun 1989).

A range of methods has been used to measure amounts of glacial erosion. These have included:

1 The use of artificial marks on rock surfaces later scraped by advancing ice.
2 The installation of platens to measure abrasional loss.
3 Measurements of the suspended, solutional and bedload content of glacial meltwater streams and of the area of the respective glacial basins.
4 The use of sediment cores from lake basins of known age which are fed by glacial meltwater.
5 Reconstructions of preglacial or interglacial land surfaces.
6 Estimates of the volume of glacial drift in a given region and its comparison with the area of the source region of that drift.
7 Cosmogenic nuclides (Colgan *et al*. 2002).

Many published rates of glacial erosion are high, typically 1,000–5,000 $m^3 km^{-2} a^{-1}$, but they vary greatly according to the measuring techniques that are used (Warburton and Beecroft 1993) and to the nature of the environment (Embleton and King 1968). Clearly, geographical location is an important control of the rate of glacial erosion. Some areas have characteristics that limit the power of glacial erosion (e.g. resistant lithologies, low relief, frozen beds), but other areas suffer severe erosion (e.g. non-resistant lithologies, proximity to fast ice streams, thawed beds, etc.).

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A.S. GOUDIE

GLACIAL ISOSTASY

'ISOSTASY' (= equal standing) refers to an equal distribution of pressure or mass within a column of rock that extends from the surface of the Earth to its interior. This concept of equilibrium is, however, perturbed by forces at the surface and within the Earth which create unequal mass distributions. With time, such inequalities are compensated for by adjustments within the lithosphere and aesthenosphere, and the rate and relaxation time for this recovery to equilibrium is a function of the Earth's rheology. The term 'glacial isostasy' is thus used to define the adjustments of the Earth's surface to the growth and disappearance of ice sheets (Andrews 1974). For example, 22,000 years ago the surface of the Earth differed fundamentally from today's geography, as large ice sheets, several kilometres thick, covered much of Canada and most of the area centred on the Baltic (the Fennoscandian ice sheet), large areas of Great Britain, and many

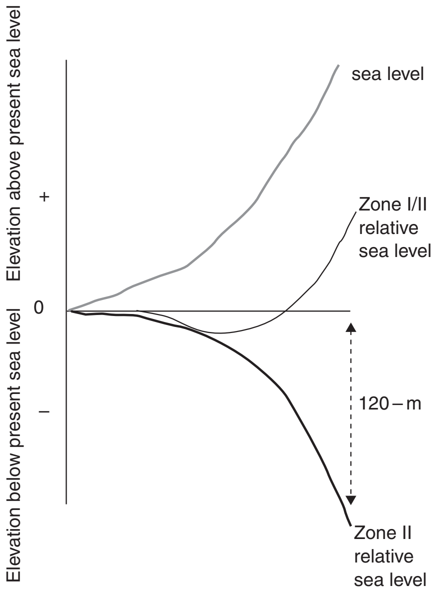

Figure 69 Schematic depiction of changes of relative sea level at sites within the borders of former ice sheets (Zone I), at sites distant (>1,000 km) from the maximum ice extent (Zone II), and sites in the transition between Zones I and II

other parts of the world (Denton and Hughes 1981). Together these ice sheets and glaciers extracted about 120 m of water from the global ocean, thus reducing the water load on the seafloor. During the last 15,000 years or so the ice sheets melted, resulting in a reduction in the load of ice sheets on continents, but the resulting meltwater has added to the load over ocean basins (Peltier 1980). Thus 'glacial isostasy' in its most complete definition includes the global-wide changes in the differential elevation of land due to ice sheet removal, and changes in relative sea level around all the world's coastlines caused by the unique combination of ice and water loading and/or unloading at each location (Figure 69, Zones I, I/II, and II) (Clark *et al*. 1978).

Observations on changes in sea level within formerly glaciated areas have been made for over 150 years, but the link between the removal of the ice load, crustal recovery and isostasy were made in north-west Europe and North America by the mid to late nineteenth century (Andrews 1974). Although significant research was carried out in the early part of the twentieth century studies of glacial isostasy bloomed in the period after ~1960 due in no small part to the development of radiocarbon dating, and to increased levels of field research in Arctic Canada, Greenland and Svalbard. In these areas materials to date the changes of sea level through time (molluscs, whalebone and driftwood) were abundant. The combination of a technique (^{14}C dating) plus exploration of vast tracts of formerly glaciated areas resulted in an explosion of data on changes in sea level and on the delimitation of former ice sheet margins through time (Andrews 1970; Blake 1975; Dyke 1998; Dyke and Peltier 2000; Forman *et al*. 1995). In the mid- to late 1970s these data-rich field observations attracted the interests of the geophysicists who now had both the mathematical tools and the computer power to tackle what is a global Earth-science problem with many ramifications (Peltier and Andrews 1976).

Studies of 'glacial isostasy' represent a significant interplay between workers in several different fields (Figure 70), including (1) the glacial geologist who maps the time-dependant changes in ice sheet extent (and more problematically, thickness); (2) the Quaternary scientist working on changes in sea level from sites within the margins of present ice sheets (near-field sites), to those at and just beyond the former ice sheet limits, and finally to workers reconstructing changes in sea level at sites far-distant (the far-field) from ice margins (say the central Pacific Islands); (3) glaciologists who combine data from (1) above with knowledge of the physics of ice to model changes in ice sheet extent and volume; (4) the geophysicists who 'tune' the behaviour of the Earth's rheology to match the data from inputs (1) and (2). The schematic interplay between these disciplines (Figure 70) has resulted in a series of successive approximations of each of the key components. This process started in 1976 (Peltier and Andrews 1976) and is still continuing.

The rheology of the Earth is most frequently modelled as a self-gravitating viscoelastic (Maxwell) solid (Cathles 1975), although a case can be made for a more complex rheology where the response is a nonlinear (power > 1) function, not unlike the behaviour of glacial ice. A simple 2-D model consists of a lithosphere of some thickness which overlies a fluid aesthenosphere. The lithosphere is rigid with the application of a small

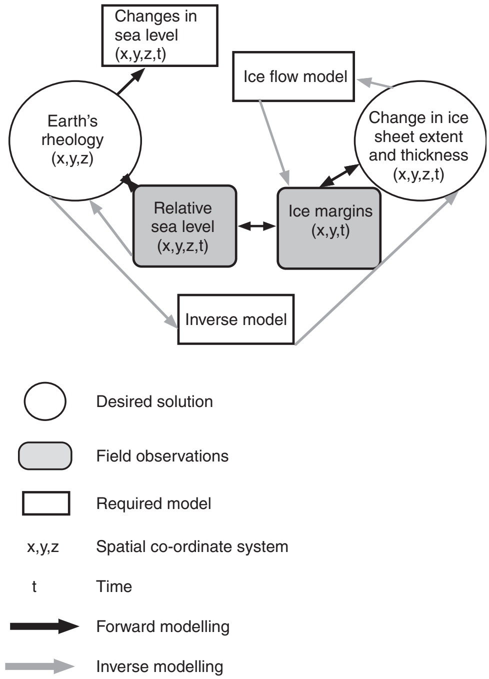

Figure 70 Schematic diagram on interactions between data and models

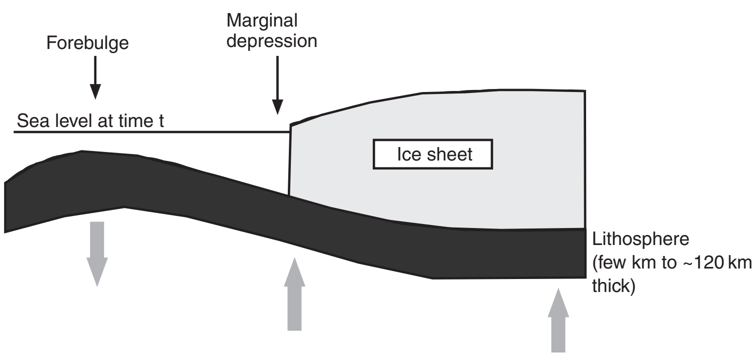

Figure 71 Two-dimensional model of an Earth with a lithosphere and aesthenosphere. The distance between the ice margin and the forebulge is a function of the thickness and elastic properties of the lithosphere. The senses of motion are those during the retreat phase of an ice sheet

load ('small' being a function of the lithosphere thickness, but on old continental shields the load may have to exceed a diameter of ~300 km, whereas on Iceland, with the mantle virtually at the surface, then the load diameter is probably a few kilometres to a few tens of kilometres at most for isostatic compensation by flow to be induced. As an ice sheet grows on land at some point the lithosphere will bend and material will be displaced by flow. It is generally assumed that the flow can be approximated by a layered Newtonian viscous fluid with a viscosity in the range of 10^{22} poises. The 2-D model of the Earth's response to a glacial load (Figure 71) shows that at the margins of the ice sheet the load is partly supported by the lithosphere so that there is a depression at the ice margin but at some distance from the ice sheet there is a zone of uplift in the forebulge. Upon retreat of the ice sheet the forebulge will collapse, hence there will be a rise in relative sea level.

Figure 70 shows the elements of the problem. Glacial isostasy has been examined in terms of 'forward models' and 'inverse models' in a full 3-D global model. In reality there is an ongoing iteration between the field scientists and the modellers so that both approaches are required (Lambeck 1995; Lambeck *et al.* 1998; Peltier 1994; Peltier 1996). In the 'forward' case, the explicit data are the positions of the ice sheet margins through time (x, y, t), and changes in relative sea level at a suite of sites from Zones I, I/II, and II. What then has to be approximated is the changes in thickness within an ice sheet (x, y, z, t). The application of this time-dependent load to a model of the Earth's rheology will result in changes in sea level (required model) where the predicted changes in sea level include not only the obvious fall of sea level in Zone I, but also account for the transfer of mass (meltwater) from the melting ice sheets to the oceans. Disagreements between the observed relative sea levels (Figure 70) and the predicted sea levels could be related to either an incorrect Earth rheology or an incorrect estimate of changes in the ice sheets in all four dimensions. In contrast, inverse modelling is an attempt to develop a model of the global ice sheet changes by taking the observed relative sea-level data, assuming a rheology, and then using these data to reconstruct the changes in the ice sheets. Appropriate ice flow models can then be applied to see if the reconstructions are glaciologically feasible and whether the reconstructed ice sheets match the data, in this case the mapped and dated ice sheet margins.

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JOHN T. ANDREWS

GLACIAL PROTECTIONISM

The belief that the erosive power of rain and rivers far exceeds that of glacier ice, and that the presence of glaciers in a region protects the landscape from much more effective fluvial attack (Davies 1969). Glaciers were thought, following Ruskin, to rest in depressions like custard in a pie dish, rather than to erode the basins. Proponents of this theory included the British geologists J.W. Judd, T.G. Bonney, E.J. Garwood and S.W. Wooldridge.

In some areas, where ice stream velocities are low and relief is limited, the erosive role of glaciers may well be passive rather than active. Some degree of protection may be afforded.

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A.S. GOUDIE

GLACIAL THEORY

The belief that in former times glaciers had been more extensive than they are today. Some suggestions as to this had originally been made at the end of the eighteenth century. In 1787 de Saussure recognized erratic boulders of palpably Alpine rocks on the slopes of the Jura Ranges, and Hutton reasoned that such far-travelled boulders must have been glacier-borne to their anomalous positions. Playfair extended these ideas in 1802, but it was in the 1820s that the *Glacial Theory*, as it came to be known, really became widely postulated. Venetz, a Swiss engineer, proposed the former expansion of the Swiss glaciers in 1821, and his ideas were supported and strengthened by Charpentier in 1834. The poet Goethe expressed the idea of 'an epoch of great cold' in 1830. However, the ideas of both Venetz and Charpentier were extended and widely publicized by their fellow countryman, Louis Agassiz, who was one of the originators of the term *Eiszeit* or *Ice Age*. In Norway Esmark put forward similar ideas in 1824, and in 1832 Bernhardi went so far as to suggest that the great German Plain had once been affected by glacier ice advancing from the North Polar region.

In spite of this convergence of opinion from numerous sources, these ideas were not easily accepted or assimilated into prevailing dogma, and for many years it was still believed that glacial till, called drift, and isolated boulders, called erratics, were the result of marine submergence, much of the debris, it was thought, having been carried on floating icebergs. Sir Charles Lyell noted debris-laden icebergs on a sea-crossing to America, and found that such a source of the drift was more in line with his belief in the power of current processes – UNIFORMITARIANISM – than a direct glacial origin.

Even towards the end of the nineteenth century some opposition still remained. In 1892, for instance, H.H. Howorth produced his massive neocatastrophist *The Glacial Nightmare and the Flood – a second appeal to common sense from the extravagance of some recent geology*, and tried to return to a fundamentalist-catastrophic interpretation of the evidence.

Conversely, others were overenthusiastic about the glacial theory and Agassiz himself postulated that glaciers reached the humid tropics in South America.

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A.S. GOUDIE

GLACIDELTAIC (GLACIODELTAIC)

The discharge of sediment-laden streams from melting glaciers into lakes or fjords usually results in accumulation of a delta. The morphology and structure of the delta depends on several factors. These include nature of the ice margin; a vertical glacier front calving into a fjord contrasts strongly with a gently sloping ice surface descending into a shallow lake overlying dead ice. The sources of meltwater, flowing directly from the ice or entering the water body at the surface, deeper within or at the base of the ice as overflows, interflows or underflows, are also important. The quantity and particle size of sediment, the channel gradient, and the location of deposition close to or distant from the ice margin all influence the form, structure and sedimentary characteristics of the delta.

Deltas formed by discharge of glacial meltwater are located either in the proglacial environment or in ice-contact situations. In the proglacial environment they are deposited in lakes and fjords, and distinct types referred to as Hjulstrom, Gilbert and Salisbury deltas have been identified. In ice-contact situations deltas may form in supraglacial lakes, subglacial water bodies and lakes of the terminoglacial environment.

Hjulstrom deltas occur either in lakes or fjords where outwash sheets (the Icelandic sandur) of gravel and sand enter shallow water with gently sloping fronts. They may also form fans or small deltas in lakes on the ice surface. The delta gravels and sands are the flood traction loads of braided sandur channels or of fan channels that on deposition in standing water become smaller in grain size away from the point of discharge. Beyond the delta front, which advances into the water as a series of lobes due to changes in channel discharge points, the suspended silts and clays are deposited as prodelta mud.

The Salisbury type delta is intermediate in type and size between the Hjulstrom and Gilbert-type deltas. It forms where high-energy flow from a subglacial tunnel mouth supplies material rapidly via sheet- and streamfloods, and topset beds accrete very rapidly.

The classic glacigenic delta is the Gilbert type that may be formed on ice, in proglacial lakes and in fjords where the water is deep enough to allow development of a distinct structure that includes bottomset, foreset and topset beds. The bottomset beds occur beneath and beyond the foreset beds and extend as prodelta clays to merge with the lake – or fjord floor clays. The bottomset beds result when the suspended sediment carried in turbid flows beyond the delta front settles. The beds are generally laminated. The laminae reflect grain-size variations between fine sand, silt and clay layers due to deposition that may be controlled by seasons, short periods or single events. The sediments become finer away from the point of discharge. They may contain occasional dropstones derived from floating ice, and convoluted laminations due to disturbance of the delta face by slumping, sediment flowage and turbidity currents. Sediment loading and dewatering will also produce convoluted laminations.

The foreset beds dip steeply (*c*.25–30°) away from the source of the glacial sands and gravels and prograde into the lake by the intermittent avalanching down the delta front of cohesionless debris flows. The foresets decrease slightly in slope and sediment size towards the delta face, and individual foresets tend to decrease in grain size upwards. Channels that result from erosion of the delta face by turbidity flows may be cut into the foreset beds. As the delta builds up to water level the glacifluvial sandur deposits extend onto the surface as topset beds. The topset gravels and sands are coarser than those of the foresets, are gently inclined (2–5°), relatively thin (*c*.1–2 m) and exhibit cut-and-fill structures related to the shifting courses of the braided channels of the sandur.

Small ponds and lakes develop on the decaying marginal and terminal parts of glaciers. Meltwater from surface streams or shallow depth within the ice may form small deltas in the shallow lakes. The delta sediments consist of inclined beds of gravel and sand interbedded with unsorted debris flow sediments from the ice surface. Subsequent melting of the underlying and laterally supporting ice causes slumping and flowage of the sediments with the development of fault and fold structures. Where preserved, the sediments form delta-kames.

Subglacial and englacial streams may enter bodies of standing water at or near the base of the ice. Reduction in stream velocity will cause sedimentation of the sands and gravels to form a delta. Decay of the ice may result in formation of a delta-kame.

More extensive lakes are frequently formed in the ice contact terminoglacial environment of glaciers and ice sheets. Where large meltwater streams enter lakes or fjords, deltas of the Gilbert type are formed. Bottomset beds may be formed where the water body is relatively large but most of the delta sediments consist of foreset beds of sand and gravel. Topset beds only develop where a sandur forms between the meltwater outlet and water body. Interstratification of debris flow deposits and foreset sands and gravels is common. Removal of ice support during decay causes collapse and pitting of the ice proximal delta margin. Such deltas have been described as kamiform and where subglacial stream outlets are numerous along an ice edge many may be developed and may coalesce laterally as a kame moraine. The Salpausselka Moraines formed in southern Finland during the final readvance stage, the Younger Dryas, of the last glaciation are over 600 km in length and largely form delta moraines.

In southern Finland the three major Salpausselka delta moraines – formed in the Baltic Ice Lake before the postglacial rise in sea level drowned the Baltic Sea – record retreat stages of the ice sheet margin. Where deltas are formed sequentially inland along fjord margins the stages of glacier retreat in the valley and relative sea-level rise in the fjord can be detected. When ice barriers impounded glacial lakes in upland valleys and formed deltas and shorelines at a number of water levels, as in Glenroy, Scotland, the delta surfaces and shorelines record the stages in draining of the glacial lake.

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ERIC A. COLHOUN

GLACIER

Glaciers are accumulations of snow and ice on the Earth's surface. They form predominantly at high latitude (polar regions) and at high elevation (on mountains). Here, two meteorological variables, temperature and precipitation, combine to yield conditions where the annual amount of snowfall (predominantly during the cold or wet season) outweighs the annual amount of snow melt (predominantly during the warm or dry season). Under these conditions, consecutive annual snow layers develop, one on top of the other, the pressures of which force snow at depth to change structure and density (recrystallization). These conditions first produce firn/névé (density of 0.400–0.830 $kg\,m^{-3}$) and, when pores are sealed off to create air bubbles, ice (density of 0.830–0.917 $kg\,m^{-3}$). These changes can occur within one year and at shallow depths in maritime regions or take hundreds to thousands of years and considerable depths in continental locations. Glaciers are said to have formed when glacier flow occurs, that is when ice is thick enough to deform plastically under its own weight (Paterson 1994).

Glacier systematics

Glaciers are commonly ordered using their geomorphological or thermal characteristics, both of which yield a tripartite system. The common characteristic of ICE SHEETs and ice caps (smaller versions of the former, i.e. $<50{,}000\,km^2$) is that they are so thick relative to the landscape relief on top of which they rest, that their surface topography and flow direction are unconstrained by the underlying topography (except near the ice margin). Their surface morphology dominantly shows the presence of ice domes, the dome-shaped central regions where ice forms and flows outward towards its perimeter, and outlet glaciers and ice streams, the narrow flow-parallel bands of faster flowing ice that are the primary routes by which the ice is evacuated in marginal areas (Plate 51A). Two ice sheets and numerous ice caps occur in the polar regions, the latter typically on upland plateaux.

The margins of ice sheets and ice caps, especially in Antarctica, frequently become afloat in the ocean that surrounds them. These floating sections, or ice shelves, normally cover shallow marine embayments or continental shelves with many islands. Eventually, slabs of ice break off from the ice cliffs that border ice shelves, thus producing (tabular) ICEBERGs.

The surface morphology and flow of glaciers (alpine glaciers, mountain glaciers) follow the morphology of the subglacial landscape. The morphology of ice fields, thinner than ice caps and lacking the characteristic ice dome, characteristically shows mountain uplands covered by ice, except for the highest ridges and peaks (see NUNATAK). Valley glaciers, elongated features confined by valley walls, occur as outlets of the ice-covered uplands and as glaciers on their own (Plate 51B). Their heads (highest section) often start in bedrock depressions (see CIRQUE) and their snouts (lowest section) reside within a valley. However, when the glacier emanates and terminates immediately beyond the valley mouth (i.e. beyond its lateral constraint), the ice tongue spreads outward to form a wide piedmont lobe, forming a piedmont glacier (Plate 51C). If outlet glaciers and valley glaciers extend offshore they form tidewater glaciers (Plate 51A). Cirque glaciers or glacierets are located within, or just barely extend beyond, the amphitheatre-like basins in which they form, and they tend to be wide rather than long (Plate 51D). Glaciers on steep mountain slopes above bedrock depressions, hanging glaciers or ice aprons, primarily loose mass as blocks of ice become detached and avalanche downslope. Sometimes, these and other avalanches supply the mass necessary for rejuvenated or regenerated glaciers to exist in locations where precipitation alone would be insufficient to support a glacier (Hambrey 1994; Sharp 1988; Sugden and John 1976).

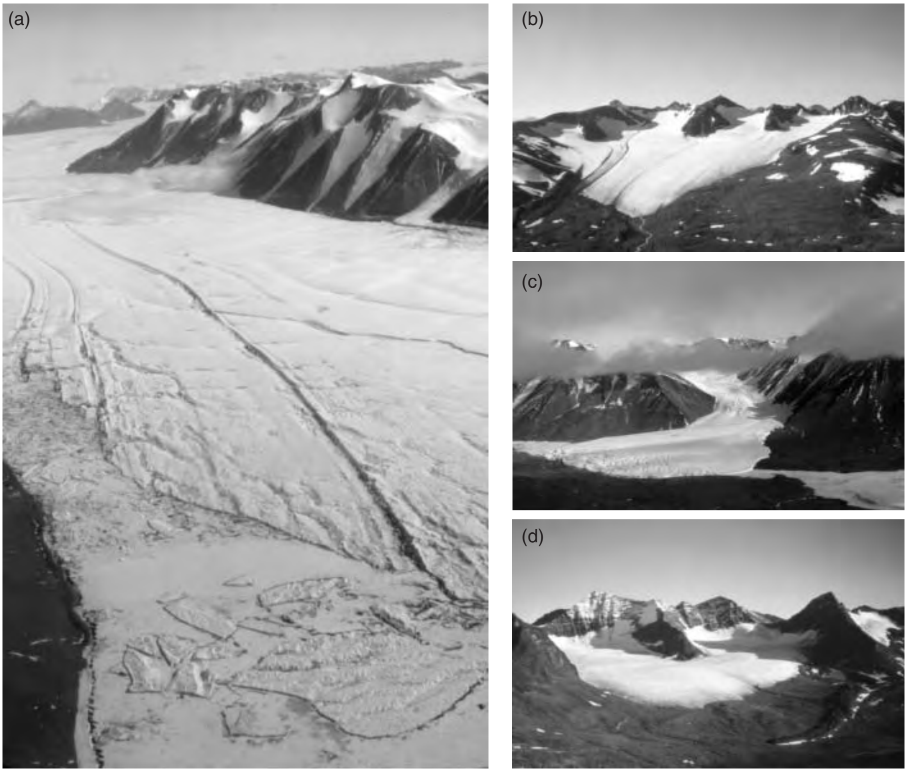

Plate 51 (a) outlet glacier, Antarctica; (b) valley glacier, Sweden; (c) piedmont glacier, Antarctica; and (d) cirque glacier, Sweden

Ordering glaciers by their thermal characteristics yields the (high) polar and temperate glacier-type end members, and the subpolar glacier type for those that combine both former characteristics. The ice in polar glaciers is below freezing throughout and meltwater is absent, except, maybe, for surface melting during short periods in summer. The ice in temperate glaciers is close to melting throughout, except, usually, for surface freezing during winter. Hence, the glacier has a temperature close to its melting point, which is 0 °C for pure ice at atmospheric pressure, but occurs at lower temperatures (pressure melting) when pressurized at depth. However, many glaciers experience both freezing and melting conditions; they are subpolar or polythermal glaciers. In polar regions they experience abundant surface melting and heating throughout the summer and in temperate regions they have a thick cold surface layer that is maintained throughout the summer (Ahlmann 1935).

Mass balance of glaciers

Most glaciers are nourished by snowfall and depleted by snow and ice melt. Because the conditions favourable for snow to fall and snow and ice to melt are related to the temperature of the atmosphere over the glacier surface, and because the mean annual temperature decreases with increasing altitude (lapse rate), the surface of a glacier normally shows two or more predictable zones

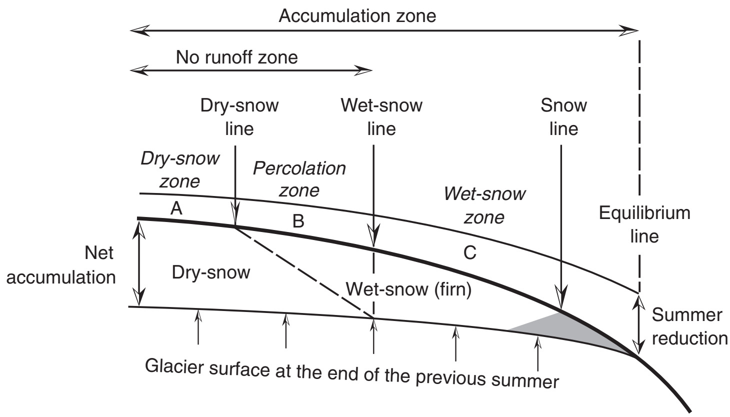

Figure 72 The surface structure of the accumulation area. The heavy line is the glacier surface at the end of the summer, the light line is the previous winter surface. The grey-shaded region is superimposed ice. Volume reduction by (A) recrystallization, (B) melt (and refreezing at depth) and (C) melt and runoff. Modified from Paterson (1994: 10)

(Figure 72). In the highest elevation zone, the accumulation zone, temperatures are at a minimum yielding conditions where the snowpack that accumulates during the winter or wet season is not melted entirely during the summer or dry season, thus creating an annual mass gain (accumulation). In the lowest elevation zone, the ablation zone, temperatures are higher and last winter's snowpack plus an additional amount of the underlying solid ice melts, creating an annual mass loss (ablation). The MASS BALANCE OF GLACIERS denotes the balance between the amount of snow remaining after the summer integrated across the accumulation area (and converted to the amount of water it represents when melted, m w.e. or metre water equivalent) and the amount of solid ice lost underneath the snowpack integrated across the ablation area (in m w.e.). When mass balance is positive, and more and more mass is added each year, the glacier will grow and expand. Conversely, when mass balance is negative over many years, the glacier will shrink and contract (retreat). When ice reaches considerable thickness a positive feedback mechanism occurs – because of the higher elevations the conditions for ice accumulation improve, especially close to the snout. The boundary between the accumulation and ablation zones is a relatively narrow zone, or line, where the annual mass balance is zero, the EQUILIBRIUM LINE OF GLACIERS (which exists on all glaciers that are not strongly out of equilibrium with contemporary climate, i.e. where the whole glacier surface becomes an accumulation area or ablation area). Some glaciers have additional boundaries, lines, in their accumulation areas (Figure 72). The dry-snow line borders the dry-snow zone, a region of extreme climate (in polar areas and at extremely high elevations), where no surface melt occurs, even in summer. The wet-snow line is the lower limit of the percolation zone, where snow melts at the surface and refreezes at depth during the summer. The refreezing of meltwater occurs at depth because the snowpack is initially cold, but the refreezing process releases heat and warms the snowpack until it reaches melting temperatures throughout by the end of the summer at the wet-snow line. Strictly taken, above the wet-snow line there is no mass loss (no runoff zone). The snow line is the lower boundary of the wet-snow zone, where all remaining snow at the end of the ablation season is at the melting temperature. The firn line, not shown, is the boundary between ice and firn at the end of the summer, and may coincide with the snow line (only in temperate regions where snow may transform to firn in one summer can this latter situation occur). Although there is mass loss throughout the wet-snow zone during the summer, it is characterized by a positive annual mass balance. Meltwater that has refrozen at depth, forming ice lenses, may become particularly extensive and form superimposed ice layers underneath the firn in the wet-snow zone. These can visually crop out at the surface of a glacier at the end of the summer season between the snow line and the equilibrium line (Oerlemans 2001; Paterson 1994).

Alternative important components in the annual mass balance of glaciers are mass gain by avalanching and rime, and mass loss through avalanching, calving and sublimation.

Glacier flow

Structural glaciology, the geomorphology of glacier surfaces, shows that bodies of ice experience differential movements between the glacier bed and the surface, between the lateral margins and the glacier centre, and between different locations along its longitudinal profile. The most conspicuous features are crevasses, foliation structures, and band- or wave ogives (Forbes bands) (Hambrey 1994: 61–69; Paterson 1994: 173–190; Sugden and John 1976: 71–78). For example, a visible sign of glacier flow, except for these ice surface structures, is the creation of a BERGSCHRUND, an opening between (stagnant ice on) the valley wall and the glacier that pulls away from the wall as it moves downslope (Figure 73).

Crevasses, which are fissures in the ice surface with a typical depth of 25–30 m, occur abundantly on glacier surfaces, are often consistent in their direction over limited distances but vary considerably along the length of a glacier, and

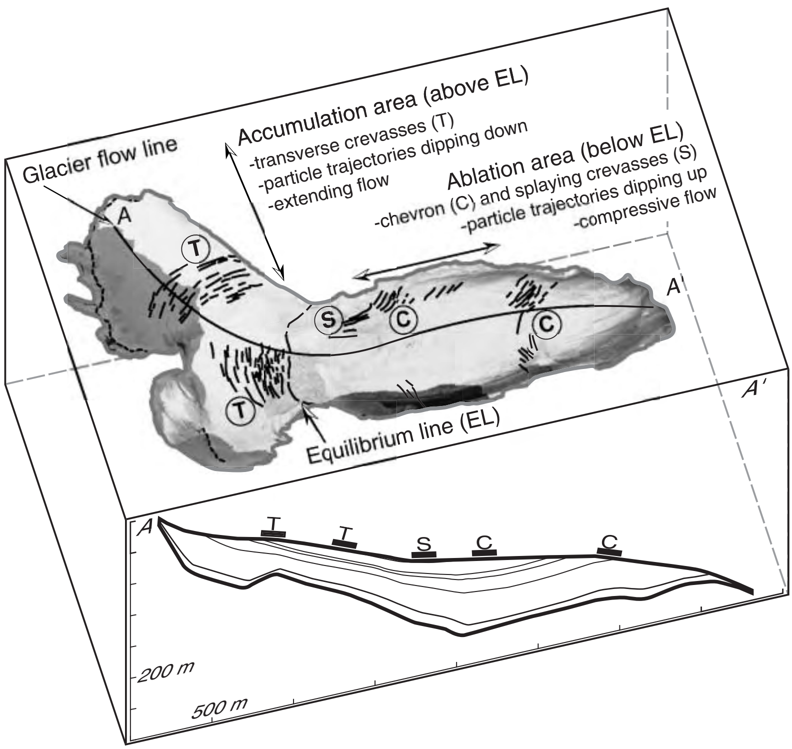

Figure 73 Glacier flow from surface structure and particle trajectories. Example is from Storglaciären, northern Sweden. Particle trajectories modified from Pohjola (1996). Note the location of the bergschrund (dashed) in the upper part of the accumulation area

will form where at least one principle stress is tensile and exceeds the tensile strength of the ice. The direction of crevasses can be predicted from a straightforward geometric analysis of stress distributions. Hence, the pattern of crevasses found in many accumulation areas of glaciers, transverse crevasses, can be related to the existence of extending flow (when ice flow accelerates downslope; i.e. where the glacier bed steepens, where the glacier is joined by another branch, on the outside of a glacier bend, and at the grounding line, where glaciers become afloat). Conversely, most glacier ablation areas reveal a pattern of crevasses, splaying crevasses, which are expected when the stress situation is dominantly compressive (when ice flow decelerates downslope; i.e. where the glacier bed flattens or rises, where the valley widens, on the inside of a bend, and at the snout; Hambrey 1994: 56). Finally, where the stress situation is neither dominantly extending, nor compressive, the lateral drag of the valley walls results in a third predictable stress and crevasse pattern, chevron crevasses, a feature common to middle sections of alpine glaciers. The width and depth of the crevasses depend primarily on the temperature of the ice, i.e. they are widest and deepest in polar glaciers.

Foliation structures result from previous structures that have been compressed (pure shear) and stretched (simple shear) to attain a direction that dominantly is parallel to ice flow. The original structures include the near-horizontal layering during accumulation (and the formation of ice lenses) and vertical or inclined elongated structures, such as crevasses, and point structures (moulins). Foliation structures, therefore, are consistent with differential motion through the ice body.

Forbes bands, or ogives, can form as glacier ice flows through an ice fall. Ice falls are causally related to the occurrence of very steep glacier beds. Here, the tensional stresses result in transverse crevassing and thinning while the ice moves through the fall. However, summer and winter conditions result in a modification of the ice as it moves through. Ice ablation during the summer results in thinner ice (and dirty because of dust collection) arriving at the foot of the fall. Precipitation during the winter, on the other hand, fills the crevasses with snow, resulting in thicker and cleaner ice arriving at the foot of the fall. For ice falls where the ice moves through within a year's time span, the resulting situation at the foot of the fall, where flow is compressive and crevasses are closed, are alternating ridges of blue ice (winter) and depressions of white ice (summer). These structures are convex in the direction of ice flow (even though the crevasses were slightly concave), indicating the differential motion in the transverse direction and, because of the crevassing, in the longitudinal direction.

The geomorphology of glacier surfaces, therefore, was the initial guidance in an understanding of some of the basic characteristics of glacier motion by the middle of the century (e.g. Tyndall 1860). Modern advancements in understanding the flow of glaciers, dating from the past fifty years, were mainly theoretical in nature (Glen 1952; Nye 1952; Weertman 1957), realizing that ice behaves like a crystalline solid, leading to theorems that have subsequently been verified experimentally (e.g. Raymond 1971).

From basic mass balance principles it must follow that a glacier of constant shape and volume (steady-state situation) must transfer all the annual mass gain in the accumulation area to the ablation area to compensate for the annual mass loss. For each transverse cross section through the glacier it is pertinent that the discharge of ice through that section per year balances the integrated annual mass balance across the up-glacier surface. Hence, this requires the discharge to be at a maximum at the equilibrium line, and normally this equates to having the thickest ice and highest ice flow velocities there, and implies the existence of extending flow in the accumulation area and compressive flow in the ablation area. Snow that falls in the accumulation area is successively compressed by subsequent layers of snow (until glacier ice is formed, which is incompressible), yielding a small velocity component towards the glacier bed. A cube of solid ice, when subjected to the range of pressures by the burden of overlying snow, firn and ice (typically less than 1 bar), will deform (stretching), thus departing on a trajectory that parallels, but is dipping slightly away from, the glacier surface (Figure 73). To satisfy mass continuity, this yields a convergence of flow in the accumulation area, and, typically, a thickening of the glacier downslope. Conversely, a cube of ice in the ablation area follows a trajectory which parallels, but is slightly directed towards, the slope of the ablating ice surface (Figure 73). This results in a divergence of flow in the ablation area, and, typically, a thinning of the glacier towards its lateral margin and snout.

As mentioned, the movement of ice is a gravity-driven internal deformation of the ice body. Because the deformation is at a maximum where the pressures are at a maximum, the deformation occurs primarily close to the glacier bed ('dragging' the overlying ice along). Because the amount of deformation higher up in the ice body, although dramatically less than at its bed, occurs in addition to the deformation for deeper ice layers, the velocity vector of ice flow close to the surface is measurably higher than at its bed.

Like other solids, ice deforms more readily when it is close to its PRESSURE MELTING POINT, a trait of temperate glaciers. Typical for pressure melting conditions is the presence of water (fluid state) and ice (solid), side by side, in the glacier body. The presence of water at the ice–bedrock interface is of importance in the motion of glaciers because its lubricating effect (especially when pressurized) facilitates the sliding of ice over its substrate (basal sliding). When ice covers sediment, basal sliding will normally be insignificant, but enhanced deformation of the water-saturated sediment may occur. Basal sliding would be very effective over a smooth bedrock surface, but, often, there are bedrock protrusions which hamper the effectiveness of the sliding process. When the obstructions are relatively small ($<10^{-2}$–10^{-1} m-size), ice melts on the upstream pressurized side, flows around the obstacle as water, and refreezes in its wake as the pressure drops, thus producing heat that can be used to help melting the ice on the upstream side. This experimental verification of the presence of regelation ice in the wake of obstacles is one of the modern findings of glacier flow. When the obstructions are relatively large (10^{-1}–10^{1} m-size), they create pressures so large that ice will deform more readily around it (by enhanced plastic deformation), a process that is less effective than regelation and thus more hampering to ice flow.

As noted, the effectiveness of glacier motion is in large part dependent on the temperature of the ice. Because most of the motion occurs in the basal ice as deformation and between the ice and the bedrock by sliding, it is specifically the subglacial temperature which is of interest. Temperate glaciers are warm-based (wet-), and their surface velocities integrate basal sliding and effective ice deformation. Polar glaciers are cold-based (dry-), which inhibits basal sliding and their surface velocities only reflect the integrated effect of ice deformation (at sub-optimal temperatures). Subpolar glaciers may have a bed at the pressure melting point and behave like temperate glaciers, or otherwise are cold-based and behave like polar glaciers.

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SEE ALSO: glacial deposition; glacial erosion; moraine

ARJEN P. STROEVEN

GLACIFLUVIAL (GLACIOFLUVIAL)

Glacifluvial is an adjective that applies to the processes, sediments and landforms produced by water flowing on, in and/or under glaciers and away from glacier snouts. Consequently, glacifluvial environments may occur in supraglacial, englacial, subglacial, ice-marginal and proglacial locations of alpine and continental glaciers both present and past. Proglacial, glacifluvial environments are often transitional to fluvial environments when the processes, sediments and landforms become dominated by non-glacial tributary inflows rather than the annual rhythm of melting ice (Lundqvist 1985).

Meltwater supply and pathways

Water enters a glacier from melting ice, snow melt, rainfall, hillslope runoff and the release of stored water. Consequently, most glacifluvial environments are dominated by strong seasonal, diurnal and episodic discharge variations. The release of stored water from subglacial, englacial, supraglacial or ice-marginal lakes and reservoirs produces floods several orders of magnitude greater than 'normal' melt-related flows. Such floods and megafloods (peak discharges estimated at 10^6–$10^7\,m^3.s^{-1}$ compared to $10^5\,m^3.s^{-1}$ for the Amazon today) are known as OUTBURST FLOODS or jökulhlaups.

The path of meltwater through a glacier is determined by water supply (amount and location), ice temperature and dynamics and basal substrate. Pathways include: (1) supraglacial meandering channels; (2) englacial passages; (3) subglacial channels cut up into the ice (ice tunnels) or down into the bed, broad flows (sheets), linked-cavities, films, canals or aquifers; and (4) proglacial channels, broad flows and jets (into standing water – lakes, oceans).

Glacifluvial processes

Glacifluvial processes include erosion, transport and deposition. The type, rate and effectiveness of meltwater erosion are influenced by the nature of the basal substrate (sediment, bedrock), meltwater supply and pathway, and sediment supply. Mechanical erosion is effective because rapidly flowing meltwater in channels and broad flows is turbulent and carries a high sediment load (see SEDIMENT LOAD AND YIELD). Mechanical erosion includes: (1) hydraulic action – the force of water against its bed lifts or drags loose debris into the flow; (2) CAVITATION – shock waves and microjets resulting from the formation and implosion of bubbles of vapour (cavities) cause rock pitting or loosen mineral grains; and (3) abrasion – the impact and grinding of rock particles carried by the flow on themselves (attrition) and on flow boundary materials causes wear. Chemical erosion, or DISSOLUTION, occurs when meltwater removes soluble minerals in bedrock and debris. Subglacial dissolution is particularly effective because freshly abraded bedrock and debris present a high surface area for chemical reactions and solutes are continually flushed from the system. Together, meltwater erosion processes act to fracture, round and wear down bedrock and rock fragments and result in a variety of erosional landforms.

Landforms resulting from meltwater erosion

Meltwater erosion produces meltwater channels, bedrock erosion marks (s-forms) and may form some DRUMLINS, flutings, ribbed and hummocky terrain (Shaw 1996).

Meltwater channels may form (1) along the ice margin, (2) proglacially, or (3) subglacially. These channels exist in a variety of sizes (cm to km wide, cm to hundreds of m deep, decimetres to tens of km long) and substrates (bedrock, sediment); channels associated with past ice sheets are typically larger than those associated with alpine glaciers today. They may be differentiated by their slope and relationship to other glacial landforms. Ice-marginal (lateral) channels typically form parallel to the ice margin of cold-based glaciers, are often left perched and nested on valley sides during glacier retreat and their slope approximates that of the glacier surface at the time of formation. Advancing glaciers may divert rivers parallel to their ice front forming URSTROMTÄLER. Proglacial channels can form braided patterns (multiple shallow channels separated by bars) across outwash plains and always follow downslope paths (Plate 52a). Catastrophic drainage of ice-dammed or proglacial lakes has produced (1) trench-like valleys or spillways, and (2) networks of dry channels and waterfalls (see SCABLAND; Plate 52b). Subglacial channels may follow upslope paths, cross-cut drumlins and contain ESKERS (Plate 52c). TUNNEL VALLEYS or channels are long (can be >100 km), wide (<4 km), flat-bottomed, overdeepened (<100 m), often radial or anabranched subglacial channel systems that formed under ancient ice sheets. They can be buried by thick fills, muting their topographic expression. They may have formed catastrophically by meltwater erosion during the waning, channelized stages of subglacial megafloods (Shaw 1996).

Bedrock erosion marks, or s-forms, come in a variety of shapes and sizes (cm to km scale) and are found on the beds of bedrock channels and on broad bedrock plains. They are classified according to their shape and to the direction of formative

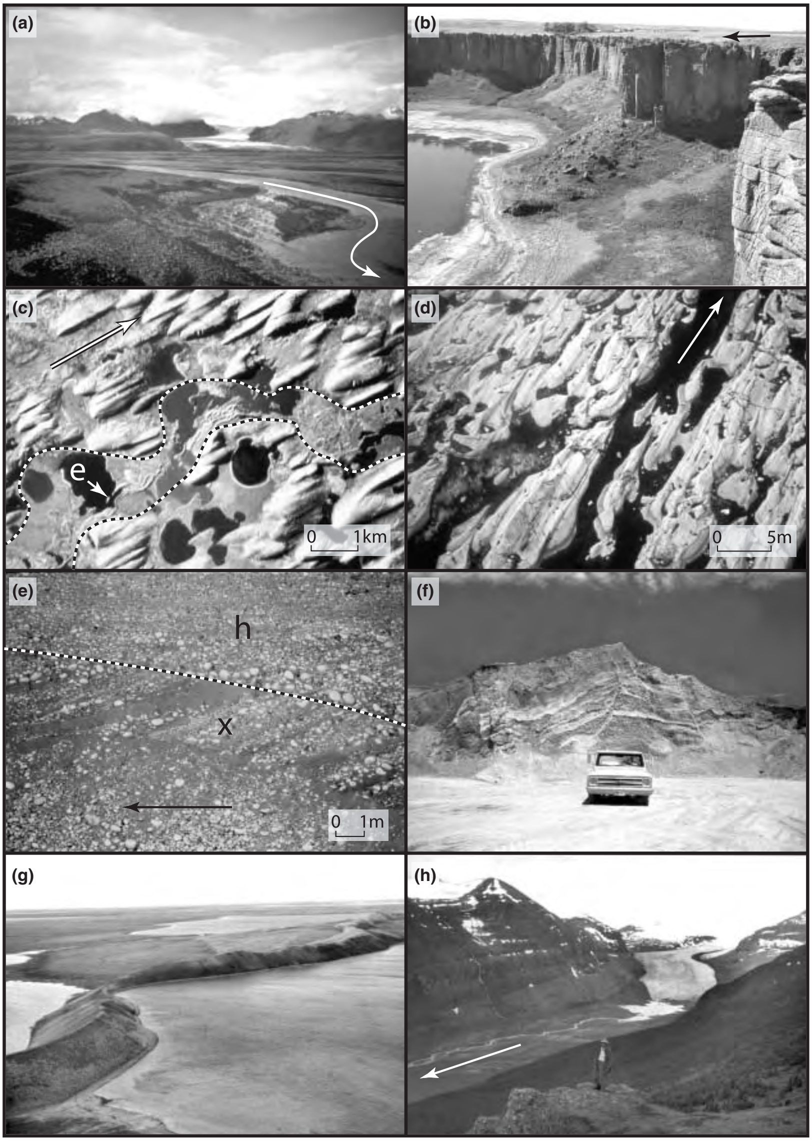

Plate 52 (a) braided outwash on Icelandic sandur (Skeiđaràrsandur, courtesy C. Simpson); (b) dry falls along Shonkin Sag spillway (USA, backwall ~55 m high); (c) cavity-fill drumlins truncated by tunnel channel (dashes) containing esker [e] (Livingstone Lake, Canada; aerial photograph A14509–77 © Her Majesty the Queen in Right of Canada, Centre for Topographic Information (Ottawa), Natural Resources Canada); (d) bedrock erosion marks (French River, Canada; Shaw 1996, © Wiley and Sons Ltd. Reproduced with permission); (e) cross-stratified [x] and horizontally-bedded [h] cobbles (Harricana esker, Canada); (f) faulted sand and gravel lithofacies (Campbellford esker, Canada); (g) sinuous esker (Victoria Island, Canada); (h) valley train (Saskatchewan glacier, Canada). Arrows indicate direction of formative water flow

flow. Mussel shell-shaped scours (muschelbrüche), sickle-shaped scours (sichelwannen; Plate 52d), comma-shaped scours and transverse troughs form mainly transverse to flow direction. Rock drumlins, rat-tails (tapering rock ridges extending from resistant rock knobs), flutes (spindle-shaped scours), furrows (Plate 52d) and cavettos (channels on vertical walls) form parallel to flow. POT-HOLEs record vertical scour. Together s-forms often exhibit a directional consistency, and a hierachical and systematic arrangement consistent with erosion by turbulent subglacial megafloods (Shaw 1996; Figure 74).

Drumlins are elongated hills of various sizes (up to ~2 km long, tens of m high, hundreds of m wide) and shapes (inverted spoon, parabolic, transverse asymmetrical, spindle-shaped; Plate 52c). They occur in en-echelon patterns and in fields spanning hundreds of kilometres. Some flutings are long (tens of km), narrow (hundreds of m) remnant ridges that occur downflow from escarpments. Ribbed terrain is composed of fields of coalescent and subparallel, convex-upflow and crenulate-downflow ridges (up to 30 m high) that are formed transverse to flow. Hummocky terrain is identified as a field of mounds (<10 m high and ~100 m diameter) and hollows. These subglacial bedforms are often transitional to one another and the material within them may be truncated at the landsurface. Consequently, some drumlins (inverted spoon-shaped drumlins), flutings, ribbed and hummocky terrain are attributed to erosion by turbulent subglacial megafloods (Shaw 1996; Figure 74).

Glacifluvial sediment

Glacifluvial sediment is mainly derived from (1) material supplied to the glacier surface from valley side rock falls, debris flows and avalanches, or (2) meltwater erosion of bedrock, sediment or debris-rich ice along its flow path. Typically, sediment transport varies with glacial environment and discharge, being greatest in subglacial channel and broad flows and during OUTBURST FLOODs. Deposition rates are generally greatest at the ice margin where meltwater issues from ice tunnels onto open outwash plains or into standing-water bodies.

Glacifluvial sediment is deposited in BEDFORMs which vary in scale from centimetres (e.g. ripples) to hundreds of metres (e.g. bars or macroforms). Bedforms are preserved in the sedimentary record as lithofacies (Plate 52e). Lithofacies are sedimentary units distinguished by their physical and/or chemical characteristics such as colour, texture, structure and mineralogy. By applying experimental relationships derived from pipe and flume studies, lithofacies are used to reconstruct former bedforms, flow conditions (see PALAEOHYDROLOGY) and glacifluvial environments. For example, cross-laminated sand records ripples deposited during low flow, whereas horizontally bedded gravel may record the movement of gravel sheets across the tops of macroforms during flood flows (Plate 52c).

Glacifluvial sediment may be found in ice-contact environments (supraglacial, englacial, subglacial and ice-marginal) or proglacial environments beyond an apron of stagnant ice. It shows many of the same characteristics as sediment of non-glacial rivers – both are typically characterized by sorted and stratified sand and gravel lithofacies. However, glacifluvial deposits differ from non-glacial fluvial deposits in several ways. (1) Glacifluvial sediment is typically coarse grained (boulder through sand sizes) as the flow velocity is generally too high for settling of the finest particles (silt and clay); such particles become trapped in lateral and distal standing-water bodies. (2) Course-grained lithofacies may exhibit relatively poor sorting (large grain-size range) and rudimentary bedding when rapidly deposited during the waning stages of an outburst flood. (3) Glacifluvial sediments typically exhibit abrupt changes in lithofacies (Plate 52e) due to pronounced seasonal and episodic changes in flow regime. (4) Ice-contact glacifluvial deposits frequently include flow deposits (diamicton) and till balls (eroded ice deposits) and exhibit structures indicative of shearing, faulting (Plate 52f), slumping and subsidence. These characteristics develop due to the proximity to a moving glacier and to the melting of buried or supporting ice.

Landforms resulting from meltwater deposition

Meltwater deposition forms ice-contact and proglacial landforms. Ice-contact landforms may include drumlins, ribbed and hummocky terrain, some crevasse-fill ridges and De Geer MORAINEs, eskers, KAMEs, kame terraces, outwash fans and some end, grounding-line and interlobate moraines. Proglacial landforms include outwash plains, valley trains and sandurs. The main differences between these two categories are (1) ice-contact landforms may contain structures

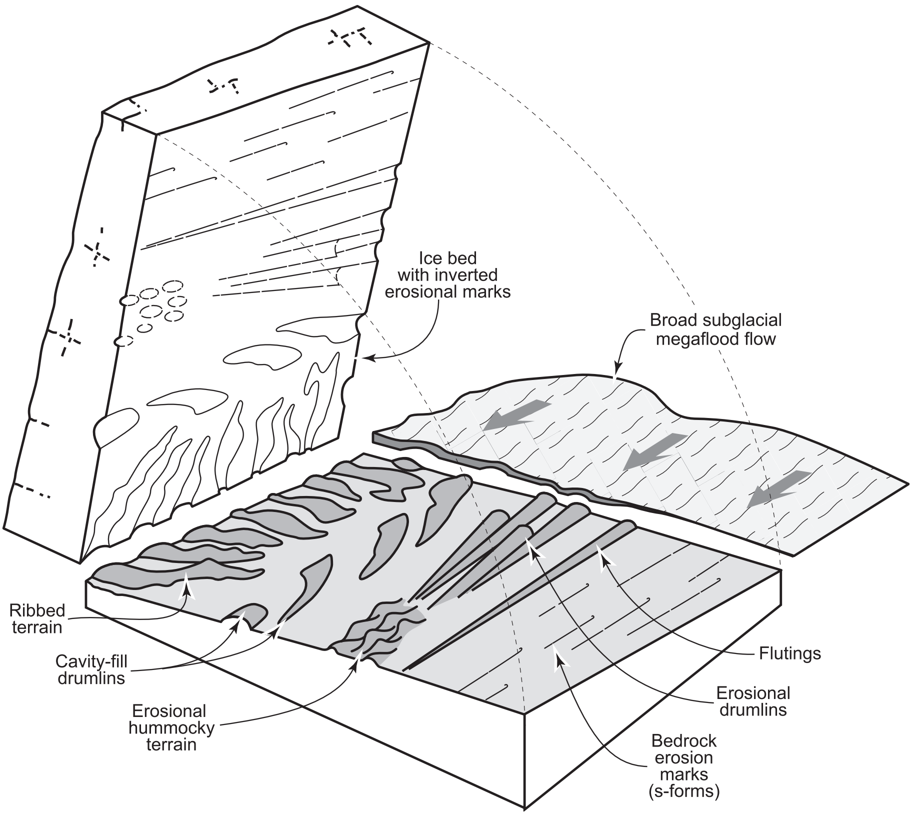

Figure 74 A model for subglacial landforms produced by broad subglacial megafloods (modified from Shaw 1996)

indicative of proximity to a moving glacier (e.g. thrust faults and shears) and/or the melting or removal of buried or supporting ice (faulting, slumping, pitted surfaces; see KETTLE AND KETTLE HOLES), and (2) proglacial landforms may contain sediment indicative of both glacifluvial and non-glacial processes.

Some drumlins (spindle, transverse asymmetrical and parabolic forms; Plate 52c), ribbed and hummocky terrain may have formed by deposition into cavities incised into the ice base during subglacial megafloods (Shaw 1996). As the broad megafloods waned, sediment carried in the flow was rapidly deposited into flow parallel, transverse and non-directional cavities forming cavity-fill drumlins, ribbed and hummocky terrain respectively (Figure 74). These landforms are composed of glacifluvial sediment with a sedimentary architecture that conforms to the landform shape.

Crevasse-fill ridges are linear, low ridges (up to ~10 m high) that are arranged in a pattern that mimics the radial and transverse crevasse patterns of a glacier. They formed by the infilling of crevasses within or at the base of glaciers (Figure 75). De Geer moraines are linear, low (<10 m) ridges that occur in fields subparallel to one another and in locations where the glacier was in contact with a standing-water body. They may form at the glacier margin during punctuated glacier retreat or in subglacial crevasses. Both ridge types may contain glacifluvial sediment.

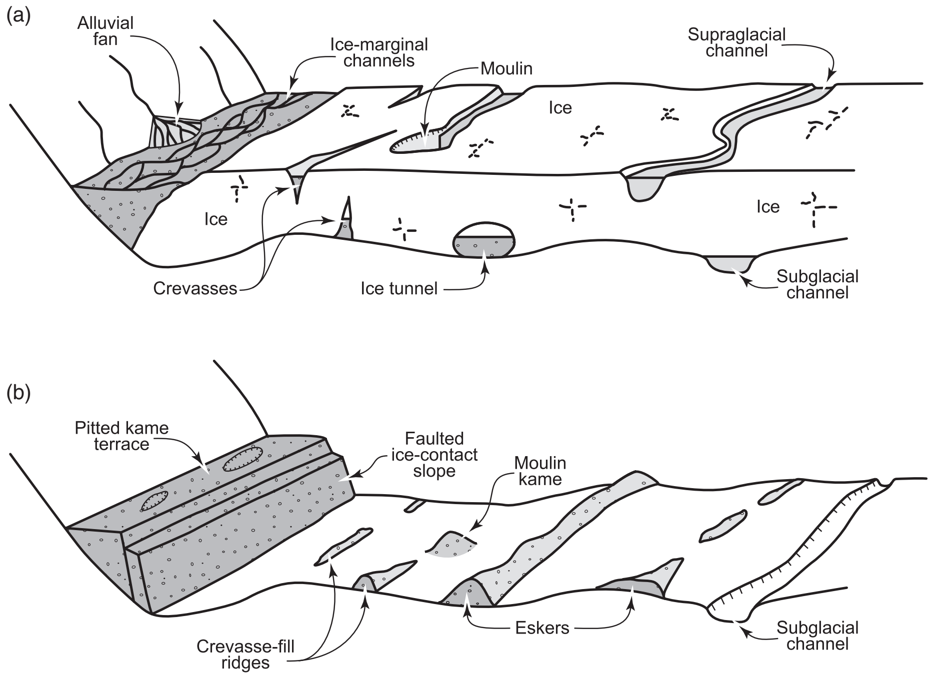

Figure 75 Development of glacifluvial ice-contact landforms (a) before and (b) after stagnant ice melt

Eskers form narrow, sinuous ridges (Plate 52g) or a series of ridges separated by broader beads. They occur in a range of sizes (m to tens of m high, m to hundreds of m wide, tens of m to hundreds of km long). They can be located in valleys or follow upslope paths. They may occur in isolation or in groups forming subparallel, deranged (not aligned with regional ice flow) or dendritic (tree-like) patterns. Eskers record the location of past ice-walled streams – mainly subglacial streams as the ridge-form is often lost when supraglacial or englacial channel deposits are lowered to the ground during glacier melt (Figure 74). Narrow ridges are tunnel deposits, whereas beads contain macroform, fan or delta sediments that formed where tunnel flow expanded into a subglacial cavity or at the ice margin. Strings of beads may indicate the punctuated retreat of the glacier front. Deranged eskers likely formed under stagnant ice. Long (hundreds of km), dendritic esker systems may have (1) developed in long and persistent (perhaps operating for hundreds of years) subglacial tunnels in stagnant ice or (2) formed almost instantaneously (over perhaps weeks or months) during the drainage of large subglacial reservoirs or supraglacial lakes.

Kames are steep-sided, variously shaped mounds of sand and gravel that were originally deposited with two or more ice-contact margins (Figure 75). Examples include moulin deposits (moulin kames) and small deltas or fans deposited at or on the ice margin (delta kames). Kame terraces are linear, often pitted benches of sand and gravel deposited by braided rivers which flowed between the valley side and the ice margin. They need not be altitudinally matched on both sides of a valley glacier.

Outwash fans are fan-shaped bodies of downstream-fining sediment with their apex at a meltwater portal. Deposition on land results in subaerial outwash fans and deposition in water (at a grounding line) results in subaqueous outwash fans. Adjacent outwash fans may coalesce forming a ridge along the ice front – an end moraine (on land) or grounding-line moraine (in water), or between ice lobes – an interlobate moraine. Past ice sheets have left many such glacifluvial moraines. Subaerial outwash fans often grade downflow into proglacial stream deposits.

Outwash plains are planar landforms containing proglacial stream deposits. Proglacial streams are often braided (Plate 52a, h) as high sediment loads, fluctuating discharge and a lack of vegetative anchoring results in a high degree of channel instability. Channels vary in width (m to hundreds of m) and bars vary in size (m to hundreds of m long, m relief). Braided streams continually evolve with each successive flood by channel scour and fill, bar development and overbank deposition. Where numerous proglacial streams issue from the ice front onto an open lowland an extensive outwash plain known as a sandur is formed (Plate 52a). When proglacial streams are hemmed in by valley sides in mountainous terrain, deposition is focused along the valley axis resulting in thick valley fills and a linear outwash plain called a valley train (Plate 52h). Where proglacial rivers enter standing-water bodies deltas (see GLACIDELTAIC) may form.

Outwash plains typically exhibit downflow changes in their morphology and composition reflecting a lessening of glacier influence and a decrease in energy away from the glacier snout. Close to the glacier (proximal) coarse gravel devoid of vegetation is arranged into longitudinal bars separated by a few large channels. The surface is often pitted. Grain size is highly variable reflecting strong melt and flood cycles. With increasing distance from the ice front transverse bars separated by a complex network of braided channels become prevalent, grain size decreases (sand and gravel are present) and grain roundness increases due to selective sorting and abrasion during transport. Lithofacies variation within and between beds is diminished as inflowing non-glacial tributaries dampen discharge fluctuations. Distally (furthest from source), flow in shallow braided channels, sheets or meandering streams deposits sand.

Megafloods and climate change

A broad suite of glacial landforms are now attributed to megafloods from past ice sheets (Figure 74; other explanations are also debated, see DRUMLINS; SUBGLACIAL GEOMORPHOLOGY). The discharge of such enormous quantities of freshwater across continents and into the oceans caused sea level to rise and may have modified ocean and atmospheric circulation, heralding climate change. As meltwater drainage, ice temperature and dynamics (movement) are linked, glacifluvial sediments and landforms contain a record of ice and water behaviour that is essential in our quest to understand climate change.

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SEE ALSO: esker; drumlin; glacier; kame; meltwater and meltwater channel; outburst flood; scabland; subglacial geomorphology; tunnel valley

TRACY A. BRENNAND

GLACILACUSTRINE (GLACIOLACUSTRINE)

Modern and ancient glacilacustrine deposits tend to be variable in grain size, mineralogy, bedding thickness and sedimentary structures, reflecting the broad range of settings in which they accumulate. Glacial lakes may originate from ice erosion of bedrock, in depressions of glacial deposits, or be impounded behind drainage barriers composed of moraine, outwash or ice (Hutchinson 1957). Today, lakes of glacial origin outnumber all other lake types combined. However, most present-day glacial lakes owe their origin to Pleistocene glacial activities and are now under no direct influence of glaciers. Thus it is useful to distinguish glacial lakes and their deposits from those of glacier-fed lakes, and to divide the latter into those bordered by an actively calving glacier (ice-contact or proglacial lakes) and those located downstream (non-contact or distal lake) (Ashley *et al*. 1985).

Most of the material deposited in glacial lakes comes from sediment in suspension and bedload in glacial meltwater streams. Additional contributions may be derived from slope processes delivering sediment directly into the lake (slope wash, avalanching, debris torrents, for example), atmospheric precipitation (including volcanic events), hydrochemical precipitation, biogenic activity, upwelling of material from groundwater flow, and resuspension from bottom current activity.

Deltas form where a meltwater stream or the glacier itself enters a lake. Sudden flow expansion causes an abrupt decrease in stream velocity and competence, which in turn results in rapid deposition of coarser material (see GLACIDELTAIC (GLACIODELTAIC)). At ice margins, other glacilacustrine sediments are also deposited, including subaquatic flow tills, formed by gravity deposits from debris-rich glacier ice standing in a lake. Icebergs can release particles either individually, dropstones, or in conical debris mounds on the lake floor.

The bulk of sediment discharge into a glacial lake comes from glacial streams during the spring and summer-melt period. Concentrations of suspended sediments are highly variable, with values ranging from a few $mg\,l^{-1}$ to $g\,l^{-1}$ in extreme cases. Density differences between inflowing stream waters and glacial lakes result largely from differences in suspended sediment concentrations and temperature. With strong density contrasts, the incoming stream water will maintain its integrity and flow into the lake as a discrete density current, either as an overflow (if its density is less than the lake water), an interflow (strong thermal stratification may result in flow along the thermocline), or underflow (if the inflowing water is more dense). The highly seasonal and weather-dependent nature of glacial-river discharge, temperature and suspended sediment concentration, together with the normal seasonal evolution of lake thermal structure, result in changing and often complex mixing and sedimentation patterns at different stages of the year. The resulting rhythmic deposition of sediments is a signature of many ice-contact and distal glacier-fed lakes.

Turbid underflows, high-density currents generated by underflowing sediment laden river water which produce quasi-continuous currents, and episodic surge-type currents formed by subaqueous slumping (velocities may range up $1\,m\,s^{-1}$) both transfer suspended sediment and a large quantity of bedload directly to deeper parts of the lake floor. A distinctive suite of graded deposits often characterized by ripple-drift and cross-laminations result. In lakes where underflows dominate, the descent of turbidity currents down the basin sides may inhibit deposition and in places may cause active erosion.

When and where underflow activity is not evident, such as during winter months or due to fluctuations in discharge, settling of particles takes place from sediment suspended in the water column. The resulting deposits, normally only a few millimetres to centimetres thick, grade from silty-clay at the base to fine clay at the top. They often terminate abruptly with a sharp contact, due to a new underflow influx of coarse material. In the most distal areas of glacial lakes, variations in sediment inflow may be sufficiently damped to give rise to homogeneous clays.

A signature of many glacial lake floor deposits are 'rhythmites'. These are pairs (couplets), composed of light-coloured, silt layers, representing spring flood or storm deposits, and dark, clay layers, with higher organic content, representing quiet deposition under winter ice. The contact between the two layers may be gradational, but more often it is sharp. Multiple laminations may occur within the more proximal silt layers, reflecting short-term fluctuations (hours and days) in sediment influx and dispersal. Local factors, load and volume of the meltwater stream, the depth of the lake and relief of its floor, the strength of the currents and the distance from the point of entry into the lake, affect the thickness of the couplets (Menounos 2002). A recurring theme in discussions of rhythmites is their periodicity. De Geer (1912) introduced the term 'varves' to describe annual couplets. Non-annual glacilacustrine rhythmites can be formed from sudden fluctuations of discharge and sediment load, sometimes from OUTBURST FLOODs, cold and warm spells of a non-annual nature, episodic slope activity, or periodic action of storms stirring up lake waters (Sturm 1979). Great care must be taken to establish a reliable, independent chronology for rhythmites, especially if they are to be used as a geochronological tool (Brauer and Negendank 2002). Varved glacilacustrine deposits have been used to interpret high-resolution records of paleoenvironmental conditions; notably, climate, glacial activity, mineralogy of drainage areas, and changes in water level, temperature and trophic state (see, for example, Karlen 1976; Leonard 1986).

Shoreline processes in glacial lakes are similar to those in lakes in other environments. Lake waters standing at particular levels create strandlines with wave-eroded scarps, beaches, small deltas and terraces. Coarse-washed gravel, cobble and boulder deposits may accumulate where waves erode older glacigenic (e.g. till) deposits. In glacial lakes, wave activity may be inhibited for part of the year by the presence of ice cover. The effects of movement of ice cover against the shore, due either to thermal expansion or wind coupling, produce small ice-push features, which may reach heights up to a few metres. The inclination of glacial strandlines (commonly 1 or $2\,\mathrm{m\,km^{-1}}$) gives important insight into the rebound and tilting since ice unloaded certain areas.

Water levels in many ice-contact lakes fluctuate widely, a consequence of meltwater filling and subsequent ice-dam collapse and drainage. This has important effects on lake-bottom sediments, through scouring and slumping, as well as ancillary effects due to changing wave base, iceberg grounding and adjustments of distribution patterns of suspended sediments.

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SEE ALSO: glacier; glacideltaic; glacifluvial

CATHERINE SOUCH

GLACIMARINE (GLACIOMARINE)

Here, the term glacimarine is preferred to other alternatives (glacial marine, glacial-marine and glaciomarine) because etymologically, words with Latin roots are joined with an 'i'. An inclusive definition is also preferred here, where the term is taken to encompass the environment, processes and deposits including landforms, sedimentary systems, stratigraphy and life forms. Glacimarine systems include a combination of glacial and marine processes that produce a penecontemporaneous mixture of primarily siliciclastic and biogenic sedimentary deposits. Terrestrial sediment is introduced by ice rafting and rainout of debris (IRD); by fluvial transport feeding turbid overflow plumes with eventual suspension settling of particles; by mass flows and rockfalls from ice contact and shoreline subaerial systems; by aeolian transport with eventual settling through water (perhaps via sea ice); and by shoreline and shelf processes such as longshore transport. Glacimarine settings occur within a range of climatic (and glaciological) regimes from polar, to subpolar, to cool temperate, and encompass fjords and nearshore areas, continental shelves and the deep sea.

Grounding-line depositional systems are formed at the contact of a glacier with the seafloor. These deposits take the form of a bank (morainal bank (less-favoured alternatives: moraine, submarine moraine and moraine bank)), a fan (grounding-line fan (less-favoured alternatives: subwash fan, glacimarine fan and submarine ice-contact fan)) and a wedge (grounding-line or grounding-zone wedge and trough-mouth fan (less-favoured alternatives: till tongue, till delta, subglacial delta and diamict apron)). Grounding-line systems include a mixture of facies: till, glacimarine diamicton (stratified or massive), gravelly mud (laminated, e.g. cyclopels and cyclopsams, or massive), poorly or well-sorted sand and gravel (stratified or massive), and interlaminated sand and mud (e.g. turbidites) (see Further reading for details).

Till with various modifiers (e.g. waterlain till and paratill), has been used as the genetic term for glacimarine diamicts; however, till is best reserved for deposition directly from glaciers without modification such as by flowage or by currents during rainout. Thus glacimarine diamict is preferred, and if genetic interpretations are possible, then such terms as debris flow deposit (or debrite), or rainout diamict (produced from ice rafting), or ice-keel turbate (produced by keels of icebergs or sea ice) may be used. Specific environmental terms for the rainout diamicts may be shelfstone diamict or bergstone diamict, depending on whether their debris source is an ice shelf or icebergs, respectively.

Beyond these ice-contact systems that extend two to several kilometres from a grounding line, are ice-proximal (to ~10 km from a grounding line) and ice-distal glacimarine systems (to thousands of km from a grounding line, e.g. Heinrich layers). These distances are relative to grounding lines and may be within an ice shelf or an iceberg zone. The main glacial components are from IRD, suspension settling and, more rarely, wind transport. Deposits are either gravelly mud or diamict depending on relative accumulation rates of IRD and matrix sediment, which often is from meltwater streams. The matrix is stratified under higher current strengths and sedimentation rates or under continuous ice cover (which control the degree of bioturbation), and is otherwise massive. However, extremely high sedimentation rates with few bottom currents can produce massive deposits.

Ice rafting occurs via three forms of ice and, if possible, recognizing the distinction is useful, such as by: ice shelves and floating glacier-tongues – ISRD, icebergs – IBRD, and sea ice – SIRD. Sea-ice rafting perhaps should be excluded from glacimarine systems because it is not strictly glacial and may occur under non-glacial conditions. However, often distinguishing SIRD from other IRD is impossible and thus it is commonly included in the glacimarine system. Ensuring that particle rafting is not by tree roots or kelp holdfasts is also important. A French term, *glaciel* has been suggested for sediment containing IBRD and SIRD, but is not commonly used.

Biogenic components in glacimarine deposits become more common with lower terrestrial sediment flux and meltwater; that is, either with distance from a glacier terminus or in colder climates. The geologically significant components include various macrofossils and microfossils, but diatoms commonly dominate and often form diatomaceous mudstone and diatomaceous ooze (diatomite). Marine productivity and diversity may depend on sea-ice extent, thickness and seasonal longevity, on sea water temperature and salinity changes, and on current up-welling (including polynya); thus these records contain high-resolution climate signals.

Morphologically significant forms produced in glacimarine settings include: fjords, cross-shelf troughs (or submarine troughs or sea valleys), inter-ice-stream ridges, mega-lineations (large-scale forms like flutes), flutes, grounding-line systems, iceberg and sea-ice scours, ploughs, or wallows, and striated boulder pavements.

The glacimarine environment includes sedimentary systems and processes that are typical of lower latitude settings, such as deltas, fan deltas, estuaries, tidal flats, linear sandy shorelines, shelves and deep water systems that commonly may include indicators of ice action described above. It includes lags, erosional surfaces, hiatuses and condensed sections produced from reworking by marine currents, from sediment starvation under large ice shelves or in ice distal areas during glacial retreat, and from isostatic rebound. By analogy with terrestrial glacial outwash and lacustrine systems, paraglacial marine settings occur where glaciers terminate on land, but their products of glacial rock flour accumulate as marine mud, perhaps including SIRD.

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ROSS D. POWELL

GLACIPRESSURE (GLACIOPRESSURE)

The term, 'Glaciopressure' was introduced by Panizza (1973) to indicate the pressure of ice on the narrow part of a valley, which is particularly intense at the confluence of glacial tongues in the areas affected by Pleistocene glacier advances. It caused rock deformations in correspondence with surfaces of structural discontinuity, like strata, fissures, etc., favouring the formation of sliding surfaces. In fact, some landslides which took place in the late Glacial and Post Glacial were observed in the Alps, and particularly in the Dolomite region: they were triggered by a tensional discharge following the loss of pressure previously exerted on the rocky slopes by two or more glaciers merging in a valley narrow. Even if the fall of large slope portions can directly affect human settlements or obstruct a whole valley, with the negative resulting consequences, the extremely high risk degree assumed by this type of phenomenon is purely theoretical. Indeed, the long time span from the withdrawal of the glacial network to the present has practically produced the total exhaustion of these events.

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MARIO PANIZZA

GLACIS D'ÉROSION

Glacis d'érosion are a form of PEDIMENT, a gently inclined slope of transportation and/or erosion that truncates rock and connects eroding slopes or scarps to areas of sediment deposition at lower levels (Oberlander 1989). Two fundamental types of pediment are recognized by Oberlander (1989): glacis d'érosion, which truncate softer rocks adjacent to a more resistant upland; and 'true' pediments, where there is no change in lithology between upland and pediment.

The name glacis d'érosion is derived from the work of French geomorphologists who studied examples of these landforms on the northern margin of the Sahara Desert, where they are particularly well developed on the flanks of the Atlas Mountains (Coque 1960). These landforms truncate weak materials such as poorly indurated Tertiary sediments, and tend to be veneered by alluvial gravels, indicating the importance of fluvial processes in their creation (Dresch 1957). The glacis piedmonts of the Atlas Mountains have a distinctive morphology, consisting of a series of coalescing flattened cones whose apices occur where stream channels debouch from upland drainage basins. The glacis long profiles range from nearly rectilinear to concave; the latter form having a slope of about 10 degrees at the top, dropping to about 3 degrees or less at the base.

Glacis d'érosion often exhibit multiple levels, or terraces, which can be traced back into the upland drainage basin where they form river terraces (see TERRACE, RIVER) (Plakht *et al.* 2000). These forms, known as stepped or nested glacis (Coque and Jauzein 1967), are formed as older glacis are incised by stream channels, which then form a younger glacis at a lower level, the new glacis being inset within the older glacis. The resulting landform appears similar in form to a telescopically segmented ALLUVIAL FAN, leading some workers to suggest that both landforms result from similar responses to environmental change (White 1991). The slope profiles of stepped glacis tend to converge downslope, the gradients decreasing from oldest to youngest. The oldest glacis are often only present as narrow residual ridges or outlying mesas, as planation of lower glacis have progressively removed upper glacis. Coque and Jauzein (1967) suggest that the number of glacis in Tunisia decreases systematically towards the south (Plate 53). Five glacis are present around Tunis and the High Steppe, south of Gafsa there are only four, the highest being present only as a few outliers.

Glacis d'érosion are thought to be erosional surfaces formed by fluvial action, cutting sequences of rocks that are easily eroded relative to the rocks of the adjacent upland. Supporting evidence for this fluvial model comes from the fact that stepped glacis are often paired on either side of contemporary channels rather like paired river terraces, and the fact that glacis are almost always covered with a layer of alluvium. This alluvial cover can be up to 15 m thick, though it rarely exceeds 10 m. Lower (younger) glacis tend to have thinner alluvial cover, and the alluvium tends to decrease in thickness towards the distal edge of the piedmont. The alluvium tends to be poorly sorted at the top of the glacis, becoming

Plate 53 A series of stepped glacis d'érosion developed on the southern flank of Djebel Sehib, southern Tunisia

better sorted downslope. In more arid areas, the alluvial cover of the glacis is frequently cemented by calcium carbonate or gypsum, forming an indurated DURICRUST. The role of duricrust in the formation of glacis is uncertain, although it may play an important part in preservation of older glacis.

Coque (1962) ascribes the formation of glacis in North Africa to slope retreat resulting from climate change; specifically a succession of humid and arid phases known to have affected the Sahara Desert during the Quaternary. He envisages a sequence of lateral planation during a humid phase, when moisture was sufficient to produce enough debris to balance the carrying capacity of streams, allowing them to erode laterally. This was followed by incision during an arid phase, when downcutting was promoted by lower sediment load in the streams. A return to more humid conditions resulted in renewed lateral planation at a lower level, forming a new glacis inset within the one above. This model is a gross oversimplification of the COMPLEX RESPONSE which river channels are now known to exhibit in response to environmental changes, but it is still generally believed that changes in the fluvial system resulting from climate changes are the basic trigger for formation of stepped glacis. The fact that the stepped glacis converge downslope indicates that changes in base level are unlikely to be involved in their formation. The widespread distribution of glacis across areas of different structural setting also rules out NEOTECTONICS as a major factor in their formation.

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SEE ALSO: alluvial fan; desert geomorphology; pediment

KEVIN WHITE

GLACITECTONIC CAVITY

Glacitectonic cavities are narrow and subhorizontal openings generated in bedrock by traction under a flowing GLACIER (Schroeder *et al.* 1986). The parallel walls take the form of irregular chevrons that follow the vertical joint pattern, while the roofs follow stratification planes. In some cases, well-compacted lodgement till forms the roof. The irregular floor of galleries is usually covered by debris issued from localized roof or walls failures.

Located less than 20 m below the surface, glacitectonic cavities can be hundreds of metres long, but typically less than 3 m wide and less than 10 m high. As their presence is only revealed by chance, from excavation work or local roof failures, they constitute hazardous constraints in

URBAN GEOMORPHOLOGY, especially in eastern Canada (Schroeder 1991).

Glacitectonic cavities are found below planar topographies, within sub-horizontal limestone or thinly bedded shale. Movement and weight of a flowing inlandsis, possibly aided by dissolution along the stratification planes, allows rock sheets to slide one on the other, leading to the spreading apart of vertical joints and to the creation of glacitectonic voids.

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JACQUES SCHROEDER

GLACITECTONICS (GLACIOTECTONICS)

Glacitectonic deformation may be defined as 'the structural deformation as a direct result of glacier movement or loading' (INQUA Work Group on Glacier Tectonics 1988). This term was first introduced by Slater (1926), and re-examined by Banham (1975). This topic has been studied a great deal in recent years, and a number of collections of papers on the subject have been published (Aber 1993; Warren and Croot 1994) and an online bibliography (http://www.emporia.edu/s/www/earthsci/biblio/biblio.htm). Additionally, recent textbooks on glacial geology include detailed sections on glacitectonic deformation (Benn and Evans 1997; van der Wateren 1995).

However, prior to the 1980s, glacitectonic deformation was thought to be a rare phenomenon, and was studied as a distinct field in glacial sedimentology. This view was first challenged when Boulton and Jones (1979) suggested that a significant proportion of glacier motion occurred not in the ice, but in a saturated weak deforming layer beneath the ice. These results showed how glacitectonic deformation was an integral part of the glacial environment, and not an unusual occurrence.

There are two types of glacitectonic deformation formed by the action of a moving glacier (Figure 76):

(a) *Proglacial deformation* which takes place at the glacier margin and is characterized by pure shear and compressional tectonics, i.e. open folds, thrusts and nappes. This results in the formation of push moraines;

(b) *Subglacial deformation* which takes place beneath the glacier and is characterized by simple shear and extensional tectonics, i.e. attenuated folds, boudins and augens, and results in the formation of deformation till and/or flutes and drumlins.

Similar styles of deformation can also occur within the ice itself (Hart 1998) as well as within permafrost (Astakhov *et al.* 1996).

Proglacial glacial tectonic structures have been relatively well studied because of their accessibility. In fact, the large number of studies of these features led many workers in the past to consider only proglacial deformation in discussions of glacitectonics. In contrast, subglacial deformation has had the least study because of the logistical

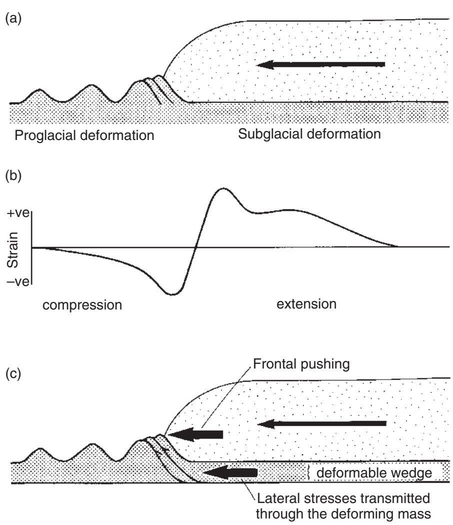

Figure 76 (a) schematic diagram showing the positions of subglacial and proglacial glacitectonic deformation; (b) the theoretical pattern of longitudinal strain; (c) schematic diagram of the forces producing proglacial deformation: frontal pushing and compressive stresses transmitted through a subglacial deformable wedge (after Hart 1998) problems involved in subglacial process studies, but the number of studies has dramatically risen in the past ten years.

Additionally, deformation also occurs within the glacial environment as a result of gravitation instabilities associated with stagnant ice and is known as dead-ice tectonics. Typical features include ice-collapse structures in an outwash plain, debris-flow mobilizations of till, and instability of subglacial sediments to produce 'crevasse infill' structures. These features are not glacitectonic structures *sensu stricto*, but may reflect the presence of saturated till in the subglacial environment, and so may be associated with subglacial deformation.

Proglacial glacitectonic deformation

Proglacial glacitectonic deformation is generally characterized by large-scale compressional folds and thrusts. The usual result of the proglacial deformation processes is to produce a topographic ridge transverse to the ice margin called a push moraine. There is often a basin up-glacier from which the material of the ridge has been removed. However, they do not always have a topographic expression. Many proglacial structures have been subsequently overridden by ice and so have become incorporated into drift deposits and have little or no topographic expression.

Push moraines are very common and can occur on scales ranging in height from 0.5 to 50 m, and in length from 1 m to several kilometres. Push moraines are associated with both contemporary glaciers and Quaternary glaciations (as well as pre-Quaternary glaciations) (see reference list). A recent review of push moraines is by Bennett (2001).

It has been argued by numerous workers that proglacial deformation can be modelled as thin-skin thrust tectonics, and the processes involved in the formation of push moraines are similar to mountain building in hard rock tectonic terrains. Using the work of Hubbert and Rubey (1959), many researchers have argued that glacitectonic nappes move along incompetent rock units or planes of weakness due to high pore-water pressures.

Although there are many processes associated with proglacial glacitectonic deformation, they can be generally divided into two types:

1 *'Foreland only' deformation* Where there is no deforming bed present, deformation may only take place in the foreland by the deformation of pre-existing sediments. This may typically include sandur sediments in terrestrial environments and shallow marine or fjord sediments in marine environments.
2 *'Deformable wedge' deformation* Where there is a deforming bed present, the subglacial and proglacial environment can be modelled as a deformable wedge which deforms by gravity spreading driven by the ice (Figure 76c). Deformation occurs due to both down-ice thrusting of the glacier into the foreland, as well as the horizontal component of the glacier's effective pressure (normal pressure minus pore-water pressure) transmitted through the subglacial layer into the foreland.

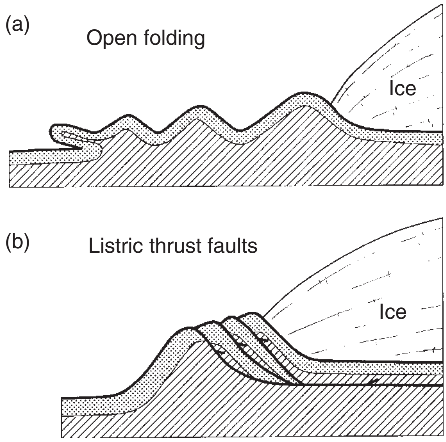

Figure 77 Schematic diagrams of different proglacial push moraines: (a) open folding; (b) listric thrust folding (after Hart and Boulton 1991)

Deformation of sediments (in both styles of push moraine) range from ductile (open folding) to brittle (thrust faulting and thrust nappes) (Figure 77). These styles of deformation reflect both the competancy of the material and increasing longitudinal compression from the simple folding to more complex nappe structures. Deformation structures are also found at the base of thrust faults and nappes, with tectonic breccias associated with brittle deformation and shear zones formed associated with ductile deformation.

Subglacial glacitectonic deformation

Although there are fewer studies of the subglacial environment due to its inaccessibility, there is still a considerable body of literature on subglacial deformation. Early descriptions of deformation structures in till included folds, laminations and blocks, boudins or rafts of soft sediments; such till was called 'deformation till'.

Subglacial deformation can occur beneath warm-based glaciers, when meltwater released from the glacier bed cannot easily escape from the system, so that pore-water pressures in the subglacial sediments build up and sediment strength is reduced:

$$\tau = (P_i - P_w) \tan \varphi$$

where τ is basal friction, P_i is ice overburden pressure, P_w is pore-water pressure and tan φ is the angle of internal friction (Coulomb's Law).

STUDY METHODS

In recent years subglacial deformation has been studied by three ways: (1) *in situ* process studies; (2) geophysical techniques; and (3) sedimentology. These methods have been discussed in detail in Hart and Rose (2001).

In situ subglacial process studies consist of the monitoring of the subglacial environment by inserting instruments into the subglacial bed via hot water drilled boreholes. This is a relatively simple technique and has been used on about ten modern glaciers including Breiðamerkurjökull (Iceland), Ice Stream B (Antarctica), Trapridge Glacier (Canada), Black Rapids Glacier (Alaska), Storglaciären (Sweden) and Bakaninbreen (Svalbard). These studies reveal the average thickness of the deforming layer is 0.5 m and indicate that deformation does occur beneath most of these glaciers. However, the importance of the effects on basal motion due to subglacial deformation ranges between 100 per cent at Black Rapids Glacier, Alaska to 13 per cent at Ice Stream B, Antarctica. Although the reason for the difference is not yet known, it has been suggested that the granulometry of the subglacial sediment may account for the difference in behaviour within the deforming layer. The glaciers with coarse-grained till appear more likely to have a higher percentage of basal motion due to sediment deformation, whilst those with more clay-rich lithologies may have only very thin deforming layers.

In addition, the presence of a deforming bed over large areas has been identified by seismic investigations in Antarctica, in particular beneath Ice Stream B and the Rutford Ice Stream.

However, most studies of subglacial deformation have been based on sedimentological studies from both modern and Quaternary glacial sequences. Most researchers have argued that the subglacial deforming layer behaves as a shear zone, which is a narrow band of high overall ductile shear located between sub-parallel walls. This deformation results in three features that will be discussed briefly below: deformation till, deformation structures and subglacial bedforms.

Deformation till

Hart and Boulton (1991) have argued that the resultant till from subglacial deformation is deformation (or deforming bed) till, which is a primary till formed from a combination of both deformation and deposition. It forms by direct melt-out of debris from the ice above, advection from till up-glacier, and changes in the thickness of the deforming layer. Where layers of deformation till are accreted on one another this is known as constructional deformation. In contrast, where the deforming layer thickens (due to changes in effective pressure, or large rafts of bedrock being thrusted into the shear zone), this is known as excavational deformation.

Deformation structures

Features typical of shear zones include: folds, boudins, augens, rotated clasts and tectonic laminations (Plate 54). The latter form from the attenuation of perturbations of the base of the deforming layer, producing ungraded laminations. However, these features will only be visible if they are formed from the mixing of sediments with different lithologies or competencies, under relatively low to medium simple shear. At very high shear strains these tectonic features can become homogenized and so macro-scale structures may not be visible. Instead criteria such as a specific till fabric (low strength associated with a thick deforming layer, high strength associated with a constrained deforming layer), or specific micromorphology structures (evidence of rotation or shears) may be used as a distinguishing criterion.

Subglacial bedforms

It has also been argued that subglacial streamlined bedforms (lineations, flutes and drumlins) are a product of subglacial deformation (Boulton

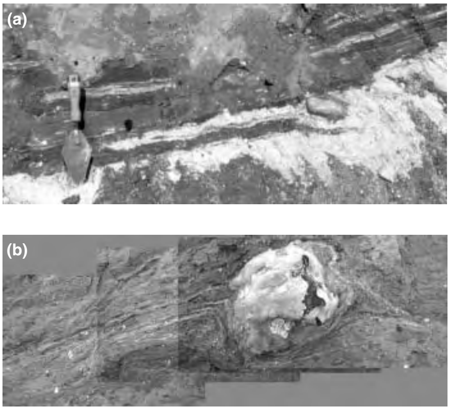

Plate 54 Examples of subglacial deformation: (a) chalk being attenuated to form tectonic laminations, West Runton, Norfolk; (b) chalk laminations flowing around on obstacle (flint clast), Weybourne, Norfolk (photographs by Kirk Martinez)

1987). These features form due to the presence of more competent masses (or cores) within the deforming layer which act as obstacles to flow. Where the core of the drumlin is weak then deformation structures will be seen, but these are relatively rare, and instead most drumlins have a competent core and a carapace composed of deformation till.

Using this sedimentological data, a number of authors have argued for wide spread subglacial deformation beneath the Pleistocene glaciers where the ice moved over the unconsolidated rocks of the European (and British) and Laurentide ice sheets.

Conclusions

Glacitectonic deformation is a fundamental process in glacier behaviour and a key component in the proglacial and subglacial sediment/ice deposition, erosion and transport system. There are very few modern glaciers that do not show evidence for proglacial deformation at their margins, and subglacial *in situ* process studies have revealed that subglacial deformation is also a common process. In addition, studies of Quaternary sediments demonstrate that such processes were also widespread in the past.

As a result, any study of the glacial environment needs to take glacitectonics into consideration, and future research needs to focus on the geotechnical properties of till to further understand the links between sediment behaviour and ice dynamics.

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JANE K. HART

GLINT

A marked gemorphological line dividing (the Canadian and the Baltic) SHIELDs from neighbouring stable platforms (the Great Plains and the Russian Plains, respectively) in the northern hemisphere. It is manifest in an ESCARPMENT which extends hundreds of kilometres and rises 20–100 m above the shield. The front of this escarpment is called a glint line. Pre-Pleistocene DENUDATION and, more significantly, differential scouring of expanding ICE SHEETs during the Pleistocene is responsible for glint formation. The Palaeozoic (Ordovician, Silurian) limestones, dolomites and sandstones of the tableland are more resistant to glacial erosion than the weathered Precambrian igneous and metamorphic rocks of the shield. Forward pushing ice was temporarily halted by the escarpment and thus allowed deep scouring at its base. After ice retreat meltwater accumulated in the depressions and glint lakes originated.

Glint is an Estonian term of Germanic origin. Once it denoted cliffs along coasts. The Baltic–Ladoga Glint extends from the islands of Estonia along the southern coast of the Gulf of Finland. Lake Ladoga and Lake Oniega occupy the depressions. Although the term is not in use in Canada and the United States, the glint line is also present there. Some major lakes, including the Great Bear, the Great Slave, Lake Winnipeg and the Great Lakes are all glint lakes, a subclass of ice-scoured lakes. Niagara Falls is the best known example of waterfalls along the glint line.

DÉNES LÓCZY

GLOBAL GEOMORPHOLOGY

The term global geomorphology embodies the notions of studying landform development at large spatial and temporal scales, of emphasizing global variations in landforms and geomorphic processes, of investigating the interactions between the land surface and other components of the Earth system, and of appreciating the particular combination of conditions for landform genesis on Earth compared with the other solid planetary bodies of the Solar System.

In focusing attention on large-scale phenomena and change over long periods of time, global geomorphology is concerned primarily either with the development of very large individual landform features, such as an entire mountain range, or with the assemblage of smaller individual landforms making up whole landscapes. At these large spatial and temporal scales the internal geomorphic processes of volcanism and tectonics generally become more significant in relation to surface geomorphic processes. A further consequence of looking at the macroscale is that short-term measurements of geomorphic processes on their own provide relatively little insight into the nature and rates of processes responsible for landscape genesis. Although numerical models (see MODELS) have been developed to investigate large-scale landscape change, data relevant to testing such models generally have to pertain to periods of thousands to millions of years, rather than the timescale of years or decades of modern process measurements.

A methodological consequence of these long timescales is that the approach to global geomorphology is predominantly historical where the emphasis is on explaining the conditions and processes responsible for development over time of a single major landform, or a regional or larger scale landscape. This contrasts with the dominance of the functional approach in small-scale surface process geomorphology where the main interest is understanding the adjustment of form to process over short periods of time.

Another distinction between global geomorphology and smaller scale approaches to landform analysis is the frame of reference that must often be employed. At the small scale it is usually sufficient to know local slope gradients and height differences, such as between interfluves and river channels, rather than absolute changes in elevation with respect to sea level. In global geomorphology, however, constraining changes in absolute elevation of the land surface above sea level is required in order to relate changes in regional topography through time to rates of crustal uplift and rates of denudation.

Historical context

A global approach to geomorphology is not new. An important theme in the study of the landforms up to the nineteenth century was the attempt to understand the origin and history of the surface of the Earth as a whole. For instance, one of the elements of Charles Lyell's concept of UNIFORMITARIANISM, was the notion of a steady-state Earth in which uplift in some areas was 'balanced' by subsidence in others, and where changes in the location of uplift and subsidence occurred over time, but the overall form of the Earth's surface did not change substantially.

Lyell's idea of regions of crustal uplift and subsidence was taken up by Charles Darwin who, during the earlier part of his career dominated by writings on geological subjects, sought to develop a global synthesis relating uplift of the continents to processes such as volcanism and mountain building. More specifically, Darwin adopted Lyell's notion of regions of subsidence in developing his own theory of coral atoll formation in which he envisaged coral reefs growing upwards from the substrate of volcanic islands grouped in broad regions of what he inferred was subsiding ocean crust. In developing his coral reef theory, Darwin provided an object lesson in historical methodology applied to understanding landform development by suggesting how spatial patterns of a range of related landforms – in this case, volcanic islands, barrier reefs, fringing reefs and atolls – could, with careful observation and reasoning, be viewed as representing stages in the development of a single landform through time.

This strategy was taken up and extended by William Morris Davis who, in his CYCLE OF EROSION, sought to develop a general evolutionary scheme for landscape development, where the form of a landscape was seen as a product of the rock structures present and the surface geomorphic processes operating, but predominantly as a function of stage of development. Although acknowledging complications arising from variations in climatic conditions, and particularly from intermittent crustal uplift, Davis's evolutionary model depended heavily on the reality of distinctive landscapes being created at different stages of the cycle of erosion. In its simplest form this involved rapid uplift from close to sea level of a low relief land surface, its progressive incision by river systems creating maximum local relief, and then an extended period of interfluve lowering with respect to valley bottoms until a low relief surface close to sea level, or PENEPLAIN, was restored.

Although often heavily dependent on particular interpretations of landscape features, and thus far less secure than Darwin's earlier exemplary treatment of coral atoll formation, the evolutionary approach advocated by Davis became the dominant strategy of global geomorphology through the first half of the twentieth century, at least amongst geomorphologists in Britain and North America. In Germany a different model of landscape change was developed by Walther Penck who emphasized the importance of the interplay between external erosional processes and internal tectonic processes causing uplift. Although more in sympathy with modern approaches to understanding large-scale landscape development, Penck's more complex approach to landform analysis never achieved the influence of the simple version of Davis's evolutionary model.

The idea of the development over millions of years of extensive, low relief erosion surfaces graded to sea level led to the development of DENUDATION CHRONOLOGY, a method of landscape analysis in which low relief erosion surfaces at different elevations were interpreted in terms of falls in base level resulting either from eustatic sea-level change (global sea-level fall), or from tectonic uplift of the land surface. Throughout the mid-twentieth century much emphasis was placed in correlating supposed remnants of particular surfaces considered to have resulted from specific uplift events. Taken to its ultimate extent, the correlation of such erosion surface remnants across the continents was seen as potentially being a chronological replacement for stratigraphy where the sedimentary record was absent or incomplete. The fullest development of denudation chronology as a methodological basis for global geomorphology is perhaps to be found in the work of Lester King who interpreted flights of low relief surfaces across different continents as representing synchronous episodes of continental uplift of global extent.

The decline in interest in global geomorphology and the corresponding move towards studies of small-scale surface process geomorphology from the 1950s occurred for two main reasons. One was the wholly inadequate dating control that was usually available for the denudation chronologies presented, the other a lack of understanding of the tectonic and surface geomorphic processes that could create erosion surfaces graded to sea level, and then subsequently preserve remnants of them when base level fell.

Renewed interest in global geomorphology

Although the revolution in Earth sciences arising from PLATE TECTONICS might have been expected to reinstate interest in global geomorphology, little attention was paid by most geomorphologists to this integrative global-scale model when it was formulated in the late 1960s and early 1970s, presumably because the focus by that time was on quantitative approaches to small-scale surface geomorphic processes. A renewed concern with global geomorphology is really only evident from the 1980s, and it occurred for a number of reasons. One was the growing availability of satellite remote sensing imagery which made evident the large-scale components of the Earth's landforms. Although initially used primarily to explore regional and subcontinental-scale landform associations, by the 1990s satellite data was being used to create digital elevation models (DEM) of the land surface at horizontal resolutions down to a few metres. This added to the growing number of digital topographic data sets being created from national archives of topographic data. By 2001 the Shuttle Radar Topography Mission had collected high resolution radar-based elevation data covering the Earth's surface between latitudes ~60°S to 60°N.

At the same time that Earth-orbiting satellites were providing images of terrestrial landscapes, there was a flood of remote sensing imagery from planetary missions, such as the Viking missions to Mars and the Voyager missions to the outer planets in the 1970s, the Magellan mission to Venus in the 1980s and the Mars Global Surveyor which provided high resolution images. Understanding of the tectonics, volcanism, surface processes and climatic history of these planetary bodies relied heavily on the interpretation of their landforms, where possible by comparisons with supposed terrestrial analogues (for instance, the outflow channels on Mars which were seen to have many similarities to the landscapes of catastrophic flooding in the Channeled Scabland of eastern Washington, USA). At the same time, the scale of landforms seen in planetary imagery pointed to the insights to be gained by studying terrestrial forms with a similar global perspective (see EXTRATERRESTRIAL GEOMORPHOLOGY).

Another important reason for increased interest in global geomorphology was the development of the computing capability necessary to numerically model regional-scale landscapes over geological time spans. Although small catchment/channel slope-scale surface process models have been developed by geomorphologists and hydrologists since the 1960s, numerical models of regional-scale landscape evolution incorporating tectonic deformation and isostasy, as well as surface processes, have been under active development only since the late 1980s.

Constraining such models requires data on denudation rates for time spans of millions of years relevant to long-term landscape development. The increasing availability of such information from the 1980s as a result of new geochronological methods and data sources is another reason for the revived interest in global geomorphology. Hydrocarbon exploration along continental margins has provided a wealth of data on rates of sediment deposition from which denudation rates on the adjacent hinterland can be estimated, at least where the sediment source area and its changes over time can be constrained. More important, however, has been the application of thermochronological techniques to infer denudational histories and denudation rates. A range of low-temperature techniques such as $^{39}Ar/^{40}Ar$ dating, fission-track thermochronology (see FISSION TRACK ANALYSIS) and helium thermochronology can now provide information on the cooling history of rocks in the upper few kilometres of the Earth's crust. This information on the timing and rate of cooling can be converted into estimates of denudation rates averaged over periods of millions of years since it is the progressive stripping of crust by denudation that is largely responsible for shallow crustal cooling. These data provide information on broad regional patterns of denudation, but they can now be related to more local denudation rates by coupling with data from cosmogenic isotope analysis (see COSMOGENIC DATING) which provides denudation rates over timescales of thousands to hundreds of thousands of years.

Key issues in global geomorphology

The most obvious issue in global geomorphology is to understand the gross variations in the Earth's continental topography and how this topography has changed over time. Why, for instance, is 82 per cent of the world's land surface over 4,000 m above sea level concentrated in the Tibetan Plateau? And what is the origin of the large area of anomalously high topography extending across southern Africa and into the adjacent Atlantic Ocean. Answers to these questions require an understanding of the interaction of internal and external processes over periods of millions of years. Crucial to answering such questions are data on changes on the elevation of the land surface over time, since the timing of the uplift of the Tibetan Plateau, for instance, is key to understanding the cause of such uplift.

Unfortunately, constraining such surface uplift has proved very difficult, not least because denudation in uplifted terrain tends to remove evidence that would be indicative of prior elevations. However, various techniques have been developed to infer past elevations in addition to the obvious strategy of using shoreline or shallow marine deposits where present. These include inferring temperature (and therefore indirectly elevation) change from the characteristics of fossil leaves on the basis that specific plant types have particular temperature tolerances and that surface uplift will elevate fossils into cooler climatic zones. This approach has been used to infer surface uplift in mountain ranges such as the Himalayas, but it requires detailed information on global and regional climatic changes which would also produce vertical shifts in climatic zones. Another approach is to use the elevation-dependent fractionation of oxygen in precipitation across mountain ranges which can be incorporated into carbonate sediments, but of considerable potential is basalt vesicle ratio analysis. This technique uses the effect of atmospheric pressure of the relative size of gas bubbles at the top and bottom of individual lava flows to infer the atmospheric pressure, and hence elevation, at the time of eruption. Notwithstanding these and other techniques, constraining changes over time in the absolute elevation of the land surface remains a problematic but fundamental issue in global geomorphology.

Another important issue is the coupling of onshore and offshore records of denudation and deposition. The growth in offshore hydrocarbon exploration along continental margins since the 1970s has greatly expanded our knowledge of their depositional history, but it has also raised the question of what controls the supply of sediment from the adjacent continental hinterland. Answering this question requires information not just on the mobilization and transport of sediment from onshore to offshore but also on tectonic mechanisms and the isostatic response to changes in crustal loading as mass is transferred offshore. Although largely irrelevant to small-scale surface process geomorphology, ISOSTASY assumes a critical role in global geomorphology since, at these larger spatial and temporal scales, flexure of the lithosphere in response to denudational unloading can have important effects on the mode of landscape development.

A further key theme in global geomorphology is the coupling between internal and external processes. Although the influence of tectonic mechanisms on surface processes through the construction of relief has been long understood, the way in which spatial variations in denudation rates can affect patterns of tectonic deformation was only fully appreciated in the 1990s. This is evident in the commonly found strike-parallel pairing of metamorphic facies in mountain ranges as a result of higher rates of denudation (and therefore greater depths of exposure) on the wetter, windward side compared with the drier, leeward side. Modelling of patterns of crustal deformation as a result of spatial variations in denudation rates has further emphasized the two-way interaction between surface and internal geomorphic processes.

The role of the land surface in interactions between tectonics and climate has also received attention in attempts to understand the long-term geological controls over the concentration of atmospheric carbon dioxide and hence, through the greenhouse effect, global climate. The key process here is the weathering of silicate minerals, a reaction which draws down CO^2 from the atmosphere. As global topography and relief changes as a result of interactions between tectonics, climate and landscape development, the global rate of CO^2 drawdown would be expected to vary, although the operation of these interactions are far from fully understood.

Finally, comparative planetary geomorphology provides the key perspective for global geomorphology. Looking at landscape development on other planetary bodies shifts our perspective from viewing terrestrial landforms as 'normal', and emphasizes that the Earth's landforms have arisen from a particular combination of its size, its composition, its distance from the sun, the composition and density of its atmosphere, and its age. The great majority of planetary bodies have surfaces dominated by impact craters, making impact cratering the dominant geomorphic process in the Solar System (although most occurred in the first 500 to 600 Ma from the birth of the Solar System around 4.5 Ga ago). The critical factor for Earth is the surface temperatures that it experiences which encompass the range over which water can exist as a solid, a liquid and a gas. This enables Earth to have an active hydrological cycle which is key to many geomorphological processes. Also Earth's size and composition means that it has a high enough internal temperature to melt rock and therefore permit volcanism and the convection that helps power plate tectonics. Earlier in its history, Mars also probably experienced short-lived episodes when there was a fairly active hydrological cycle including oceans; this is the main period of the channel formation on Mars. By contrast, the high surface temperatures on Venus, largely resulting from an intense greenhouse effect associated with a dense CO^2–rich atmosphere, has prevented the existence of liquid water; thus its surface is dominated by the effects of volcanism and impact cratering.

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MIKE SUMMERFIELD

GLOBAL WARMING

There is now a widespread appreciation that the build-up of greenhouse gases in the atmosphere (carbon dioxide, methane, nitrous oxide, CFCs, etc.) will create an enhanced greenhouse effect that will cause global warming. Details of the degree of warming that will occur and of the associated changes in other climatic variables are provided in the reports of the Intergovernmental Panel on Climate Change (2001). If such changes occur over coming decades certain landscapes and geomorphological processes will be modified (Table 22).

Some landscapes, 'geomorphological hot spots', will be especially sensitive because they are located in zones where it is forecast that climate will change to an above average degree. In the high latitudes of Canada or Russia the degree of warming may be three or four times greater than the global average. It may also be the case with respect to some critical areas where particularly substantial changes in precipitation may result from global warming. For example, various scenarios suggest the High Plains of the United States of America will become markedly drier. Other landscapes will be highly sensitive because certain landscape-forming processes are so closely controlled by climatic conditions. If such landscapes are close to particular climatic thresholds then quite modest amounts of climatic change can flip them from one state to another. In this entry attention will be paid to some of these hot spots.

Tundra and permafrost terrains

High latitude tundra and PERMAFROST terrains may be regarded as one of these sensitive zones. They are likely to undergo especially substantial temperature change. In addition, the condition of permafrost is particularly closely controlled by temperature conditions. By definition it cannot occur where mean annual temperatures are positive, and the latitudinal limits of different types of permafrost can be related to varying degrees of negative temperatures. Thus the equatorward limit of continuous permafrost may approximate to the $-5\,°C$ isotherm and the equatorward limit of discontinuous or sporadic permafrost to the $-2\,°C$ isotherm. It is likely that the latitudinal limits of permafrost will be displaced polewards by 100 to 250 km for every 1 °C rise in mean annual temperature. The quickest loss of permafrost would occur in terrains underlain by surface material with low ice contents. The slowest response would be in ice-rich materials, which require more heat to thaw. Snow or the presence of thick, insulating organic layers (i.e. peat) might also buffer the effects of increased surface temperatures in some areas.

There is historical evidence that permafrost can degrade speedily. For instance, during the warm 'optimum' of the Holocene (*c*.6,000 years ago) the southern limit of discontinuous permafrost in the Russian Arctic was up to 600 km north of its present position (Koster 1994). Similarly, researchers have demonstrated that along the Mackenzie Highway (Canada), between 1962 and 1988, the southern fringe of the discontinuous zone had moved north by about 120 km in response to an increase over the same period of 1 °C mean annual temperature (Kwong and Tau 1994).

Table 22 Some geomorphologic consequences of global warming

Hydrologic
Increased evapotranspiration loss
Increased percentage of precipitation as rainfall at expense of winter snowfall
Increased precipitation as snowfall in very high latitudes
Possible increased risk of cyclones (greater spread, frequency and intensity)
Changes in state of peatbogs and wetlands
Less vegetational use of water because of increased CO_2 effect on stomatal closure
Vegetational controls
Major changes in latitudinal extent of biomes
Reduction in boreal forest, increase in grassland, etc.
Major changes in altitudinal distribution of vegetation types (*c*.500 m for 3 °C)
Growth enhancement by CO_2 fertilization
Cryospheric
Permafrost, decay, thermokarst, increased thickness of active layer, instability of slopes, river banks, and shorelines
Changes in glacier and ice-sheet rates of ablation and accumulation
Sea-ice melting
Coastal
Inundation of low-lying areas (including wetlands, deltas, reefs, lagoons, etc.)
Accelerated coast recession (particularly of sandy beaches)
Changes in rate of reef growth
Spread of mangrove swamp
Aeolian
Increased dust storm activity and dune movement in areas of moisture deficit
Soil erosion
Changes in response to changes in land use, fires, natural vegetation cover, rainfall erosivity, etc.
Changes resulting from soil erodibility modification (e.g. sodium and organic contents)
Subsidence
Desiccation of clays under conditions of summer drought

Woo *et al*. (1992) made certain predictions based on the assumption that a greenhouse warming of 4–5 °C causes a spatially uniform increase in surface temperature of the same magnitude over northern Canada. They suggested that permafrost in over half of what is now the discontinuous zone could be eliminated, that the boundary between continuous and discontinuous permafrost might shift northwards by hundreds of kilometres and that a warmer climate could ultimately eliminate continuous permafrost from the whole of the mainland of North America, restricting its presence only to the Arctic Archipelago.

In areas where rapid permafrost melting occurs, the consequences will be legion. They include ground subsidence (THERMOKARST), increased erosion of shorelines and riverbanks, and an increase in debris flow activity and other forms of slope instability.

High latitude areas may also be particularly susceptible to changes in precipitation and runoff. Areas which are currently very dry, because the air is so cold, may become moister as warmer winters cause more snow to fall, thereby creating a likelihood of increased summer runoff. In somewhat warmer environments, where substantial winter snowfall occurs, there might be a tendency in a warmer world for a decrease in the proportion of winter precipitation that falls as snow. There would thus be greater winter rainfall and runoff, but less overall precipitation to enter snowpacks to be held over until spring snowmelt. This in turn would have adverse consequences both for late spring and summer runoff levels in rivers and for soil moisture levels. Other factors may also modify runoff. For example, as permafrost thaws, groundwater recharge may increase and surface runoff decrease.

Glaciers and ice sheets

Glaciers and ice sheets will be highly susceptible to a rise in temperature. Although there has been considerable debate as to whether or not polar ice caps might respond catastrophically to global warming because of an increase in ablation, accelerated melting of tidewater snouts, the cliffing of termini by a rising sea level, or the removal of the buttressing effects of ice shelves as they melt (Huybrechts *et al.* 1990), it is probably valley glaciers in alpine situations which will respond most quickly and markedly to climatic warming. Such glaciers are highly responsive, as is made evident by their frequent and rapid fluctuations during the Neoglacials of the Holocene. Although topographic controls and changes in precipitation and cloudiness are significant controls of glacier state, it is highly likely that most alpine glaciers will show increasing rates of retreat in a warmer world. Indeed, given the rates of retreat (20–70 cm year) experienced in many mountainous areas in response to the warming episode since the 1880s, it is probable that many glaciers will disappear altogether, from areas as diverse as the Highlands of East Africa or the Southern Alps of New Zealand.

Desert margins

The history of desert margins indicates that in the past they too have been sensitive to environmental change. This in turn suggests that they are likely to be susceptible to future environmental changes. Thus closed depressions have fluctuated repeatedly from being dry and saline to being full and fresh. Valley bottoms and hillsides have alternated between cut-and-fill, and dunefields have at some times been mobile and at other times stable (Forman *et al.* 2001). Many dry regions will suffer large diminutions in runoff (Arnell 1999), with annual totals likely to be reduced by over 60 per cent. Indeed Shiklomanov (1999) has suggested that in arid and semi-arid areas an increase in mean annual temperature by 1 to 2 °C and a 10 per cent decrease in precipitation could reduce annual river runoff by up to 40–70 per cent.

In the case of closed depressions, the dating of high water levels in lakes in the tropics and subtropics shows that many of them have had a complex history during the Holocene and that their water levels have varied considerably. High levels were a feature of the Saharan region around 8,000 years ago, a time when global temperatures were probably slightly greater than today. Very large numbers of freshwater deposits date from this time, even in the dry heart of the Sahara. Some stream courses (e.g. the Wadi Howar) were active.

In the case of river and slope systems, they too have fluctuated between phases of stability or alluviation and phases of erosion and incision. Even over the last century or so the valley systems of the American south-west, called ARROYOS, have experienced trenching and filling episodes in response to climatic and other stimuli (e.g. land use change). Of particular importance have been changes in the amount and intensity of precipitation. Crucial in this respect is the response of vegetation cover to rainfall events, for in semi-arid areas it is not only highly dependent on moisture availability but also controls the erodibility of the ground surface (Elliot *et al.* 1999).

Changes in precipitation and evapotranspiration rates also have a marked impact on aeolian environments and processes. Rates of deflation, sand and dust entrainment and dune formation are closely related to soil moisture conditions and vegetation cover. Areas that are at present marginal with respect to aeolian processes will be particularly susceptible, and this has been made evident, for example, through recent studies of the semi-arid portions of the United States (e.g. the High Plains). Repeatedly throughout the Holocene they have flipped from a state of vegetated stability to states of drought-induced surface instability. Thermoluminescent and optical dates have made evident their sensitivity to quite minor perturbations. Geomorphologists, using the output from General Circulation Models (GCMs), combined with a dune mobility index which incorporates wind strength and the ratio of mean annual precipitation to potential evapotranspiration, have shown that with global warming, sand dunes and sandsheets on the Great Plains are likely to become reactivated over a significant part of the region, particularly if the frequencies of wind speeds above the threshold velocity for sand movement were to increase by even a moderate amount (Muhs and Maat 1993). The same applies to dust storm generation in the Great Plains and the Canadian Prairies, where the application of GCMs shows that conditions comparable to the devastating dust-bowl years of the 1930s are likely to be experienced.

Tropical coastlines

Tropical coastlines are a further very sensitive environment with respect to future climatic change. This is for three main reasons: the relationship between tropical cyclone activity and the sea-surface temperature (SST), the temperature tolerances of coral reefs, and the effects that temperature change and sea-level rise have upon mangrove swamps.

Tropical cyclones are important agents of geomorphological change. They scour out river channels, deposit debris fans, cause slope failures, build up or break down coastal barriers, transform the nature of some coral islands (either building them up or erasing them), and change the turbidity and salinity of lagoons. Were their frequency, intensity and geographical spread to change it would have significant implications. It is not, however, entirely clear just how much these important characteristics will change. Intuitively one would expect cyclone activity to become more frequent, intense and extensive if sea-surface temperatures were to rise, because SST is a clear control of where they develop. Indeed, there is a threshold at about 26.5–27.0 °C. However, the Intergovernmental Panel and some individual scientists are far from convinced that global warming will invariably stimulate cyclone activity.

Coral reefs may be sensitive to warming, partly because of the role that cyclones play in their evolution, partly because their growth can be retarded or accelerated because of changes in SSTs, and partly because their existence is so closely related to sea level.

In the 1980s there were widespread fears that if rates of sea-level rise were high (perhaps 2 to 3 m or more by 2100) then coral reefs would be unable to keep up and submergence of whole atolls might occur. Particular concern was expressed about the potential fate of Pacific Island groups, and of the Maldives in the Indian Ocean. However, with the reduced expectations for the degree of sea-level rise that may occur, there has arisen a belief that coral reefs may survive and even prosper with moderate rates of sea-level rise. As is the case with marshes and other wetlands, reefs are dynamic features that may be able to respond adequately to sea-level rise. It is also important to realize that their condition depends on factors other than the rate of submergence.

Increased sea-surface temperatures could have deleterious consequences for corals which are near their thermal maximum. Most coral species cannot tolerate temperatures greater than about 30 °C and even a rise in seawater temperature of 1–2 °C could adversely affect many shallow-water coral species. Increased temperatures in recent years have been identified as a cause of widespread coral bleaching (loss of symbiotic zooxanthellae). Those corals stressed by temperature or pollution might well find it more difficult to cope with rapidly rising sea levels than would healthy coral. Moreover, it is possible that increased ultraviolet radiation because of ozone layer depletion could aggravate bleaching and mortality caused by global warming. Various studies suggest that coral bleaching was a widespread feature in the warm years of the 1980s and 1990s (Goreau and Hayes 1994).

However, Kinsey and Hopley (1991) believe that few of the reefs in the world are so close to the limits of temperature tolerance that they are likely to fail to adapt satisfactorily to an increase in ocean temperature of 1–2 °C, provided that there are not very many more short-term temperature deviations. Indeed, in general they believe that reef growth will be stimulated by the rising sea levels of a warmer world, and they predict that reef productivity could double in the next hundred years from around 900 to 1,800 million tonnes per year. They do, however, point to a range of subsidiary factors that could serve to diminish the increase in productivity: increased cloud cover in a warmer world could reduce calcification because of reduced rates of photosynthesis; increased rainfall levels and hurricane activity could cause storm damage and freshwater kills; and a drop in seawater pH might adversely affect calcification.

However, reef accretion is not the sole response of reefs to sea-level rise, for reef tops are frequently surmounted by small islands (cays and motus) composed of clastic debris. Such islands might be very susceptible to sea-level rise. On the other hand, were warmer seas to produce more storms, then the deposition of large amounts of very coarse debris could in some circumstances lead to their enhanced development. However, the situation is complex, and in some cases potential vertical reef accretion could be reduced by storm attack. One also needs to consider changes in tropical storm frequency as well as changes in tropical storm magnitude, for high storm frequencies might change the relative importance of corals and calcareous algae (Spencer 1994).

Other coastlines

There are other coastlines that will also be substantially modified by sea-level rise resulting from global warming. These include sandy beaches, miscellaneous types of saltmarsh and areas of land subsidence.

Sandy beaches are held to be sensitive because of the so-called BRUUN RULE (Bruun 1962, see Plate 55). This predicts future rates of coastal erosion in response to rising sea level. Bruun envisaged a profile of equilibrium in which the volume of material removed during shoreline retreat is transferred onto the adjacent shoreface/inner shelf, thus maintaining the original bottom profile and nearshore shallow conditions. With a rise in sea level additional sediment has to be added to the below-water portion of the beach profile. One source of such material is beach erosion, and estimates of beach erosion of *c.*100 m for every 1 m rise in sea level have been postulated. However, although the concept is intuitively appealing, it is also difficult to confirm or quantify without precise bathymetric surveys and integration of complex nearshore profiles over a long period of time. Moreover, an appreciable time-lag may occur in shoreline response which is highly dependent upon local storm frequency. Furthermore, the model is essentially a two-dimensional one in which the role of longshore sediment movement is not considered. It is also assumed that no substantial offshore leakage of sediment occurs. Accurate determination of sediment budgets in three dimensions is still replete with problems. Whatever the problems of modelling, however, sandy beaches will tend to disappear from locations where they are already narrow and backed by high ground or swamp and marsh, but will probably tend to persist where they can retreat across wide beach ridge plains.

Saltmarshes, including MANGROVE SWAMPs, are potentially highly vulnerable to sea-level rise, particularly where sea defences and other barriers prevent the landward migration of marshes as sea-level rises. However, saltmarshes are dynamic features and in some situations may well be able to cope, even with quite rapid rises of sea level. Indeed, some important sediment trapping plants may extend their range in response to warming. Such plants include mangroves (e.g. in New Zealand) and also *Spartina anglica* (e.g. in northern Europe). They would tend to lead to an acceleration in marsh accretion.

One way of attempting to predict the effects of increasing rates of sea-level rise is to study those areas where the rates of sea-level rise are currently high because of subsidence. On the coast of south-east England, where rise occurs at a rate of 5 mm per year, saltmarshes appear to cope. Sediments eroded from the outer edge appear to contribute to the sediments which are accreted on the inner marsh surface. Moreover, UK saltmarshes have current rates of accretion that are the same order of magnitude as, or greater than, the predicted rates of sea-level rise.

Plate 55 The main railway line between France and Spain near Barcelona. Note the severe erosion of the coastline, and the abandoned track in the foreground. Sandy beaches of this type will be especially sensitive to the effects of accelerated sea-level rise associated with global warming

Reed (1990) suggests that saltmarshes in riverine settings may receive sufficient inputs of sediment that they are able to accrete sufficiently rapidly to keep pace with projected rises of sea level. Likewise, some vegetation associations, e.g. *Spartina* swards, may be relatively more effective than others at encouraging accretion, and organic matter accumulation may itself be significant in promoting vertical build-up of some marsh surfaces. For marshes that are dependent upon inorganic sediment accretion, increased storm activity and beach erosion which might be associated with the greenhouse effect could conceivably mobilize sufficient sediments in coastal areas to increase their sediment supply.

One particular type of marsh that may be affected by anthropogenically accelerated sea-level rise is the mangrove swamp. Mangroves may respond rather differently to other marshes because their main plants are relatively long-lived trees and shrubs. This means that the speed of zonation change will be less. The degree of disruption is likely to be greatest in microtidal areas, where any rise in sea level represents a larger proportion of the total tidal range than in macrotidal areas. However, the setting of mangrove swamps will be very important in determining how they respond. River-dominated systems with large allochthonous sediment supply will have faster rates of shoreline progradation and deltaic plain accretion and so may be able to keep pace with relatively rapid rates of sea-level rise. By contrast, in reef settings in which sedimentation is primarily autochthonous, mangrove surfaces are less likely to be able to keep up with sea-level rises (Ellison and Stoddart 1990).

The ability of mangrove propagules to take root and become established in intertidal areas subjected to a higher mean sea level is in part dependent on species. In general the larger propagule species (e.g. *Rhizophora* spp) can become established in rather deeper water than can the smaller (e.g. *Avicennia* spp). The latter has aerial roots which project only vertically above tidal muds for short distances.

Mangrove colonization and migration would also be influenced by salinity conditions so that any speculations about mangrove response to sea-level rise must also incorporate allowance for change in rainfall and freshwater runoff.

In arid areas, such as the Middle East, great lengths of coastline are fringed by low level salt-plains (SABKHA). These features are generally regarded as equilibrium forms that are produced by depositional processes (e.g. alluvial siltation, aeolian inputs, evaporite formation, faecal pellet deposition) and planation processes (e.g. wind erosion and storm surge effects). They tend to occur at or about high tide level. Because of the range of depositional processes involved in their development they might be able to adjust to a rising sea level but quantitative data on present and past rates of accretion are sparse.

A crucial issue with all types of wetlands is the nature of the hinterland. Under natural conditions many marshes and swamps are backed by low-lying estuarine and alluvial land which could be displaced if a rising sea level were to drive the marshes landward. However, in many parts of the world sea defences, bunds and other structures have been built at the inner margins and these will prevent colonization of the hinterland. Experiments are now being conducted to see whether saltmarsh development can be promoted by the deliberate breaching of sea defences.

One final type of sensitive coastal environment is that where coastal submergence is taking place. The combination of local submergence with global sea-level rise will make these coasts especially prone to inundation. Some areas are subject to natural subsidence as a result of sediment loading on the crust (e.g. deltas) or because of tectonic processes, but some key areas are subjected to accelerated (i.e. anthropogenic) subsidence. This is brought about primarily by mining of ground water or hydrocarbons and can be especially serious in the case of coastal mega-cities, portions of which are either close to or beneath current sea level (e.g. Bangkok and Tokyo).

Although there may still be uncertainties about whether global warming will occur and about the various impacts of such warming should it occur, and although the degree of climatic and sea-level change that is being postulated might at first sight appear relatively modest, it would be wrong to be complacent about the potential geomorphological impacts brought about by global warming. Our knowledge of how geomorphological systems have reacted to the climatic fluctuations of the Holocene, and our knowledge of the intimate relationships between some geomorphological processes and climatic conditions, both lead us to the conclusion that some environments will respond in a manner that will be substantial in degree and which will have numerous consequences for human occupation of these environments.

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A.S. GOUDIE

GOLDICH WEATHERING SERIES

The types and proportions of various minerals in a weathering profile are usually quite different from the original bedrock. Some minerals seem to survive more or less unaltered even after being subjected to prolonged WEATHERING, while others decompose more rapidly. In many weathering studies, the silicate minerals, the primary constituents of igneous and metamorphic rocks, are arranged into an order of susceptibility to chemical weathering. The most commonly cited order was first proposed by S.S. Goldich (1938), based on a detailed study of the mineralogic changes of granitoid rocks during weathering (Figure 78).

Goldich (1938) concluded that minerals which form at high temperatures and pressures (olivine, amphiboles, pyroxenes, calcium plagioclase), and hence are the first to precipitate, are markedly less stable and weather much more quickly than minerals which crystallize at lower temperatures and pressures (sodium plagioclase, potassium feldspar, micas and quartz) (Figure 78). This sequence is the reverse of BOWEN'S REACTION SERIES, which ranks minerals in their order of crystallization from a melt.

CHEMICAL WEATHERING reactions are with the cations that bind the silica structural units together. Thus the relative strength between the oxygen and cations in each mineral and the structure of the bonding are both significant. The isolated Si-O tetrahedra in olivine are the least stable in weathering; while quartz, which is completely formed of interlocking silica tetrahedra with no intervening cations, is the most stable. If muscovite and the plagioclases are disregarded, the order of the Goldich weathering series coincides with the classification of silicate structures, based on increasing Si–O–Si bonds from zero in olivine to four in quartz.

Although Goldich's series has widespread applications and usually works well, local exceptions have been documented.

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SEE ALSO: Bowen's reaction series; chemical weathering

CATHERINE SOUCH

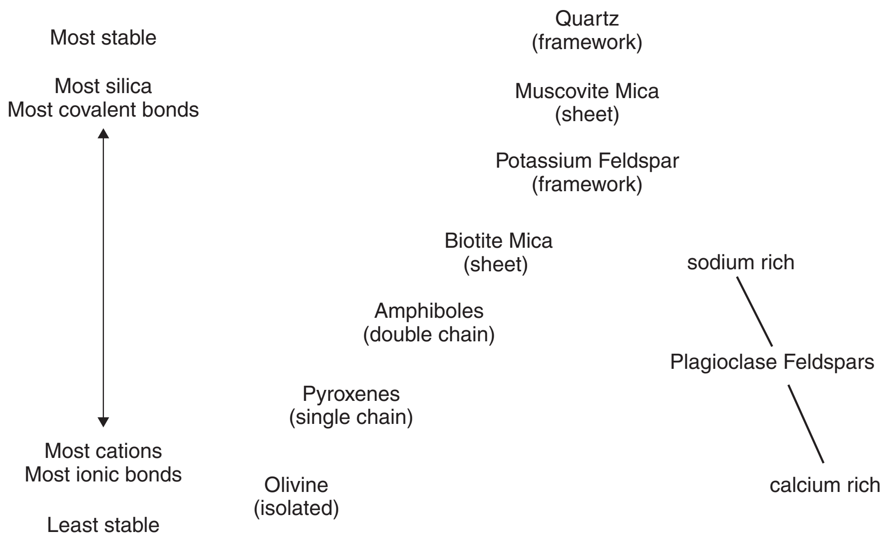

Figure 78 Goldich weathering series

GORGE AND RAVINE

Gorges, which may be hundreds of metres deep, are caused either by incision of a river against an uplifting landmass, the superimposition of a channel across resistant rock, the outburst of floodwaters across a landscape, or by the headward retreat of a KNICKPOINT or WATERFALL (Rashleigh 1935; Derricourt 1976; Tinkler *et al.* 1994; van der Beek *et al.* 2001). Ravines are much smaller gashes (the order of metres to tens of metres wide and deep) cut into the weak bedrock, or frequently into superficial sediments such as glacial deposits or deeply weathered horizons. The term ravine is frequently used in the context of soil erosion and land degradation, and a ravine, or ravine network, will have steep, weakly consolidated side slopes, flat channel bottoms characterized by a heavy sediment load, and a clear break of slope with the surface above. Present academic literature seems to find the technical use of the word limited to south and east Asia (Raj *et al.* 1999), otherwise it is used as a synonym for gully. It may occur as the generic part of a place name: Yamuna Ravine in India, Elk Ravine, New Hampshire, USA.

Neither of the terms *gorge* nor *ravine* is well defined in the literature, although a Gorge (e.g. the Three Rivers Gorge in western China) is generally understood to be typical of rivers of larger sizes. The word 'ravine' tends to imply a small deeply incised channel in a low-order drainage basin. Both terms imply a river deeply incised below the surrounding landscape, local slope processes being unable to reduce the side slopes at the same rate that the river is incising into the terrain. Thus there is often little sensitivity to the local topography. Large, deep gorges require mechanically strong country rock, although the typically steep valley slopes may still be susceptible to failure by rock fall and rockslides. Gorges are frequently found in areas where drainage is antecedent upon actively growing fold systems such as the Himalayan ranges, or where it is superimposed (superposed) upon more resistant rocks from weaker cover rocks.

In exceptionally large river systems the term gorge usually refers to the deeply incised, and often scarcely accessible inner gorge (Kelsey 1988). Bedrock channels often contain an inner channel, where bedload transport rates are highest, and erosion is enhanced. Because of the large variations in discharge which have occurred in many high latitude drainage basins during the Quaternary, it is unclear to what extent the inner channel of a river system carrying large glacial outflow discharges becomes the inner gorge of its non-glacial successor. Excavations for the Boulder Dam on the Black Canyon of the Colorado revealed an unsuspected inner gorge up to 25 m below flanking bedrock edges to the channel (Legget 1939: 322–323).

Catastrophic scale outburst floods of glacial stored waters are another mechanism, unsuspected until recent decades, for the formation of gorges (Baker 1978; O'Connor 1993; Rathburn 1993; Knudsen *et al*. 2001). Such floods may have been repeated many times during the Quaternary, their cumulative sum affect being what we now see.

Scheidegger *et al*. (1994) argue for strong structural control by large-scale regional joints and fault systems, in the geographical layout of large gorges. However overall trends in gorge orientation usually owe their origin to regional scale topographic trends (Baker 1978; Rathburn *et al*. 1993), to which structural control merely adds local detail. Subsequent river erosion may generate entrenched or incised meandering patterns unrelated to local or regional structure.

Buried gorges are not uncommon in glaciated terrains, many being found in the Great Lakes region of eastern Canada (Davis 1884; Karrow and Terasmae 1970; Greenhouse and Karrow 1994). The infill of permeable glacial sediments within a bedrock gorge often produces localities favourable for groundwater exploration (Farvolden 1969).

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KEITH J. TINKLER

GPS

The Global Positioning System (GPS) is a constellation of satellites developed by the US Department of Defense to provide precise positioning and navigation information. GPS receivers determine position through repeated measurements of digitally tagged radio signals from the satellites. Conceived for military purposes, the commercial applications for positioning information have blossomed. Analysts have suggested that the global GPS market is worth over US$16 billion (in 2002). Among the varied users of GPS are geomorphologists requiring geo-referenced positioning information for field terrain. However, the wide variety of systems available and the enormous range in cost and capability requires caution on the part of the user and an ability to assimilate a multitude of jargon and proprietary software.

Initially, the Department of Defense used a procedure termed Selective Availability to dither the precise time code and degrade the accuracy of the signal. This has been set to zero since May 2000, improving reliability and consistency though applications requiring sub-metre accuracy (and hence all fieldwork requiring elevation data) continue to require differential GPS (DGPS). DGPS relies on a static reference receiver at a known control point which logs bias errors over the same time period that another receiver (the 'rover') is occupying the points of interest. The measured errors are used to correct the rover position either by downloading and 'post-processing' the data, or by receiving corrections via radio telemetry (known as 'real time kinematic'). The control point can be operated by the user or be a commercial ground station broadcasting corrections. The reference frame for GPS output is the World Geodetic System 1984 (WGS-84), a geocentric system returning ellipsoid co-ordinates in latitude and longitude. Altitude is derived as elevation above the ellipsoid and some knowledge is needed to integrate GPS-derived height data with existing levelling data or to translate positions into a local datum. Fortunately there are many textbooks providing technical details (e.g. Hofmann-Wellenhof *et al.* 2001). Relatively few papers consider explicitly GPS applications in geomorphology (Cornelius *et al.* 1994; Fix and Burt 1995; Higgitt and Warburton 1999) but an increasing number make routine use of GPS as part of the data-gathering procedure. Four broad areas of application can be identified:

Rectification

Global referencing is essential in most geomorphological research. GPS can assist observations where detailed maps are lacking or it can be used for registering ground markers to analyse aerial photographs or remotely sensed imagery. This is useful for assessing change in sequential imagery such as the dynamics of land degradation (Gillieson *et al.* 1994). In terrain remote from conventional benchmarks, GPS can save much time in establishing the elevation of sample points.

Detailed topographic survey

The speed of GPS data capture offers scope for producing accurate digital elevation models (DEMs) of moderately sized field areas. The abundance of points in a GPS survey generates topographic attributes which can be used as input in hydrological models. A related commercial development is 'precision agriculture' where GPS receivers mounted to farm vehicles produce detailed information about spatial variations in crop yields or soil conditions. One consequence of precision, as highlighted by Wilson *et al.* (1998), is the recognition that calculated topographic attributes are sensitive to the resolution and distribution of survey points. By implication, estimation of topographic variables from a limited number of survey points may be prone to large errors. A dense network of GPS survey points around a catchment can provide a more enlightened summary about the statistical distribution of slope characteristics.

Measuring change in landforms

GPS is ideal for measuring sequential change in landform characteristics. Geologists have made extensive use of networks of high precision GPS for identifying ground movements associated with earthquakes and volcanic eruptions. Geomorphological applications are apparent in neotectonics and landslide research. Where budgets are more restrictive, repeat surveys provide similar information. This has been used to construct detailed maps of river channel change (Brasington *et al.* 2000). As the object of geomorphological study is usually inanimate, there are no parallels to ecological applications that examine animal behaviour. The methodology to determine grazing patterns by fitting ungulates with GPS collars might be adapted to keep track of students during field trips!

Geomorphological mapping

Where acquisition of elevation data are not critical, GPS can be an effective mapping tool. The outline of geomorphological features (e.g. the edge of river terraces or landslides) or point patterns (e.g. glacial erratics) can be obtained speedily in terrain where conventional surveying is impractical and the features cannot be determined sufficiently from aerial photography. GPS software has a facility to tag attribute information to the data and can be integrated into GIS. It should be remembered that the GPS receiver requires an unobstructed path to the satellites and hence mapping in mountainous terrain, urban environments or under forest cover can be problematic.

In each of the categories above, the speed and frequency of positioning is the essential difference enabling GPS to provide data that would be difficult or impossible to derive from conventional surveying methods. As such, GPS is not a technique producing completely new data but rather an application that improves accuracy and/or frequency of measurement coupled with efficient data-processing capability. The cost of GPS receivers spans at least two orders of magnitude. High precision GPS is not only expensive but requires a thorough understanding of surveying principles and the equipment can be bulky. Mapping grade GPS is highly portable and can be operated by a single user where safety considerations allow. The required accuracy should dictate the specification of GPS but its subsequent application in geomorphology is wide-ranging.

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DAVID HIGGITT

GRADE, CONCEPT OF

Since the end of the seventeenth century (Dury 1966; Chorley 2000), various engineers have been concerned both with the regulation of natural rivers and with the operation and construction of artificial channels. This required an interest in geometric stability or equilibrium (grade) which tended to be roughly constant or subjected to limited oscillation over a recognized period of time. Such stability could arise from some sort of balance between, for example, fluid shear stress and material resistance, or some equalization between those sedimentary processes (e.g. cut-and-fill) which control channel morphology. The concept entered mainstream geomorphology through G.K. Gilbert (1877), for whom the major geometrical evidence of the graded state was a smooth, concave-up river long profile. Grade was also accommodated within the Davisian cycle, and for Davis (1902) the elimination of breaks of slope was the hallmark of the graded condition. A major contribution to understanding the concept of grade was made by Mackin (1948) who defined a graded river as: 'one in which, over a period of years, slope and channel characteristics are delicately adjusted to provide, with available discharge . . . just the velocity required for the transportation of the load supplied from the drainage basin'.

In 1965 Schumm and Lichty introduced the concept of a time-span intermediate between the longer interval of 'cyclic time' and the shorter period of 'steady time'. They defined graded time as 'a short span of cyclic time during which a graded condition or dynamic equilibrium exists'. Later, Schumm (1977) saw a graded stream as 'a process-response system in steady-state equilibrium, and the equilibrium is maintained by self-regulation or negative feedback, which operates to counteract or reduce the effects of external change on the system so that it returns to an equilibrium condition'.

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A.S. GOUDIE

GRADED TIME

The most concise description of graded time is derived from Mackin (1948):

> A graded river is one in which, over a period of years, slope and channel characteristics are delicately adjusted to provide, with available discharge, just the velocity required for the transportation of the load supplied from the drainage basin. The graded stream is a system in equilibrium; its diagnostic characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change.

It is clear that 'graded time' is the time over which a stream is in balance in this way.

A second, less useful, term is the 'time to grade' (see GRADE, CONCEPT OF) or the time that it takes for a river to attain a graded condition. This is not a simple idea because the graded condition is not reached throughout all parts of a system at the same time. In rivers, for example, W.M. Davis (1902) stated that grade would be attained first in the lower reaches and then extend upstream. It would also be attained first in the most adjustable materials.

The term, graded time, therefore has a spatial dimension. Davis (1899) stated that when the trunk streams were graded the stage of early maturity had been reached, when the smaller headwaters were graded maturity was well advanced and when even the wet river rills and the waste mantle were graded the stage of old age had been attained.

The idea that once grade had been achieved the balance of forces, sediment loads and forms would remain adjusted even though the landscape was still being slowly lowered has always been an uncomfortable element of the geographical cycle.

The timeless aspect of the concept of grade does not sit happily within the timebound cyclical framework.

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DENYS BRUNSDEN

GRANITE GEOMORPHOLOGY

Granite terrains of the world, whether in lowland, upland or mountain settings, often have distinctive morphology, different from one typical for the surrounding country rock. Although it would probably be impossible to find a landform endemic for granite, many are most prominent if bedrock is granitic. Examples include boulders, TORS, INSELBERGS, BORNHARDTS, INTERMONTANE BASINS, and a range of microforms such as WEATHERING PITS or TAFONI (Twidale 1982). They usually form through selective bedrock weathering, either in subsurface (see DEEP WEATHERING) or at the topographic surface, followed by evacuation of the loose products of rock disintegration. However, there is no 'standard' granite landscape, as these can be significantly different, even if located adjacent to each other. Granite is known to support extensive plains of extreme flatness and, by contast, high-mountain, highly dissected terrains. In spite of widespread presence of a weathering mantle, bedrock frequently crops out at the topographic surface and tors and boulder fields are characteristic landmarks. Granite is typically, but by no means universally, more resistant to weathering and erosion than surrounding country rock, and therefore tends to form upland terrains and to support topographic steps.

Lithological and structural properties of granite, such as mineral composition, texture and joint density, which are often highly variable within a single granite intrusion, are the keys to understanding the selectivity of weathering and the prominence of many small- and medium-scale granite landforms.

Granites are usually fairly regularly jointed according to an orthogonal pattern, i.e. they are cut by three subsets of joints perpendicular to themselves, which delimit cuboid block compartments. As fractures guide movement of ground water through the rock mass, weathering acts most efficiently along joints and preferentially attacks the sides and edges of joint-bound cubes, which results in their progressive rounding and the typical multi-convex appearance of many granite landscapes. In the subsurface, weathering attack along joints transforms sharp-edged blocks into rounded core stones surrounded by a thoroughly disintegrated mass. Furthermore, because of variable joint density over short distances (< 10 m) significant differences in the intensity of rock disintegration may occur. Less fractured parts are left standing as rock pillars or castellated tors, whereas adjacent more closely jointed compartments are disintegrated into block rubble or GRUS. Evacuation of weathered material reveals a range of topographically negative features, common for granite areas. These include rock basins developed either at joint intersections or between master joints, and linear joint-guided valleys.

Many post-orogenic granites are typically very massive, with large-scale SHEETING joints being dominant. Joint spacing in such granites can be extremely wide, more than 10 m apart. In these areas topography usually follows the curvature of sheeting planes, bornhardts are common, and minor weathering features on rock surfaces often grow to gigantic dimensions.

Rock texture is equally important. Coarse variants of granite with abundant large phenocrysts of K-feldspar usually support a varied, rough relief, with big boulders, inselbergs, and intervening basins. The majority of domed inselbergs and bornhardts seems to be built of massive, coarse-grained granite. Likewise, minor features on rock surfaces are best developed within coarse granite. Finer variants tend to give rise to a more subdued topography, often with frequent angular tors.

Another factor important for the development of granite topography is mineralogical and chemical composition of the rock, including proportions between quartz and feldspar, between different types of feldspar, silica content, and proportions between potassium, sodium and calcium. Potassium-rich granites tend to be more resistant and therefore often form higher ground and give rise to spectacular inselberg landscapes, whereas granites with high plagioclase content typically underlie gently rolling terrains and low ground (Brook 1978; Pye *et al.* 1986).

In areas, where high precipitation and humidity levels favour deep weathering, subsurface decomposition of granite becomes crucial in the evolution of topography (Twidale 1982). Granite terrains, except for those in arid areas or in high mountains, usually carry a spatially extensive, thick mantle of weathering residuals. A very wide range of thicknesses has been reported, from only a few to as much as 200–300 m (Ollier 1984). There are different types of weathering mantles developing on granite, but rather shallow grus and more advanced geochemically, kaolinite-rich covers are most typical. This division likely reflects environmental conditions during weathering, including climatic conditions, their change through time and geomorphic stability of the surface. What both categories of granite weathering mantles have in common though, is the rough topography of the weathering mantle/bedrock interface (i.e. WEATHERING FRONT), attributable to the selectivity of deep weathering, and the usually sharp nature of this boundary. Therefore, stripping of the pre-weathered material often reveals complicated bedrock topography, with numerous low domes, isolated boulders and tor-like bedrock projections separated by basins and linear hollows. Indeed, many granite landscapes are interpreted to be the product of two-stage development, with the phase, or phases, of deep selective weathering followed by stripping and exposure of weathering front topography (Plate 56). The presence of bornhardts, tors and rounded boulders is occasionally used to infer the two-stage evolution, even if no remnants of any weathering mantle are left and no independent evidence exists that such ever existed. However, examples from areas with a long history of aridity such as the Namib Desert demonstrate that deep weathering is not a necessary precursor to the development of multi-convex granite topography, which primarily reflects structural control (Selby 1982).

The majority of detailed studies has concentrated on prominent medium-scale landforms such as boulder fields, tors, bornhardts and inselbergs, and pediments, or distinctive minor features of rock surfaces. Analyses of entire landform assemblages and their evolution through time are fewer. One of the attempts has been made by Thomas (1974) who distinguished multi-concave, multi-convex and stepped or

(a)


(b)


Plate 56 Granite landscapes share many of their characteristics regardless of the climatic zone in which they occur. Both the granite landscape of (a) the Erongo massif in arid Namibia and (b) the humid Estrela Mountains in central Portugal are dominated by massive domes, big rounded boulders scattered around and basins formed through selective joint-guided weathering

multistorey landscapes. In multi-concave terrains, topographic basins of various sizes are dominant features. Their occurrence may be related to either inliers of less resistant granite or to the occurrence of initially more jointed rock compartments (Thorp 1967; Johansson *et al.* 2001). Multi-convex terrains are those typified by closely spaced domes, or similar upstanding rock masses, so there is little space left for basins to develop. They are common in homogeneous, poorly jointed intrusions, where lines of structural weakness available for exploitation by weathering are few. Another type of multi-convex landscape is one dominated by low hills weathered throughout, possibly with a solid rock core.

Stepped landscapes are characterized by the presence of topographic scarps separating successive levels or 'storeys'. They typically occur in areas subjected to recent, but moderate uplift which was proceeding concurrently with weathering and stripping. Since the scarps are apparently not tectonically controlled, it is proposed that they form due to reduced rates of advance of the weathering front at progressively higher topographic levels, whereas their exact location reflects the occurrence of a more massive granite (Wahrhaftig 1965; Bremer 1993). In each of these landscape types, spatial patterns of individual landforms are largely controlled by lithology and structure.

Specific landform assemblages typify ring complexes, made of concentrically arranged intrusions of granite and other rocks, differing in mineralogy and texture, and intersected by dykes. Depending on the susceptibility of particular complex-forming rock units to weathering and erosion, a concentric pattern of uplands alternating with basins develops. Most resistant dykes form linear ridges, sculpted into jagged rock crests.

In addition, granite landscapes may take the form of a plain, either rock-cut or deeply weathered, as it is common in Australia, parts of Africa, or Scandinavia. Within strongly uplifted and highly dissected areas, a mountainous all-slope topography evolves (Twidale 1982). In both cases, structural control is less obvious and its influence surpassed by the high efficacy of planation or dissection.

Granite geomorphology has played an important part in CLIMATIC GEOMORPHOLOGY, and especially in the attempts to use specific landforms as indicators of specific climatic conditions. For instance, claims have been made that granite domes evolve in the humid tropics, boulder heaps are more typical for seasonally dry areas, whereas small-scale flutings indicate hot and humid conditions (Wilhelmy 1958). Moreover, the apparent durability of granite and its ability to withstand high compressive and tensile stresses have been used to support the claim that granite landforms, once formed under distinctive environmental conditions, may survive many subsequent environmental changes. In some Central European studies, minor granite landforms have been used to establish the chronology of denudation and environmental change since the mid-Tertiary. Increasing recognition of pervasive structural and lithological control on the evolution of granite landforms, as well as of the crucial role of subsurface weathering, have seriously undermined the basis of the climatic approach to granite geomorphology. At present, a consensus appears to have been reached that the evolution and appearance of granite landscapes are primarily controlled by structure, and similarities of structures explain why granite landform assemblages in contrasting geographical settings often look very much the same.

Many of the geomorphologically classic landforms and landscapes are underlain by granite. Examples include the tors of Dartmoor in southwest England, domes and U-shaped glacial valleys of the Yosemite National Park in Sierra Nevada, USA, sugar-loaf hills in Rio de Janeiro, African inselberg landscapes of Nigeria, Kenya and Namibia, fluted coastal outcrops in the Seychelles, and the Wave Rock in Western Australia.

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PIOTR MIGOŃ

GRANULAR DISINTEGRATION

Granular disintegration is the physical disintegration of rock into individual grains and rock crystals. The product of granular disintegration is usually coarse-grained, loose debris, which can be easily removed by erosive agents such as wind, water and gravity. This form of rock breakdown occurs commonly in coarse-grained rocks such as sandstone, dolerite and granite. Clay-rich rocks are thought to be particularly susceptible (Smith *et al.* 1994).

The surface grains and rock crystals which become detached may be unweathered and unaltered. They may also remain *in situ* but could be easily removed by light brushing with the hand. Where the product of granular disintegration remains *in situ* and accumulates, a gritty SAPROLITE is produced, known as GRUS. When loose material is removed, the fresh surface beneath may be pitted and uneven. The loose material may accumulate as a sandy deposit.

There are a number of mechanical and chemical mechanisms of granular disintegration and it is likely that the process can be attributed to several or all of these. It is equally likely that more than one of the mechanisms operates simultaneously in many cases:

1. *Solution of soluble cement* Sandstones cemented by soluble calcareous material are particularly susceptible to granular disintegration due to this mechanism.
2. *Stress induced by volumetric expansion* The growth of salt and ice crystals leads to a volumetric expansion. Under certain conditions, this can produce sufficient force to rupture the rock and this is most likely to occur at locations of weakness such as grain boundaries. There is ample evidence that rocks readily disintegrate in salt-rich environments due to salt crystallization (Evans 1970). Experimental work has also shown salt to be extremely effective in the

physical breakdown of rock (e.g. Goudie *et al.* 1970). Chemical weathering processes such as hydrolysis may involve expansion of minerals sufficient to produce crystal fracture.

3 *Release of residual stress* This is stress in rock due to primary crystallization or lithification. These stresses exist in a balanced state in unweathered rock. However, residual stresses can become unbalanced, and therefore released, by erosion, weathering and mass movement. The stresses generated can be large enough to cause crack propagation (e.g. Bock 1979).

4 *Water adsorption* Repeated wetting and drying may be responsible for the disintegration of fine-grained rocks such as mudstone (see SLAKING). Water molecules are absorbed onto mineral surfaces and may produce force sufficient to prise particles apart.

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DAWN T. NICHOLSON

GRAVEL-BED RIVER

An alluvial river in which the average diameter of bed materials exceeds 2 mm. An upper grain-size limit is seldom identified, but channels with beds predominantly composed of very large, essentially immobile boulders (> 256 mm) may be regarded as a distinct type or subcategory, especially if they exhibit step-pool morphology (see STEP-POOL SYSTEM). The primary distinction is with SAND-BED RIVERS (bed material 0.063–2 mm). Gravel-bed rivers dominate in upland and piedmont settings where the sediment supplied to the channel is coarse and poorly sorted. With distance downstream, bed materials become smaller (see DOWNSTREAM FINING) and an abrupt gravel–sand transition often terminates the gravel reach.

Although gravel-bed rivers transport significant quantities of sand, much of it in suspension, the proportion of BEDLOAD transport is apt to be higher than in sand-bed channels. Parker (in press) usefully defines a limit case wherein the median bed-particle size is greater than 25 mm and bedload transport dominates. Examination of such channels reveals that the boundary shear stresses generated by modest flows (for example, bankfull) are barely capable of moving median grain-sizes. In sharp contrast to sand-beds, gravel-bed channels are therefore characterized by hydraulic stresses that rarely exceed the entrainment thresholds of particles exposed at the bed surface, and large floods are required to generate significant sediment transport. Sediment yield is limited by the competence of flows to move the coarse load, rather than the availability of mobile sediments per se. This is a definitive characteristic of gravel-bed rivers, though in many environments vertical sorting of the bed material (ARMOURING) does significantly limit the availability of potentially mobile, subsurface sediments.

Close to the threshold for motion, the coarse armour layer remains intact and transport involves individual grain movements across its largely unbroken surface. Once rotated out of bed pockets by lift and drag forces, particles roll and bounce across the bed, intermittently stopping in stable positions from where they may be entrained again by instantaneous turbulent stresses. Particles cover relatively short distances, potentially falling into stable interstices or pockets in the armour layer. This marginal transport regime dominates during most floods, with the armour layer moderating sediment supply and grain velocities. However, as flow intensity increases, larger areas of the armour layer are breached, the number of particles in motion rises and, during exceptional floods, most of the bed may be mobile. Even during mass transport, flows are seldom sufficiently deep to form mobile bedforms of the geometry found in sand-bed channels (steep ripples and dunes), but low-amplitude forms known as gravel sheets are common, and their passage generates bedload pulses.

A number of small-scale bedforms are recognized in gravel-bed rivers and are important because they influence near-bed hydraulics and, like bed armour, moderate sediment supply and entrainment. Pebble clusters that form when large obstacle clasts distort the flow and impede the passage of other clasts, are repeating, streamlined features. Transverse ribs are regularly spaced, linear ridges of coarse clasts that form perpendicular to the flow under supercritical conditions. Stone cells are reticulate structures that may form where transverse ribs and pebble clusters intersect.

These micro-bedforms protrude above the bed surface and therefore contribute to overall flow resistance, as do channel-scale grain accumulations (bars and riffles) that retard the passage of water. However, in contrast to sand-bed channels where bedforms dominate boundary resistance, grain roughness is regarded as the dominant component in gravel-bed rivers (see ROUGHNESS).

The longitudinal profiles of gravel-bed rivers typically exhibit significant concavity that reflects adjustment to downstream fining and the associated reduction in competence required to transport a given load. Channel gradients therefore vary significantly from as much as 0.1 to as little as 0.001.

Cross-sectional form is determined by numerous variables in addition to bed-material size. Nevertheless, for a given discharge gravel-bed rivers do tend to be shallower than sand-bed channels and have higher width–depth ratios. This reflects the dominance of bedload transport and a lack of fine-grained, floodplain deposition, that together promote lateral instability and channel widening. Local variability of width and depth is exacerbated in gravel-bed rivers by well-developed riffle-pool sequences. Analogous bed topography is evident in some sand-bed channels, bedrock channels, and as step-pools in boulder-bed channels, but riffle-pools are best developed in gravelly channels with heterogeneous bed materials.

Gravel-bed rivers may be straight, meandering, anabranching (see ANABRANCHING AND ANASTOMOSING RIVER) or braided (see BRAIDED RIVER). To the extent that bedload transport dominates, and stabilizing, cohesive, floodplain sediments are lacking, gravel-bed channels tend to exhibit larger meander wavelengths and a propensity to wander or braid. Wandering is a type of anabranching that represents a transitional stage between meandering and braiding, with some sinuosity, low-level braiding and stable mid-channel islands. Wandering channels tend to have lower slopes and less abundant bedload than fully braided channels. Relative to sand-beds, gravel-bed channels require steeper slopes to generate full braiding.

Bed material size is a fundamental control of river form and function, and characteristic morphological and process attributes do justify the general distinction that is made by geomorphologists between gravel- and sand-bed rivers. The presence of a gravel–sand transition in many rivers and the widely reported deficiency of fluvial sediments in the range 1 to 4 mm – the so-called 'grain-size gap' – reinforce this binary categorization. However, all gravel-bed rivers contain sand, and many gravel-bed rivers transport larger volumes of sand than gravel. The sand is apparent to varying degrees as a patchy surface veneer (for example in pools) and in the subsurface matrix. This suggests that the two-fold, sand versus gravel, classification is rather simplistic and potentially limiting. It may obscure important attributes that are peculiar to channels containing particular mixtures of sand and gravel. Indeed, there is increasing evidence that understanding channel hydraulics, sediment transport and the formation of fluvial deposits in 'gravel-bed' rivers depends upon explicit recognition that bimodal gravel and sand mixtures often dominate the bed materials (e.g. Sambrook-Smith 1996).

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SEE ALSO: armouring; bedload; downstream fining; roughness; sand-bed river

STEPHEN RICE

GRÈZE LITÉE

Stratified TALUS deposits displaying well-developed cm-thick beds and composed of small angular clasts (Guillien 1951). They have also been referred to as éboulis ordonnés and stratified screes, though some authors find slight differences between these terms (mostly related to the slope gradient and the mean clast size). The most diagnostic features of grèzes litées are (1) the internal structure of the deposit, organized in parallel beds of around 10 to 25 cm thick, and (2) the small size of the clasts as a result of very frequent freezing and thawing cycles on a gelivable (frost susceptible) rock substratum.

The sedimentary structure shows alternating matrix-rich (matrix-supported) and openwork (clast-supported) beds. In many cases openwork beds show fining upward textures. In longitudinal sections the base of the matrix-rich beds is affected by festoons, with increasing size downslope, and even with the development of lobate fronts (Bertran *et al.* 1992). Undulations are relatively frequent in frontal sections. The presence of blocks within the clast-supported beds defines another more heterogeneous type of talus deposit called groizes litées. In carbonate-rich deposits the presence of carbonate cemented crusts is relatively frequent, as a result of percolation and water circulation.

Most authors consider that grèzes litées are better developed in limestone areas, at the foot of large vertical or sub-vertical cliffs. This is the case of Charentes (France), where the best examples have been studied, and many other localities in the Alps and the Pyrenees. However, they have also been described in crystalline, volcanic and metamorphic rock areas (for instance, in the Chilean Andes, Vosges, the French Central Massif and the Atlas in Morocco). These deposits can be up to 40 m thick. In all cases the grèzes litées are most frequent in middle latitudes, with a periglacial climate. In these regions the annual number of freezing and thawing cycles is high (even more than 200 days per year), providing the best conditions to break down the rocks and to accumulate large volumes of debris. Under these conditions the cliffs erode backward rapidly and they are partially fossilized by the grèzes litées.

Grèzes litées have been described in a wide range of slope gradients (between 5 and 35°), though, in general, gentler than in ordinary talus with non-stratified screes, thus excluding an origin based only on gravity. Most active grèzes litées are located on sunny aspects or in snow-free hillslopes. Pleistocene deposits, related to former cold-climate phases, are located in almost any aspect, depending not only on the altitude but also on local topography and wind direction (García-Ruiz *et al.* 2001).

Several hypotheses have been used to explain the development of grèzes litées. Tricart and Cailleux (1967) stressed the importance of in-mass transport (especially solifluction) accompanied by pipkrake (needle-ice) activity. Bertran *et al.* (1992) and Francou (1988) confirm the decisive role of continual burial of stone-banked sheets. This implies the existence of large solifluction sheets in which pipkrakes cause a vertical sorting of the material, displacing the clasts towards the surface. The movement of the front of the stone-banked sheets produces the accumulation of continuous layers of clast-supported and matrix-supported beds. The slow mass movement is responsible for the occurrence of frontal and lateral festoons and undulations. The presence of debris flows also contributes to the characteristic alternating structure (Van Steijn *et al.* 1995), though the limits and continuity of the beds can be poorly developed. Slopewash processes are almost completely excluded as the main mechanism, since most of the rock fragments are oriented parallel to the slope gradient. Furthermore, the absence of rills, longitudinal sorting and cross-bedding suggests the inability of overland flow to redistribute the debris along the talus.

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JOSÉ M. GARCÍA-RUIZ

GROUND WATER

Groundwater processes

Ground water is an important source of water for domestic use, irrigation and industrial uses and concern about the quantity and quality of ground water withdrawals is global in nature. It is a critical link in the hydrologic cycle, as it is a major source of water in rivers and lakes. Ground water is water under positive (greater than atmospheric) pressure in the saturated zone. The fluctuating water table marks the upper boundary of saturation in unconfined aquifers. Recharge can occur by infiltration of rainwater or snowmelt and by horizontal or vertical seepage from surface-water bodies. Ground water leaves the system by discharge into rivers, lakes or the ocean, by transpiration from deep-rooted plants, or by evaporation when the water table is close to the surface. Ground water is in continual motion, with velocities that are typically less than $1\,\mathrm{m\ day^{-1}}$.

The most important geologic factors controlling the movement of ground water are lithology, stratigraphy and structure, and combinations of these conditions produce a great variety of groundwater flow patterns. The term aquifer is used to define a geologic unit that can store and transmit enough water to be hydrologically or economically significant. Layers of rock which are impermeable are termed aquicludes and semi-permeable rocks, which retard the flow, are termed aquitards. An unconfined aquifer is open to the atmosphere and its hydrostatic level is the water table. In a confined aquifer water is held between confining layers (aquitards or aquicludes) and is not vertically connected to the atmosphere.

In humid regions, ground water may be an important contributor to streamflow, with water entering the channel by effluent seepage to form the baseflow discharge. If ground water input is significant, the streams will be characterized by relatively low temporal flow variability. By contrast, in arid regions streamflow often percolates into permeable beds to contribute to the water table. Such streams are referred to as influent.

Ground water as a geomorphological agent

Ground water is a significant geomorphological agent in many environments, both arid and humid, hot and cold. It influences cave formation in karst terrains; water chemistry and surface morphology of playas or PANs; the erosion of rock faces and formation of alcoves and caves; cliff retreat and mass movement; and canyon growth by basal sapping processes. Ground water can impede wind erosion in arid areas where the water table lies close to the surface and can affect dune type. Over time, the role of ground water in geomorphic development is strongly affected by fluctuations in the height of the water table which result from climate change and human pumpage. Excessive groundwater withdrawal and falling water tables can lead to surface subsidence, vegetation death and dune mobilization.

The term KARST is given to limestone terrains that include such distinctive landforms as caves, springs, blind valleys and dolines. The dominant erosional process is dissolution and the region is typified by lack of surface water and the development of stream sinks or dolines. A unique pattern of drainage results from karst processes. Solution creates and enlarges voids, which then integrate to allow the transmission of large amounts of water underground, thereby promoting further solution. In karst areas underground drainage is developed at the expense of surface flow networks. Solution and weakening of silicic rocks to form karst-like topography has also been noted in arid environments, such as the Bungle Bungle of north-western Australia, but it is uncertain whether such landforms are wholly or partly inherited from more humid climatic periods.

Ground water plays a role in mass movement and channel formation by the process known as sapping. Concentrated seepage caused by ground water convergence is capable of slowly eroding materials at valley head or cliff bases, undermining overlying structures, and causing failure and headward retreat. The term spring sapping is often used when a point-source spring is involved, whereas seepage erosion may be employed where the groundwater discharge is less concentrated. Computer modelling suggests that scalloped escarpments develop where groundwater flow is diffuse, whereas elongation into channels or canyons results from higher and more concentrated seepage discharges, often associated with growth updip along fracture systems characterized by higher hydraulic conductivities. In rocks which are susceptible, chemical weathering renders the rocks even more permeable. Although many sapping networks, for instance those of coastal Italy and the Colorado Plateau, have developed in highly jointed bedrock, field research by Schumm *et al.* (1995) illustrates that similar networks can develop in highly permeable sands without significant structural controls.

Common morphological characteristics of valleys in which sapping plays a dominant role include amphitheatre-shaped headwalls, relatively constant valley width from source to outlet,

Plate 57 Long Canyon and Cow Canyon are tributaries to the Colorado River, developed in the Navajo Sandstone. The morphology of these valleys, with theatre-shaped heads and relatively constant valley width from source to outlet, is consistent with their formation by groundwater sapping

high and often steep valley sidewalls, a degree of structural control, short and stubby tributaries, and a longitudinal profile which is relatively straight (Plate 57). Similarly, simulation models (Howard 1995) of groundwater sapping produce canyons which are weakly branched, nearly constant in width and terminate in rounded headwalls.

Interest in groundwater outflow processes as an important factor in valley network development was stimulated by imagery of Mars which revealed features that were broadly similar in morphology to those on Earth (Laity and Malin 1985). It is now widely believed that many valleys on Mars were probably the result of erosion by groundwater sapping (Gulick 2001), although the actual mechanism is still subject to conjecture and debate. On Earth, an ever-increasing number of research papers illustrate that groundwater sapping is a global process, which can occur in a number of diverse lithologic and hydrologic settings. Valleys formed by sapping processes have been identified in Libya, Egypt, England, the Netherlands, the United States (Vermont; the Colorado plateau; Hawaii; Florida), New Zealand, Japan and Botswana. Valleys and escarpments which maintain the characteristic forms outlined above, but which lack modern seepage, may be relict from previously wetter climates (for instance, in Egypt). Other systems may include both active and relict components.

In addition to forming valleys by headward erosion, sapping at zones of groundwater discharge also contributes to the backwasting of scarps. Slopes are undermined and collapse owing to the removal of basal support by fluid flow which weakens rock at sites of concentrated seepage or diffuse discharge. These processes have received particular attention when considering scarp retreat in sandstone-shale sequences of the American south-west. Additionally, slopes of sandstone, granite, tuff or other massive rock form may be modified to include alveolar weathering or tafoni. The term ‘dry sapping’ has been applied to the formation of these features, for although the rock surfaces may be encrusted by salts, they do not appear to be damp. By contrast, larger alcoves formed by ‘wet sapping’ show wet surfaces on at least a seasonal basis (Howard and Selby 1994).

Boulders and inselberg landscapes in arid regions, when exposed by mantle stripping, are collectively referred to as etch forms. Such landforms may have developed over periods of 100 My or more and had an origin beneath deep mantles (tens or hundreds of metres in depth) which weathered as a response to ground water, under the control of geothermal heat. It has been proposed that the residual rock masses formed at the basal weathering front and were later exposed as the regolith was stripped.

Playa surfaces in deserts vary considerably owing to the range of unique hydrologic environments. Where ground water discharges seasonally or perennially onto the playa surfaces salt crystallization is characteristic and salt crusts of varying thickness form. The surface expression of groundwater discharge and salt crystallization includes extremely irregular micro-topography, polygonal forms and salt ridges. Solution phenomena such as pits and sinkholes may occur. Beneath the surface, the sediments are usually wet, soft and sticky. Spring mounds, elevated forms which may have a central pool, form where the water table is higher than the playa surface.

Landscape changes associated with groundwater overdraft

When more water is withdrawn from an aquifer by pumping than can be returned by natural recharge, the system is considered to be in overdraft. Such conditions have geomorphic impacts. Overdraft conditions have led to measurable SUBSIDENCE of the ground (one to ten metres) in such areas as Mexico City, Tokyo, Hanoi, the Central Valley of California, the Houston–Galveston area of Texas, and Las Vegas, Nevada. Surface geomorphic expression includes the development of fissure systems, such as those at Yucca dry lake, Nevada, where parallel fissures are as much as 2 km long and perhaps 500 m deep. In Bangkok, Thailand, subsidence averages 1.5–2.2 cm/yr, but has occurred at rates as high as 10 cm/yr, causing damage to buildings and infrastructure. As the city is almost at sea level, the most serious impact has been flooding at the end of the rainy season.

Ground water plays a significant role in aeolian and fluvial systems of deserts. In channel systems, ground water forced to the surface by faulting or bedrock may flow at the surface for short distances in zones marked by dense phreatophytic vegetation – plants whose root systems draw water directly from ground water, and which dominate the riparian habitat. Phreatophytes affect hydraulic roughness and depositional processes, and their loss owing to water table drawdown often precedes episodes of stream widening.

Dune systems also change in response to a decline in water table elevation. In a 'wet aeolian system' the water table lies at or close to the surface and various stabilizing agents, such as vegetation, deflation lags, or cements allow accumulation while the system remains active. Drawdown of the water table may lead to a change to a 'dry aeolian system', where neither the water table nor vegetation exerts any significant influence, and surface behaviour is largely controlled by aerodynamic configuration (Kocurek 1998). In the Mojave Desert of California, sand released from degrading nebkhas (vegetation-anchored dunes) has reaccumulated downwind in migrating sand streaks and barchan dunes. Problems to nearby settlement include dune encroachment and blowing dust episodes.

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SEE ALSO: canyon; etching, etchplain and etchplanation; karst; pan; sandstone geomorphology; spalling

JULIE E. LAITY

GROYNE

Groynes are shore-perpendicular structures that are emplaced to control sand movement along a beach by altering processes in the swash and surf zones and providing a physical barrier to sediment moved as littoral drift.

Groynes change patterns of wave-refraction, wave-breaking and surf-zone circulation, generate rip currents, trap sediments on the updrift beach, reduce sediment inputs to the downdrift beach and redirect sediment offshore. Geomorphic effects include creation of wider beaches with steeper foreshores on the updrift sides, narrower beaches with flatter foreshores on the downdrift sides, lobate deposition zones downdrift of the tips of the structures and pronounced breaks in shoreline orientation (Everts 1979). The locally wider updrift beaches can enhance aeolian transport, and the subaerial portions of the groynes can form effective traps for blown sand, increasing the potential for creation of dunes on their updrift side (Nersesian *et al.* 1992; Nordstrom 2000). However, shoreline recession rates may be greatly increased on the downdrift side of groynes (Everts 1979; Nersesian *et al.* 1992), leading to truncation of beaches and dunes and loss of habitat.

Shortening, lowering or notching of existing groynes or construction of permeable pile groynes or submerged groynes have been suggested to allow for some sediment to bypass the structures in order to reduce downdrift erosion rates. Permeable pile groynes can reduce the longshore current while eliminating the effect of the structurally induced rip current, creating a more linear shoreline than occurs with an impermeable groyne and creating an underwater terrace that can reduce the erosion potential of waves crossing it (Trampenau *et al.* 1996). Submerged groynes retain the original aesthetics of the landscape, and allow beach traffic to proceed unimpeded, but their effects have been poorly studied (Aminti *et al*. 2003). T-groynes, built with a short, shore-parallel seaward end, are favoured in some areas to reduce scour and redirect rip currents, thereby reducing unwanted sedimentation offshore, but they can leave the beach in the centre badly depleted (McDowell *et al*. 1993).

Groynes can be used to best advantage when they are located where (1) sediment transport diverges from a nodal region; (2) there is no source of sand, such as downdrift of a breakwater or jetty; (3) transport of sand downdrift is undesirable; (4) the longevity of BEACH NOURISHMENT must be increased; (5) an entire reach will be stabilized; and (6) currents are especially strong at inlets (Kraus *et al*. 1994). Groynes also have considerable recreational value for fishing because they create new habitat and provide access to deep water. Combined pier/groyne structures have been built to enhance this value.

New groynes or alterations to existing groynes are now often included in beach nourishment plans, but groynes have been banned or strongly discouraged in some management policies (Truitt *et al*. 1993; Kraus *et al*. 1994). Instances of removal of groynes have been reported (McDowell *et al*. 1993), but there is little documentation of the results on beach change. Alterations to groynes to allow for some bypass of sediment are more common than removal and are better documented (Rankin and Kraus 2003).

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KARL F. NORDSTROM

GRUS

A product of *in situ* GRANULAR DISINTEGRATION of coarse-grained rocks characterized by its specific grain-size distribution, where the sand (0.1–2.0 mm) and gravel (> 2.0 mm) fraction predominate and may constitute up to 100 per cent of the total. The percentage of finer particles liberated by weathering is often negligible (Migoń and Thomas 2002). Thus, grus is not associated with any particular bedrock, although some rocks, e.g. mudstones, are unlikely to produce grus because of their grain-size composition. Granitic rocks, gneiss and migmatites are parent rocks that typically break down into grus.

The term is also used by sedimentologists to describe a product of accumulation of weathering-derived, poorly sorted, angular quartz and feldspar grains that have been subjected to very limited transport, usually towards the base of an outcrop. Such a sedimentary veneer of grus is particularly widespread in arid and semi-arid areas, where slope wash redistributes products of rock disintegration across PEDIMENTS.

Grus as the product of current superficial weathering of rock outcrops should not be confused with grus weathering mantles, which may be many metres thick and can be found in geological records. Grus saprolites may be defined as *in situ* weathering profiles, consisting almost entirely, or predominantly, of grus throughout, that grades into unweathered parent rock. Grus may occur at the base of a deep weathering profile and would represent a transitional stage in alteration of solid rock into a clayey weathering mantle, although there is evidence that many tropical deep weathering profiles do not have a basal zone of grusification and the transition zone is very thin. 'Arenaceous mantles' and 'sandy saprolites' are usually used as synonyms of grus mantles.

Grus saprolites are diversified in terms of their internal structure, depth and lithology. Many of them are homogeneous throughout, yet some contain frequent core stones, zones of more advanced breakdown along fractures or show sharp lateral or vertical contacts between weathered and unweathered parent rock. Core stones within grus profiles may be as large as 3–4 m across and be either closely spaced, separated by weathered fractures, or in isolation in an otherwise strongly disintegrated rock mass. Various depths of grus saprolites are reported and profiles more than 10 m deep are not uncommon. Mineralogical changes associated with grusification are usually slight and the content of secondary clay minerals in grus profiles is often insignificant (< 2 per cent). Among clays, interstratified minerals, kaolinite, halloisyte and vermiculite are the most common. The occurrence of gibbsite is reported, but its percentage is usually low and is likely to represent a transitional stage in the formation of kaolinite.

The origin of deep grus saprolites is still unclear and several mechanisms have been suggested to be responsible for opening of microfractures within and between the grains in the near-surface zones (Pye 1985; Irfan 1996). Microfracturing results from de-stressing of quartz and feldspars during weathering and may be enhanced by expansion of biotite after its HYDRATION. Development of intergranular porosity in response to partial solution along grain boundaries and transgranular microcracks may be an important contributing agent. However, advanced chemical processes play rather a subordinate, if any, role as indicated by minor amounts of secondary clay, preservation of easily weatherable minerals such as biotite and plagioclase, and limited degree of corrosion of quartz and feldspar grains.

Grus mantles are particularly widespread in areas of temperate climate, but they in fact occur in a variety of climatic zones and may be found in every climatic regime, both humid and semi-arid (Migoń and Thomas 2002). In low latitudes they occur alongside products of more advanced alteration, such as ferrallitic saprolites, as for instance in south-east Brazil. This distribution contradicts the claim, often made in the past, that production of grus is primarily controlled by climatic conditions, and that it is specific for a humid temperate climate. Generalization is further inhibited by the possibility that many grus mantles are not the result of weathering under contemporary climatic conditions, but are inherited from a geological past and different climatic regimes, and by the fact that many grus profiles are evidently truncated.

From a geomorphological point of view, grus mantles occur in three major settings. First, they are common within elevated plateaux and uplands, beneath gentle upper slopes and along valley sides. Second, they occur in hilly and inselberg landscapes, but hills may either be weathered throughout into grus or have only their lower slopes underlain by a grus mantle. Third, they are associated with highly dissected mountain areas. In some subtropical mountains, watershed ridges, spurs and isolated hills are very often deeply weathered and only the cores of unweathered bedrock protrude as massive domes from the widespread saprolitic mantle (Thomas 1994).

The common association of thick grus mantles with areas of moderate to high relief, often with a recent history of uplift, across the world's morphoclimatic belts, implies that dissected terrains of moderate relief are particularly suitable for thick grus to develop. This is because of free drainage, strong hydraulic gradient, tensional stress and rock dilatation. On the other hand, surface instability prevents grus profiles from attaining geochemical and mineralogical maturity. It has therefore been proposed that the deep grus phenomenon is a response of weathering systems to rapid relief differentiation, whether by tectonics, erosion, or both, and associated enhancement of groundwater circulation, although it is not exclusive to such settings (Migoń and Thomas 2002).

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SEE ALSO: granite geomorphology; weathering

PIOTR MIGOŃ

GULLY

'Gully' can refer correctly, if uncommonly, to clefts down cliffs and to several sorts of seafloor channels. Those uses stem logically from the root sense of a narrow passageway, following derivation from the Latin *gula,* meaning throat, the French *goulet,* meaning a narrow entry or passage (including into a harbour or bay), and the Middle English *golet,* or gullet.

Minor uses aside, however, *gully* predominantly denotes a small and narrow but relatively deeply incised stream course, difficult to cross or to ascend, for which words like valley and gorge are too grandiose. It ideally connotes a young cut, with steep sides and a steep headwall, that has been carved out of unconsolidated regolith, typically by ephemeral flow from rainstorms or meltwater. However, these are not required attributes and exceptions abound.

Gullies are very variable in terms of processes of initiation and growth, as well as conditions of substrate, vegetation and climate, so they vary greatly in appearance and can show distinct regional differences. Thus the literature contains diverse usages and many local synonyms, including DONGA, vocaroca, ramp and lavaka. Among the variations, gullies may be slit-like to lobate (expanded at the head end), and continuous or discontinuous (depending on whether or not the gully has become connected at grade to the main drainage system). They can grow downward from mid-hillslope positions or from the hill-toe up, or they can develop along valley floors. Most gullies have a relatively simple, single thalweg, but gully heads can split during headward retreat, thereby creating a dendritic shape, and once in a while two branches rejoin head-to-head, creating a ring valley around a central pedestal or hillock.

Gully has distinct but poorly defined connotations of size, and even vaguer implications of cross-sectional shape. A gully is bigger than a RILL, which is a small entrenched rivulet, small enough to be crossed by a wheeled vehicle or to be eliminated by ploughing. An ARROYO (or wadi or barranca) represents an entrenched stream, not necessarily very deep, that has a somewhat wider floor than a gully – one might hope to drive up an arroyo, but probably not up a gully. A gully has a greater width to depth ratio than a slot canyon, and is neither as deep nor as wide as a box canyon or a gorge (see GORGE AND RAVINE), all of which would also likely have rock walls, unlike typical gullies. A gully is ideally narrower and shallower than a ravine, although no size limits have been specified. Gullies and gorges can share equally steep and enclosing walls, but ravines can be more V-shaped in cross section. Floors of ravines are ideally less enclosed than floors of gullies, but need not offer easier access. Ravines can be cut in regolith or rock. Overall, gully and ravine overlap considerably (the French *ravine* explicitly includes both gullies and larger valleys). In popular usage, gullies, ravines and gorges are perhaps best separated by their implications of lethality: a fall into a gorge could easily be fatal, whereas only the terminally unlucky would die by falling into a gully, and falls into ravines are unpredictable. Gullies might therefore be considered to range approximately from 5 or 10 m long, 1 or 1.5 m wide, and about as deep, arguably up to the order of several hundred metres long, many tens of metres wide at the original ground level, and perhaps twenty or thirty metres deep. Use of 'gully' at the larger end of that spectrum seems most supportable when there is a gradation in size from similar but smaller gullies nearby.

Probably the most dramatic and common causes of gullies involve human misuse of the land, where sites are made vulnerable to erosion by deforestation or by a more general devegetation, via logging, burning, overgrazing, or establishment of fields (especially when on hillsides and when unterraced or ploughed down the slope rather than across it). Additional proximal causes related to humans include runoff along paths or tracks that run straight downhill, regrading of hillsides and diversion or concentration of runoff from roads or building sites uphill. Gullying can also be initiated when soil compaction tips the balance from infiltration to runoff, or following mass movement after a hillside is undercut or overloaded.

However, there are also many natural causes for gullying. Exceptionally intense or prolonged storms are a very common culprit in erosion, but critical increases in rainfall can also come about from such climate changes as increased annual rainfall, or increased storminess with no net increase in rainfall. Critical levels of devegetation can be reached by aridification or by natural fires.

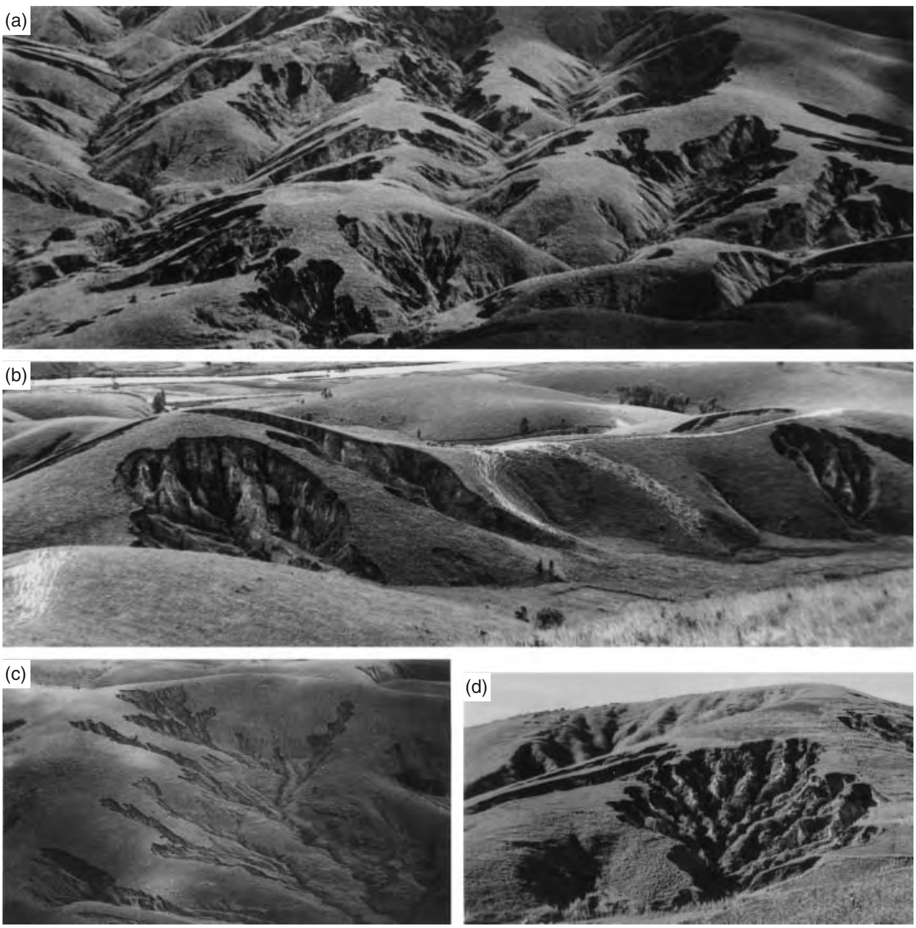

Plate 58 Examples of gully erosion in hills of weak saprolite in Madagascar. (a) Intense gully erosion. Concave-up runoff profiles are replacing smoothly convex hill profiles that formed by infiltration and chemical weathering. (b) Gullies at various stages of evolution. The biggest gully has expanded headward up dip, through the ridge crest. Ridge-crest cattle trails attest to endemic overgrazing in Malagasy hills. Many of these gullies receive no runoff from upslope. (c) These long and narrow gullies apparently represent entrenchment of a pre-existing dendritic drainage system. (d) Erosion dominated by runoff generated within the gully

Gully erosion may also be initiated by natural slope collapse, varying in style from deep slumps to soil slips, which can be triggered by rainstorms, earthquakes or undercutting by springs or rivers. Another cause is the collapse of pipes (see PIPE AND PIPING), which are natural subsurface passageways through soils, enlarged by rapid drainage along burrows, root voids, interconnected soil pores and the like. Critical increases in runoff relative to infiltration may happen naturally due to the plugging of a soil with fines or (especially in laterites) hardening due to exposure and drying following devegetation. Gullying can also be caused by incision or headward extension of first-order streams, as a result of uplift of headwater regions, subsidence downstream, a fall in base level, an increase in runoff and discharge, or breaching of a dam or a sill.

Note that the causes that make a site vulnerable to gullying can be very different from the specific initiators of erosion, such as when fires remove vegetation, but erosion starts only when sufficiently intense or prolonged rains attack the site. Thus, gullies may easily have different proximal and ultimate causes, and causes may operate at the gully, or upstream or downstream. Thus also, many gullies show threshold-related behaviour rather than linear responses to processes. Overall, gullying tends to be diagnostic of recently disrupted or non-equilibrium landscapes, whether perturbed by natural or cultural agents.

Once erosion is initiated, it can continue by a variety of processes, and the processes may change during growth. Rain attack and slope wash on the walls can be surprisingly effective, as can spalling or collapse of the walls from multiple wetting/drying cycles. Stream incision along the gully floor and concomitant undercutting of sidewalls are major growth processes. Another very effective process is headward retreat of a waterfall at the head end, which is notably associated with gullies that formed due to a lowering of base level. However, some gullies form entirely from rain that falls within the gully itself and have little or no exterior catchment area and no streams or rivulets flowing into them. In some situations, such as in thick saprolite in Madagascar, once the gullies have cut deeply enough to intersect wet zones near bedrock, groundwater seepage out of the bases of the headwall and adjacent sidewalls keeps the bases of those parts of the walls moist and vulnerable to further erosion and spalling, hence causing a dramatic increase in growth around the head end (Plate 58). This in turn creates very distinctive lobate or teardrop shapes, with broad arcuate heads and tiny exits. In these instances, the ultimate size of a gully is determined by the limits of the supply of subsurface water rather than the limits of the watershed on the surface. Gullies that grow by seepage and sapping can in some situations cut far back into high ground, effectively becoming amphitheatre-headed valleys. On occasion, they can even grow backward through ridge crests, for example when layering in the regolith dips through a ridge and delivers GROUND WATER from one side of a hill to the other.

In most cases, the shape and position of a gully reflect its causes and growth processes. For example, collapse of pipes, downcutting along stream beds, and headward retreat by springs or waterfalls create very long gullies, whereas growth by seepage and sapping can lead to a lobate shape with a broad and arcuate headscarp. Gullies that have grown along tracks and paths may lie along the crests of hill spurs, if those provided the easiest routes up the hill.

Most valley-side gullies represent new or future extensions of drainage networks into hillsides. Thus gullies commonly cut deeply into high ground, but they typically start within it and rarely pass through it at a low level, unlike most gorges. However, arroyo-like valley-floor gullies (which could form in the floor of a gorge) typically represent renewed incision of pre-existing drainages. Because they form most easily in unconsolidated material, gullies are common not only in colluvium on hillsides and alluviated valley floors but also in loess, till, outwash, loose fine volcaniclastics, laterite, saprolite and anthropogenic fill. The low resistance to erosion of these materials means that long-lived gullies are most typically associated with infrequent flow, which gives gullies a common but non-causal and non-exclusive association with relatively dry climates. (In contrast, ravines are more likely to have resistant walls and permanent flow.) It also means that gullies can grow extraordinarily rapidly. Thick regolith is generally a prerequisite for impressively deep gullies.

Left to themselves, gullies will eventually stop expanding, albeit possibly only when they have consumed their entire upslope watershed or have run out of erosible material. At that time, it will either have become a box canyon or ravine, or, if still cut into regolith, the upper walls will crumble back and the floor and lower walls will fill in and be buried, and the whole will become overgrown. Such a gully will become less angular and more rounded, with smoothly concave longitudinal and transverse profiles. Most will ultimately evolve into minor hillside hollows or reentrants.

Possibilities for remediating a gully include diverting drainage away from its head end; revegetating its walls and floors and/or the surrounding hillside; contour ploughing the surrounding hillside to promote infiltration rather than runoff; establishing small sediment-trapping dams along the floor, filling it in, and regrading the entire hillside. Nevertheless, the remedy should be matched to the specific causes and growth processes dominating each gully: disturbance of the surface (e.g. contour ploughing) will not be helpful if removal of vegetation is the critical initiator of erosion, and diversion of surface drainage may merely create a bigger problem somewhere else. If the gully is growing by sapping at the base of the walls, then work on the upper walls and the surrounding hillside may be at best a waste of energy, and effort should instead be concentrated on burying the wet zone, for example by promoting deposition along the gully floor. Overall, gullies are easier to prevent than to cure.

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NEIL A. WELLS

GUYOT

Although occasionally used to refer to any sizeable underwater ocean-floor edifice, the term 'guyot' should be confined to those that are flat-topped and were once above the ocean surface. This is to distinguish them from seamounts which are underwater volcanoes that have never been above the ocean surface.

The flat tops of guyots were thought for many years to be erosional – an expression of wave truncation of their summits during an island's slow submergence. The first to be studied in detail were in the Hawaiian chain, and submergence and summit truncation were thought to be natural and unavoidable consequences for an island moving along the chain (Hess 1946).

As ideas of Earth-surface mobility became fashionable in the 1960s, so it was realized that the distribution of guyots about mid-ocean ridges (seafloor spreading centres) was significant. Most such guyots had evidently originated at the mid-ocean ridge and then, following a period as a subaerial volcanic island (or atoll), they were submerged as they moved down the ridge's steep flanks and became guyots.

Another important step in the understanding of the significance of guyots came when they were found to be mixed in with atolls in various Pacific island groups like the Marshall Islands, Austral, Tuamotu and several in Kiribati (and some in the Indian and Atlantic Oceans). It has become clear that these guyots were once ATOLLS. That they are no longer is due to various reasons, including the morphology of the ocean floor in these regions (particularly the presence of intraplate swells) and oceanographic factors (principally temperature) which inhibit coral growth (Menard 1984).

Guyot morphology and location

The existence of low-relief surfaces on the summits of guyots is clear evidence for most authors of wave truncation, specifically shoreline erosion (at typical rates of 1 km/Ma) coincident with island subsidence (e.g. Vogt and Smoot 1984). Coral reefs on some guyots demonstrate that they are drowned atolls (see below). Phosphorites on certain Pacific guyots (at depths of 550–1,100 m) also derive from subaerial avian phosphorites (Cullen and Burnett 1986) demonstrating that these islands were once above the ocean surface.

Following observations and insights of Charles Darwin, it was proposed in 1982 that an oceanographic threshold existed in the Hawaiian chain which explained why atolls became converted to guyots at around 29 °N. It was proposed that at the 'Darwin Point', the gross carbonate production by corals was no longer sufficient for atoll reefs to regrow during periods of sea-level rise and so the atoll which had once existed became drowned (Grigg 1982). More recently it has been argued that other factors such as climate and sea-level history, palaeolatitude, seawater temperature and light all contribute to the Darwin Point which has shifted in the Hawaii region between 24 and 30°N within the last 34 Ma (Flood 2001).

Although guyots are commonly located at the older ends of hotspot island chains where these cross the Darwin Point, other guyots are located in equally instructive locations. For example, the morphology of guyots which have been pulled down into the Tonga–Kermadec Trench (southwest Pacific) has given us insights into the nature of tectonic processes across oceanic plate convergent boundaries (Coulbourn *et al.* 1989).

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PATRICK D. NUNN

GYPCRETE

The accumulation of gypsum ($CaSO_4 \cdot 2H_2O$) within a soil or sediment profile leads to the formation of gypsic horizons, which are called gypcrete. Gypcrete is a member of the dryland DURICRUST family, which also includes CALCRETE and SILCRETE. As gypcrete is more soluble than other duricrusts it seldom produces MESA landscapes with a gypsum CAPROCK. The process of gypcrete formation and its global distribution differs from that of other duricrusts. Nevertheless, calcrete and gypcrete may occur in the same profile and can be associated with salts such as halite. Gypcrete may also be host to one of the most commonly recognized forms of gypsum which is the desert rose.

Gypsum, the building material in the formation of gypcrete, is a very common mineral. It can be found throughout the world and forms under present-day evaporitic conditions in inland salt lakes (PANs), coastal salt flats (SABKHAs), springs, CAVEs, organic rich submarine sediments and in dryland soils. It is most commonly associated with massive sedimentary bedrock sequences, in particular those of evaporitic lake or sea basins. Gypsum formation generally requires evaporation of water and due to its solubility forms preferentially in areas of very low rainfall. It may also produce GYPSUM KARST if massive gypsum horizons are subjected to significant precipitation. Surficial gypsum does occur in all arid regions including the polar deserts, but massive gypcrete appears to be restricted to the drier subtropical desert regions of North Africa, the Middle East, southwestern Africa, southwestern America, South America, Central Asia and Western Australia.

Significant primary sources of gypsum formation are pans and sabkhas, which may also host gypcrete. These environments often feature shallow groundwater tables that are subject to substantial evaporation rates, which lead to the formation of evaporites above the water table in close proximity or at the surface. An upward migration of water and formation of gypsum is described as *per ascensum* gypcrete formation. When shallow ground water evaporates into a sandy substrate around a pan or sabkha margin, desert rose gypcrete forms. These crystals can be up to 30 cm in size and can be joined to form a single massive horizon.

Pans may also be subject to pronounced surface salinity gradients between the point of freshwater input and the point of saline brine formation at the centre. Such gradients are often accompanied by distinct evaporitic zones, which in a circular pan are arranged in concentric belts. Under such conditions, sulphate formation often follows the formation of carbonates and precedes the precipitation of chlorides and other salts. Pan surface gypsum may form hard gypsum crusts, which can develop a thrust polygon pattern.

Pan or sabkha environments may also accumulate unconsolidated fine gypsum crystals and powdery gypsum soils on their surface. Such freshly precipitated gypsum may not always form a hardened surface crust, but may be subject to aeolian dispersal (see AEOLIAN PROCESSES) Gypsum deflation may lead to gypsiferous lunette dunes, in particular at pan margins and will lead to the accumulation of gypsum-rich dust in the downwind environment. Aeolian dispersal of gypsum from pans is common and may be relatively rapid as indicated by significant burial of Roman artefacts in Tunisia (Drake 1997). It may also produce regional-scale gypcrete as demonstrated in the Namib Desert region (Eckardt *et al.* 2001).

Pedogenic accumulations form during sporadic rain, which dissolves surface dust and reprecipitates gypsum in stable soils or sediment profiles. Regolith cover in particular traps gypsum dust and gradually incorporates gypsum into the stable subsurface soil or sediments below the STONE PAVEMENT. It has been suggested that pavement surfaces are displaced upward during the process of gypsum dust entrapment (McFadden *et al.* 1987). The external and primary aeolian input of gypsum into a soil results in the relative downward migration of gypsum into the profile and is described as the *per descensum* mode of gypcrete formation. This process generally takes place in the absence of ground water. The resulting gypcrete can be massive in character, and may partly consolidate the surface regolith of a stone pavement. Stone pavement surfaces that cover significant gypcrete accumulations sometimes develop polygonal surface patterns.

The various crusts outlined above may differ considerably in terms of thickness, strength, composition and purity. The structure of gypcrete may range from powdery, nodular to massive horizons that vary in thickness from a few centimetres to many metres. Gypsum crystals may vary in size from microcrystalline (powdery) to massive (desert rose) and may include alabasterine morphologies as well as transparent lenticular clasts. A single crust may undergo multiple stages of formation with reworking, removal, production and storage of crust occurring simultaneously. This can produce a spatially complex and varied morphology. As a result no typical gypcrete profile exists. Gypsum crusts can however be defined as 'accumulations at or within 10 m of the land surface from 0.10 m to 5.0 m thick containing more than 15 per cent by weight of gypsum and at least 5.0 per cent by weight more gypsum than the underlying bedrock' (Watson 1985).

The formation of gypcrete is not only determined by climate but also by the provision of the elements, which produce gypsum. In particular sulphate is not as common as the elements required to form silcrete or calcrete. Sulphur isotopes have demonstrated that the formation of gypsum is dependent on the supply of dissolved sulphate or pre-existing sulphate accumulations such as bedrock gypsum. Gypcrete formation is thereby directly linked to the regional sulphur cycle (Eckardt 2001). Dissolved sulphate in surface or ground water leading to the formation of gypsum may be derived from the dissolution of sulphates or sulphides in the bedrock, the marine atmospheric contribution of sea spray, marine dimethyl sulphide (CH_3SCH_3) also known as DMS (Eckardt and Spiro 1999), gaseous hydrogen sulphide (H_2S) or the dissolution of aeolian sulphate from terrestrial gypsum dust sources.

We still have little information on exact rates of gypcrete formation and the response of gypcrete to climatic change (Plate 59). Attempts have been made to examine the micropetrography of gypcrete and to infer palaeoenvironmental conditions from such observations (Watson 1988). Dating of gypsum and gypcrete is possible using U-series dating and thermal luminescence techniques.

Plate 59 Example of a desert rose, the most commonly recognized form of gypsum. It forms with the evaporation of shallow ground water into a sandy substrate around a pan or sabkha margin

We have also been able to map gypcrete using remote sensing. Due to the distinct spectral response of gypsum in the mid-infrared region of the spectrum (2.08–2.35 μm) it can be separated from many other dryland features (White and Drake 1993).

Some powdery gypcrete accumulations may attain purities which make them attractive to mining while other deposits, in particular those fed by ground-water in Namibia and Australia, are known to be associated with high concentrations of uranium (Carlisle1978). In some areas pure gypcrete is also being used as a road paving material.

Gypcrete formation needs to be examined in the context of regional dryland processes, which include a combination of fluvial and aeolian processes. An understanding of pan or sabkha chemistry and hydrology is particularly important, as these are significant aeolian point sources of gypsum dispersal.

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SEE ALSO: duricrust; gypsum karst; pan; sabkha

FRANK ECKARDT

GYPSUM KARST

Karst associated with gypsum and anhydrite rocks is generally referred to as 'Gypsum Karst' and has received little appreciation by geomorphologists if compared to the normal (limestone) KARST. However, gypsum karst is widely spread in the world where the global gypsum-anhydrite outcrop exceeds 7 million km^2, the largest areas being in the northern hemisphere, particularly in the United States, Russia and the Mediterranean basin.

Due to the high solubility of calcium sulphate, the gypsum karst life cycle is commonly far shorter than that of carbonate karst. The average experimental values for gypsum degradation within the Mediterranean area were 0.91 mm/1,000 mm of rain (Cucchi *et al.* 1998) and therefore no outcrop of such rock may survive more than a few hundred thousand years if exposed to the meteorological agents. The actual evolution depends greatly upon the geological history of the particular region so that intra-Messinian and even older gypsum karst may have been preserved until the present.

Exposed karst forms

Medium- to large-sized gypsum karst landforms (DOLINES, BLIND VALLEYS, polje-like depressions) are very similar in genesis and morphology to those found on carbonate rocks, while meso-, micro- and nano-forms may sometimes be peculiar to a gypsum environment. The differences in meso-, micro- and nano-forms are normally the direct consequence of the fact that the size of the crystals in different gypsum outcrops may range from over a metre to a fraction of a mm, while in the carbonate rocks the crystal size is normally around a mm.

The most peculiar gypsum karst meso-form are 'tumulos', while 'weathering crusts' are amongst the micro-forms. All of them develop in gypsum formations characterized by a crystal size of 1–10 cm, which is normal for the Messinian gypsum in the Mediterranean area. Their evolution is produced by the increase of volume of the superficial gypsum stratum induced by the dis-aggregation of the rock texture, as a consequence of the anisotropic behaviour of the gypsum crystals with respect to temperature changes (Calaforra 1998).

Finally, whereas in carbonate rocks some of the micro- and most of the nano-forms are the result of biological activity, the outcrop of very large gypsum crystals (up to 1 m or more in length) together with their high solubility allows for the evolution of nano-forms, the morphology of which is simply controlled by the structure of the crystal lattice (Forti 1996).

Deep karst forms

In gypsum karst the single active speleogenetic mechanism is simple dissolution. Therefore, here deep forms are not so varied as in carbonate karst, where plenty of different speleogenetic mechanisms are active. Moreover, most of the dissolution–erosion forms (pits, canyons, domes, scallops, large collapse chambers) are quite similar to those present in carbonate ones.

Gypsum CAVEs are generally very simple linear or crudely dendritic caves that directly connect sink points and resurgence. They are commonly referred to as 'through caves' and consist of a principal drainage tube running along the water table with few and short, often subvertical, effluents; through caves are common in almost every entrenched and denuded gypsum karst area.

The deepest gypsum caves currently known rarely exceed 200 m in depth, being far shallower than those in carbonate karst: the reason is that always in mountainous regions, where the potential drained depth is greatest, gypsum formations are fragmented and do not favour the development of such vertically extensive sequences as do carbonates.

For the same reason the length of a gypsum cave rarely exceeds 2–5 km even if, in peculiar hydrogeological conditions (basal and/or lateral injection and dispersed inputs), complex dendritic 2- or 3-dimensional (multistorey) maze caves may develop up to several tens of kilometres. Podolia (Ukraine) is the 'type' region in which such caves have been explored and studied.

Chemical deposits

Chemical deposits (Hill and Forti 1997) are rather uncommon if compared with those present in carbonate caves: this depends mainly on the scarce chemical reactivity of gypsum. Normally they consist of calcite and gypsum and the local prevalence of one or the other mineral depends on climate.

Calcite SPELEOTHEMs show no morphologic peculiarities to distinguish them from similar deposits in limestone caves. Although, in most cases, their depositional mechanism is unlike that which dominates in a limestone environment (supersaturation due to CO_2 loss) being the product of the incongruent dissolution of gypsum by water with a high initial carbon dioxide content. Incongruent dissolution also explains the existence of unique forms like 'calcite blades' and 'half calcite bubbles'.

Gypsum speleothems have a more ubiquitous distribution. They present obvious morphological differences compared to calcite ones, due to their distinct genetic mechanism, which involves supersaturation due to evaporation. This genetic mechanism is also responsible for several unique forms such as 'gypsum balls', 'gypsum hollow stalagmites' and 'gypsum powder'.

Climatic influence on the chemical deposits

In gypsum caves climatic factors have a strong influence upon calcite and/or gypsum deposits. The completely different depositional mechanisms (incongruent dissolution for calcite and evaporation for gypsum) are influenced in very different ways by climatic variables: therefore climate strictly controls (far more than in carbonate karst) what chemical deposit can develop in a given gypsum cave. This close relationship with climate gives deposits preserved in the gypsum environment a potentially very great importance on the basis of their application to palaeoclimatology (Figure 79) and as indicators for present-day climatic changes.

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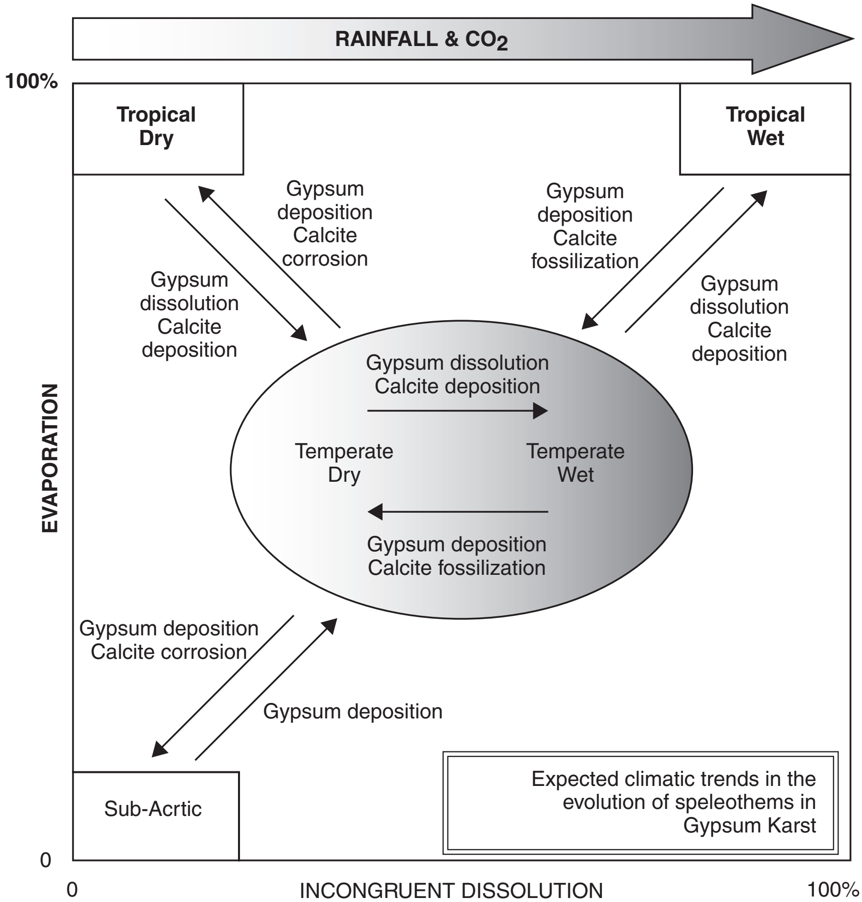

Figure 79 Expected climatic trends in the evolution of speleothems in gypsum caves

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PAOLO FORTI

H

HALDENHANG

Haldenhang is a German geomorphic expression introduced by W. Penck (1924, see translation 1953) for a 'basal slope – the less steep slope found at the foot of a rock wall, usually beneath an accumulation of talus'.

Figure 80 illustrates the formation of a haldenhang at the foot of a rock wall: all parts of the rock face but one are subject to erosion through rockfall, namely its base. The material there cannot fail because there is no gradient beneath it. This results in a parallel retreat of the rock face, during which its foot moves gradually upwards. Thus a rock slope of lower inclination appears, the haldenhang. Underneath a rapidly weathering rock face, the haldenhang may be covered by rockfall debris, forming a SCREE or TALUS. Over time the rock face will suffer incremental reduction in its height finally to be replaced by the haldenhang.

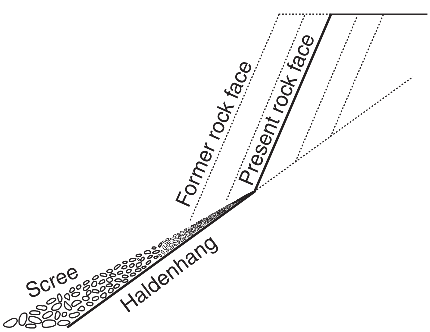

Figure 80 Formation of a haldenhang at the foot of a rock wall

W. Penck built his observations on haldenhang formation into his classic model of landform evolution and used them for explaining the erosional processes responsible for the downwearing of a relief (see SLOPE, EVOLUTION). According to Penck (1953), waning slope development starts with steep valley slopes being replaced by haldenhang of less inclination. Weathering of the surface material of the haldenhang will eventually produce finer material susceptible to creep and rainwash. Thus erosion of the haldenhang starts, resulting in a parallel retreat of the haldenhang and the development of a lower slope segment at its base. The whole process of slope retreat by mutual consumption produces concave and progressively lower slopes.

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CHRISTINE EMBLETON-HAMANN

HAMADA

Large rocky, unvegetated plateaux spread over dozens of kilometres in the Sahara, the Australian deserts and Libya, e.g. Hamadas of Dra, and in the Guir, north-western Sahara (Mabbutt 1977).

The surface of hamadas shows STONE PAVEMENTs which could either be residual and result from the disintegration of rock formations below or consist of boulders transported only short distances, forming a reg. The reg acts as a protective layer to the underlying formations of the hamada which represent forms of stabilized relief. The hamadas of the Sahara rest on erosion surfaces of varied age: Cretaceous, Oligocene, Miocene, Pliocene or Quaternary (Conrad 1969). The hamadas of the northern Sahara are characterized by the extension onto the southern Atlas piedmont of the detrital and lacustrine facies of the *torba*, which is a non-stratified sediment, interrupted by one or several levels of silicified dolomitic limestones, called the *hamadienne carapace*. The geochemical interpretation of the torba and the carapace is that it formed by continental sedimentation in a lacustrine environment as indicated by the abundance of neoformed attapulgite, dolomite and calcite.

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HANGING VALLEY

A tributary valley in which the floor at the lower end is notably higher than the floor of the main valley in the area of junction. Hanging valleys are a hallmark of GLACIAL EROSION in mountains, because the greater bulk of the trunk glacier was able to cut a larger valley cross section than those of smaller tributary glaciers, in consequence of which the floor of the main valley was eroded to a lower level. The relationship between the size of valley glacier troughs and ice discharge was first noticed by A. Penck (1905) who termed it as the 'law of adjusted cross-sections'. Indeed, geomorphometric assessments undertaken in more recent years strongly support his idea that discharge and trough size are mutually adjusted (Benn and Evans 1998: 365).

Hanging valleys have a variety of forms. In high mountain areas the cross profile will exhibit the typical U-shape of glacial erosion. If the tributary valley was not glaciated or only occupied by thin, cold-based ice the preglacial V-shape might prevail. In some cases a waterfall cascades over the lip into the main valley, but in high mountain areas headward erosion of the tributary stream has usually cut a narrow gorge into the lower reaches of the hanging valley floor.

Outside previously glaciated areas hanging valleys sometimes occur along youthful fault scarps or along coasts where the rate of cliff retreat is higher than the adjustment potential of the smaller streams, e.g. in the chalk cliffs in the south of England. Hanging valleys can also develop in karst areas where surface streams flow directly on the groundwater surface. If the main river is downcutting rapidly, the water level will be progressively lowered and the smaller tributaries will eventually turn into DRY VALLEYs hanging above the main valley.

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CHRISTINE EMBLETON-HAMANN

HEADWARD EROSION

Headward erosion is the process by which a stream extends upstream towards the catchment divide. Headward erosion occurs at a range of scales, from rills to large rivers, and in all environments. Drylands have been central to research into headward erosion because conditions that favour GULLY and ARROYO development are frequently found in these regions. Headward erosion may also be caused by RIVER CAPTURE.

In any channel network, approximately half of the total length of channels is in un-branched (first-order) fingertip tributaries. Environmental changes that promote channel extension therefore have a large potential impact on the landscape. During discharge events channel heads may advance great distances upslope, or retreat downslope if the hollow refills. In extreme cases, gullies can grow in length by tens of metres per year, and may also incise their channels creating steep ravine banks (Bull and Kirkby 2002). One possible end result of these processes is the creation of BADLANDs, where there is little or no remaining land suitable for agriculture.

Headward erosion occurs at the channel head (see CONTRIBUTING AREA). In terms of landscape dynamics, the channel head is one of the most important elements of the coupled hillslope-channel system. The location of the channel head controls the distance to the catchment divide and therefore influences the drainage density and average hillslope length of a catchment (although bifurcation frequency, confluence angles and tributary spacing are also important) (Bull and Kirkby 2002). The position of the channel head is controlled by the balance of sediment supply and sediment removal (Kirkby 1980; Dietrich and Dunne 1993). A change in any factor that influences this balance, such as fluctuations in climate or land use, alter the surface erodibility, sediment supply and runoff rates and may therefore result in headward erosion.

Channel extension results from a complex array of processes that reflect variations in slope, soil type, soil thickness, vegetation type and vegetation density. These processes include overland flow, pipe initiation and collapse, mass failures and hillslope processes (which have the reverse effect to headward erosion by infilling the channels).

Overland flow occurs in small rills or as sheets of moderate depth over large surfaces. For erosion to occur the rate of rainfall must be sufficient to produce runoff, and the shear stress produced by the moving water must exceed the resistance of the soil surface. Erodibility is a function of the permeability of the surface, the physical and chemical properties that determine the cohesiveness of the soil, and the vegetation.

In some areas there is a close association between piping (see PIPE AND PIPING and headward erosion. The erosive effects of flow through subsurface channels may result in TUNNEL EROSION and subsequent collapse to cause headward erosion. Piping intensity reflects a critical interaction between climate conditions, soil/regolith characteristics and local hydraulic gradients.

Mass failures also occur at channel heads to cause headward erosion. Failure of steep channel heads occurs when the driving forces exceed resisting forces. Channel heads are loaded by three different forces: (1) the weight of the soil, (2) the weight of water added by infiltration or a rise in the water table, and (3) seepage forces of percolating water (Bradford and Piest 1977). The change in water content is important because it has a strong influence on the shearing resistance of the soil. The shear strength is also influenced by freeze–thaw cycles and wetting–drying cycles. Vertical tension cracks tend to decrease overall stability by reducing cohesion, and when these are filled with water the pore-water pressure increases dramatically, often resulting in failure.

Hillslope processes such as rainsplash, wetting and drying cycles and frost action operate to infill channels, and hence reverse headward erosion. For incisions to grow the rate of sediment transport out of an incision must also exceed the rate of sediment input at the same point, otherwise filling will occur. The inter-rill hillslope processes involved are rainsplash and rainflow. Both processes depend on raindrop impact to detach soil material. Mass failures may also act to fill channels if failed material is not removed, but builds up at the base of headcuts.

Traditionally there are two conceptual approaches to understanding processes operating at the channel head, the stability approach (Smith and Bretherton 1972) and the threshold approach (Horton 1945). The stability approach emphasizes that the channel head represents the point where sediment transport increases faster than linearly downslope. This usually requires wash processes to dominate. The threshold approach takes the view that the channel head represents a point at which processes not acting upslope become important. The balance of sediment still determines whether the channel head becomes stable or migrates, but changing process domains drive incision. However, it is not clear whether there is always a change in process at the headcut, or whether a change in the intensity of the process operating, or a variation in the spatial distribution causes incision. The different approaches tend to be better suited to different environments and determine the two extremes of a range of factors that combine to produce channel heads. These models assist our understanding and prediction of headward erosion.

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SEE ALSO: arroyo; badland; donga; gully; pipe and piping; tunnel erosion

LOUISE BRACKEN (NÉE BULL)

HIGH-ENERGY WINDOW

Neumann (1972) suggested that in the mid-Holocene on tropical coasts there was a period when wave energy was greater than now. This occurred during the phase when the present sea level was being first approached by the Flandrian (Holocene) transgression and prior to the protective development of coral reefs. The 'window' may have operated on a more local scale on individual reefs with waves breaking not on margins of an extensive reef flat as now, but more extensively over a shallowly submerged reef top prior to the development of the reef flat (Hopley 1984).

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A.S. GOUDIE

HILLSLOPE-CHANNEL COUPLING

Fluxes from hillslopes to the channel system are controlled by the connectivity of process domains between different elements of the catchment system. Brunsden (1993) defines coupled systems as being ones where there is a free transmission of energy between elements, for example where a river channel directly undercuts a hillslope, whereas decoupled systems are ones where a barrier is present, for example in the case of a FLOODPLAIN buffering the input of the hillslope to the channel. The extent to which a hillslope is coupled to the channel is thus a function of any factor that affects its connectivity, and may relate to spatial variability of properties such as soil texture or vegetation cover (see OVERLAND FLOW). A floodplain may cause a hillslope to be strongly coupled to the channel if it has a low enough infiltration rate, or at times when it is already saturated. The main channel may be decoupled from the hillslope by the presence of minor channels running along the edge of floodplains (YAZOO channels) or human-made drainage channels.

The strength of coupling may affect the type of process that occurs on either side of the boundary. A channel directly undercutting a hillslope may cause the local gradient to be steep enough to initiate RILLS or gullies (see GULLY) on the hillslope, or may lead to failure of the base of the slope (e.g. Harvey 1994). In all cases, the amount of sediment fed into the channel will increase, and may cause it to avulse (see AVULSION) or change its planform (see BRAIDED RIVERS). The rate of removal of sediment from the base of a hillslope relative to its supply by processes on the slope will also affect the form of slope evolution (see SLOPE, EVOLUTION) in the longer term. Coupled slopes will tend to have more convex lower profiles whereas decoupled slopes will encourage deposition at the slope base leading to concave lower profiles. Strongly coupled slopes will also be more sensitive to changes elsewhere in the catchment system.

Consideration of the strength of coupling may also be important in an APPLIED GEOMORPHOLOGY context. SLOPE STABILITY from undercutting is again an important process here, while Burt and Haycock (1993) discuss the impact of floodplain buffers on water quality, for example due to pollutants carried by runoff (see RUNOFF GENERATION) from hillslopes.

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HILLSLOPE, FORM

What are hillslopes?

Most of the Earth's surface is occupied by hillslopes. Hillslopes therefore constitute a basic element of all landscapes (Finlayson and Statham 1980) and a fundamental component of geomorphologic systems (see SYSTEMS IN GEOMORPHOLOGY). However, there is an 'amazing absence of any precise definition' of hillslopes (Schumm and Mosley 1973; Dehn *et al*. 2001). Hillslopes have a very large variety of sizes and forms; and several more or less synonymous terms are used to describe the phenomenon hillslope, e.g. valley slope, hillside slope, mountain flank. The description of hillslope form is a fundamental problem in geomorphology (see GEOMORPHOMETRY).

Generally, a hillslope is a landform unit, that is, a part of the Earth's surface, with specific characteristics (see LAND SYSTEM). As a basic characterization, a hillslope can be defined as an *inclined landform unit* with a slope angle larger than a lower threshold β_{min} (delimiting hillslopes from plains) and smaller than a higher threshold β_{max} (delimiting hillslopes from vertical walls like cliffs or overhangs), which is *limited by an upper and a lower landform unit* (Dehn *et al*. 2001). A definition of hillslopes additionally has to include position within the landscape as an *external context*. A valley, for example, can only exist with its accompanying hillslopes. Moreover, *size and scale* context are important properties for definitions of hillslopes: a hiker in Grand Canyon might identify the components of the valley side as an individual hillslope itself, whereas a pilot flying over the scene defines the whole canyon side as a hillslope. Hillslopes are formed as the result of hillslope processes (see HILLSLOPE PROCESS) acting over different timescales. Therefore, hillslopes are units, where downslope component of gravitational stress (g sin β) plays a dominant role for acting hydrologic and geomorphologic processes. However, a hillslope is usually the product of a variety of processes interacting in space and time; therefore, hillslopes form sequences of hillslope units with different characteristics (compare Figure 81; see SLOPE, EVOLUTION).

Therefore, fundamental properties for the definition of a hillslope are: (1) local geometry, (2) external landform relationships, (3) scale, and (4) related processes. The utilization of these fundamental properties into a definition of hillslope depends on the perception or the specific application, that is, a specific semantic model for hillslopes (Dehn *et al*. 2001). In geomorphometry, hillslope forms are usually described as arrangements of individual hillslope units. This concept facilitates description and classification of hillslopes, and enables modelling of interaction of hillslope form with forming geomorphologic processes. Hillslope analysis therefore incorporates two major connected aspects: the decomposition of a hillslope profile into units, and the aggregation of a hillslope by arrangements of form units. These analysis steps are carried out in three dimensions for a hillslope or, in a simplified way, two dimensionally for a hillslope profile. The related terminology used here is listed in Table 23.

Hillslope units

Hillslope analysis is carried out by subdivision of a hillslope into different hillslope units. There have been several approaches to standardize hillslope units using qualitative terms (e.g. Speight 1990). Most commonly, a hillslope is described by a series of basic units describing changes in slope, curvature and processes along the hillslope profile.

- Ridge/crest/interfluve: convex/rectilinear unit; most stable unit in landscape, if of considerable width; mainly vertical water transport; more poorly drained soils.
- Shoulder/upper midslope: convex element; unstable unit due to erosion processes; minimum soil thickness.
- Backslope/midslope: usually rectilinear segment; unstable unit; intensive lateral drainage; sediment transport; soils of varying depth.
- Footslope/lower midslope: concave element; sediment deposition; unstable unit; soil thickness tends to increase.
- Toeslope/floodplain: concave/rectilinear unit; sediment input from upstream and hillslope; unstable unit; thicker soils.

Table 23 Basic components and terminology for hillslope analysis

Hillslope component	Definition
Hillslope profile	Flowline connecting drainage divide with thalweg
Hillslope toposequence	Arrangement of hillslope units within the hillslope
(Hill)slope unit	Part of hillslope with specific characteristics: segment or element
Segment	Unit of homogeneous slope angle
Element	Unit of homogeneous curvature
Convex element	Element with a downslope increase in angle
Concave element	Element with a downslope decrease in angle
Maximum segment	Segment, steeper than units above and below
Minimum segment	Segment, gentler than units above and below
Crest segment	Segment bounded by downward slopes in opposite directions
Basal segment	Segment bounded by upward slopes in opposite directions
Irregular unit	Slope unit with frequent changes of both angle and curvature

Source: Young (1972, modified and extended)

In the direction of contours, hillslopes are usually stratified into the elements HILLSLOPE HOLLOWs, spurs (or noses), and rectilinear valley sideslopes using plan curvature. Another method for hillslope unit classification is based on the position within the DRAINAGE BASIN: Young (1972: 4) distinguishes the 'component slopes': valley-head slopes, spur-end slopes and valley side slopes. Speight (1990) provided an exhaustive list of nomenclatures of different land elements, including many hillslope units. Hillslope units therefore generally incorporate different aspects of land surface form: (1) slope angle, (2) curvature, (3) position within drainage basin and (4) position within hillslope. These properties are used to derive quantitative models of hillslope units (see below). Moreover, hillslope units are related to different geomorphic processes and REGOLITH properties (see above, compare Speight 1990). This leads to the utilization of hillslope units for soil-landscape modelling (see SOIL GEOMORPHOLOGY), formalized for example in the concept of the CATENA.

One of the early quantitative approaches in hillslope analysis, based entirely on geomorphometric properties, was established by Savigear (1952) using the components profile intercept (constant slope gradient), slope segment (constant slope gradient consisting of several profile intercepts), and slope element (constant convex or concave curvature). These units are delimited by breaks of slope, which are characterized by a distinct change in slope gradient. This approach has been extended and quantified by Young (1972), who subdivided the slope into convex, concave and rectilinear units (Table 23).

Those early approaches mostly concentrated on quantitative description of hillslope profiles; however, a hillslope is not simply a linear feature, but a two-dimensional landform unit within the three-dimensional space, acting as a boundary layer of a three-dimensional lithological body. Characterization of local hillslope form is therefore based on the land surface derivatives: gradient, which has two components, slope angle and aspect angle; and curvature, which is usually described by two components in profile and contour or tangential directions (see MORPHOMETRIC PROPERTIES). Curvature can be classified into convex, concave and rectilinear surfaces (Young 1972). Hence, the combination of three slope profile curvature characteristics and three plan curvature characteristics leads to nine possible hillslope units, which are defined by Dikau (1989) as basic form elements of the landscape (Figure 81). They deliver a disjunctive description of the hillslope surface into units of homogeneous curvature characteristics.

Slope angle has been used to describe hillslopes by slope segments (Table 23). Young (1972: 173) compared several classifications of slope angle and proposed a system of seven classes: 0°–2° level to very gentle; 2°–5° gentle; 5°–10° moderate; 10°–18° moderately steep; 18°–30° steep; 30°–45° very steep; >45° precipitous to vertical/overhanging. Limiting angles describe the range of slope

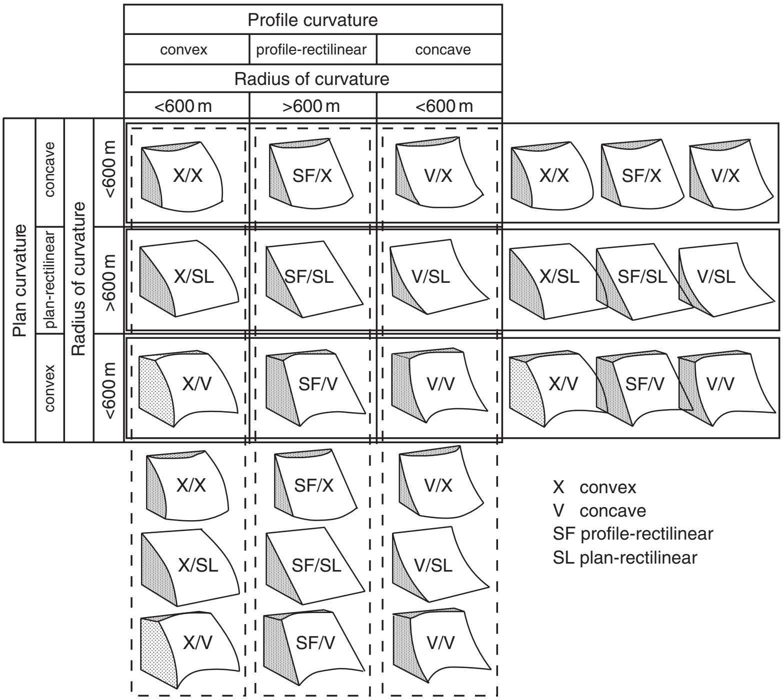

Figure 81 Fundamental hillslope form elements classified by plan and profile curvature (Dikau 1989)

angles within which specific slope forms occur. They include maximum and minimum limiting angles, which are related to the environmental conditions and the corresponding geomorphic processes. The angle of repose (see RESPOSE, ANGLE OF) defines the maximum angle for a given granular material type. Young (1972: 165) gives some figures for limiting angles of hillslope units under different environmental and lithological conditions.

An additional parameter for description of hillslope units is the position in the toposequence (Table 23) of the hillslope. Usually qualitative terms as upper slope, mid-slope, and lower slope are applied. Young (1972) used neighbourhood relationships to upper/lower units to describe hillslope units (see e.g. maximum and minimum segment in Table 23). However, for more complex profiles, these measures fail to describe absolute hillslope position, and quantitative rules, derived from total hillslope length or height, need to be introduced.

Investigations on hillslope forms are centred on process-form relationships, i.e. the explanation of a specific slope form (slope unit) by hillslope evolution, and contemporary processes. Young (1972: 92) gives a series of classical explanations of convex, concave and rectilinear slopes. Convex slope elements generally indicate erosion processes, which increase with slope length (surface wash), additionally soil creep and weathering has been identified as the dominating process regime for convex slopes. Rectilinear slopes generally indicate spatially homogeneous erosion conditions, i.e. the slope retreats parallel, or static transport units. Concave parts of hillslopes are explained by sediment accumulation due to constant base levels and/or by surface wash as an analogue to graded river profiles. However, as hillslopes are complex phenomena with an evolutionary history over long timescales, many interactions of processes and components can occur. Therefore, such simple assumptions generally do not match with the specific hillslope case and can only be used as guidelines.

Hillslope profiles and toposequences

Complete hillslopes are often represented by *hillslope profiles*. According to Young (1972) and Parsons (1988) a hillslope profile can be defined as a line on a land surface, connecting a starting point at the drainage divide with an end point at the thalweg, following the direction of the steepest slope. Hillslope profiles have been used to characterize various types of terrain using typical distribution of slope angles. Differences in the frequency distributions of slope angle are related to lithology (material resistance), climate (stress through rainfall, temperature), and evolutionary state of the slope (see limiting angles above). Hillslope profiles usually cover several process domains. Often the upper section of a slope is characterized by erosion, the middle section by transport and the basal section by deposition. Therefore, *toposequences* are used to describe the arrangement of different units within the hillslope profile. Characterization of these sequences delivers information about the slope system. It can be used to classify hillslopes. A toposequence may include one simple slope (single-sequence, e.g. a rectilinear element connecting ridge and valley), or two or more units (multi-sequence, e.g. convexo-concave slopes) (Speight 1990: 14). For multi-sequences, the order of the slope units (e.g. XMV for convex–rectilinear–concave slope) and the proportional length of the same units can be used to characterize the whole profile (Young 1972: 189).

The description of slope units within the context of the entire two-dimensional slope in three-dimensional space is also part of a toposequential analysis. Dikau (1989) and Schmidt and Dikau (1999) used parameters such as the neighbourhood relationship, distance to drainage divide, or height difference to the drainage channel to classify and aggregate complex hillslope systems.

Different models of hillslope profile form have been developed, which relate a specific hillslope toposequence to evolutionary history and contemporary geomorphic processes. Wood (1942) introduced the term 'waxing slopes' for convex hillslope units on crests developed by weathering processes acting on a cliff top. Likewise, Wood defines 'waning slopes' as depositional concave hillslope units developing at the base of a SCREE by sediment sorting due to aquatic hillslope processes.

King (see Young 1972: 37) developed a classical four-unit toposequential model based on work of Wood (1942). The crest (waxing slope) is a convex element of little erosion by weathering and creep processes. The evolution of the whole hillslope is driven by an active scarp segment of steep slope angle (rill erosion, mass movements). The downslope debris slope segment is formed by sediment provided by the scarp and determined by the angle of repose of the coarser material. The PEDIMENT (waning slope) is a rectilinear-concave, upward erosional element, produced by surface wash that connects to alluvial plains. Dalrymple *et al.* (1969) developed King's toposequence into a nine-unit slope model (Figure 82). The sequence consists of three low erosional upslope units, an intensively erosional unit (4), a transformational midslope (5), the depositional colluvial footslope (6) and three low-angled units associated with fluvial work. Conceptual hillslope models like these contribute to an understanding of the function of hillslope units and hillslope sequences, and can be utilized to classify hillslopes according to dominant process regimes. Numerical hillslope models (see MODELS) are used today to simulate process behaviour and evolution predicted by conceptual hillslope models, and thereby contribute to understanding of hillslope form and evolution based on current knowledge of process physics.

Measurement and analysis of hillslopes

There exist numerous techniques to measure hillslope froms. A selection of direct manual methods based on field observations and indirect measurements from maps and aerial photographs are described in Goudie (1990). Advances in computer technologies and the availability of DIGITAL ELEVATION MODELS (DEMs) have significantly revolutionized hillslope form analysis in the last decades. GIS technologies (see GIS) with algorithms to calculate morphometric properties, including slope gradient, slope curvature and flow paths, are now common tools for geomorphometric hillslope analysis (Schmidt and Dikau 1999). Raster-based DEMs are available at increasingly higher resolutions and, through the development of satellite data, on a global extent. As a result many numerical geomorphometric applications are raster-based. A typical GIS for hillslope analysis includes the following components (Table 24). Local morphometric properties of hillslopes

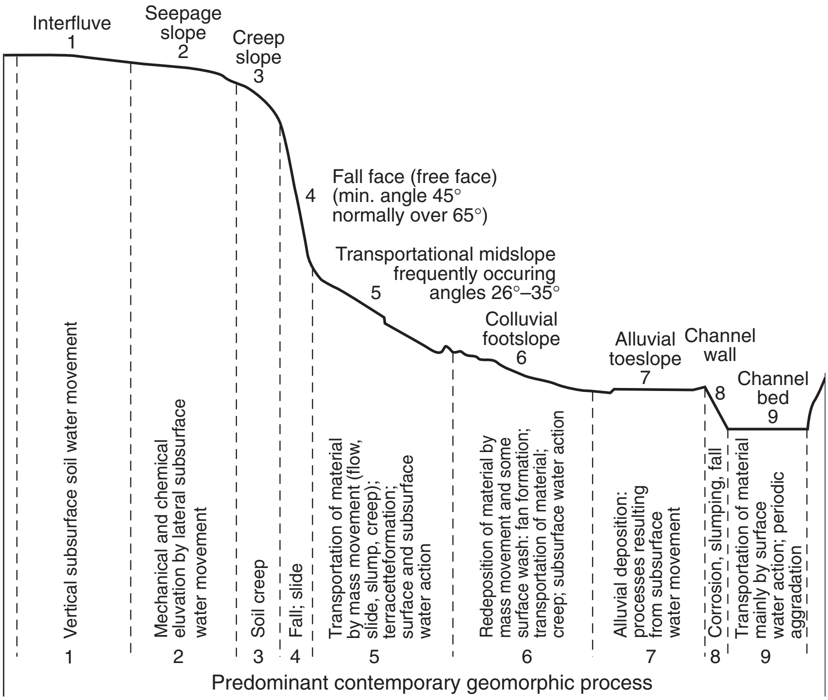

Figure 82 The nine-unit slope model of Dalrymple *et al.* (1969) (modified)

(height, curvature, gradient; see Schmidt and Dikau 1999) are generated from gridded DEMs through local interpolation, whereas complex parameters are based on flow routing algorithms. As slope and curvature are strongly dependent on scale, effects of DEM resolution on these parameters have to be considered, and preferably a specific scale for calculation of derivatives should be chosen. Hillslope units (areal geomorphometric objects after Schmidt and Dikau 1999) can be derived by GIS-based classification of slope, curvature and hillslope position (Dikau 1989). Linear hillslope profiles can be derived directly from a DEM by flow routing algorithms (Rasemann *et al.* 2003). Hillslopes as toposequences (geomorphometric objects of higher level after Schmidt and Dikau 1999) are derived by combining hillslope units according to their properties in gradient, curvature, position, and neighbourhood relationships. Hillslopes are characterized using representative geomorphometric parameters (Schmidt and Dikau 1999): frequency distributions and statistical moments of slope angle have been used to describe different types of slope profiles (Young 1972; Schumm and Mosley 1973). However, modern GIS technologies allow the calculation of many more hillslope parameters, including toposequential characteristics (Schmidt and Dikau 1999).

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Table 24 Hillslope analysis in a GIS

Hillslope parameters and objects	Algorithm	Input
Local geomorphometric parameters – local geometry		
Slope angle and aspect, profile and plan curvature	Local interpolator	DEM
Complex primary geomorphometric parameters – hillslope position		
Flowdirection	Local classifier	DEM
Upslope contributing area	Flow routing	Flowdirection
Downslope/upslope flowlength	Flow routing	Flowdirection
Hillslope position	Algebraic operation	Flowlength
Geomorphometric objects – hillslope units		
Form elements	Classification	Curvature
Slope segments	Classification	Slope angle
Hillslope units	Classification	Hillslope position
Geomorphometric objects – hillslopes		
Hillslope profile/flowpath	Flow routing	Flowdirection
Hillslope toposequence	Neighbourhood analysis	Form elements, slope segments
Representative parameters – hillslope characteristics		
Frequency distribution	Grid overlay	Hillslope profiles, Hillslope toposequences
Statistical moments of parameters	Grid overlay	Hillslope profiles, Hillslope toposequences

Notes: Local geomorphometric parameters are derived through a local interpolator. Complex geomorphometric parameters, which are related to landscape position, are derived through flow routing. Hillslope units and hillslope profiles are derived as geomorphometric objects. Hillslopes as a toposequential arrangement of units can be analysed by topological relationships of hillslope units

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SEE ALSO: catena; cliff, coastal; digital elevation model; drainage basin; geomorphometry; GIS; hillslope hollow; hillslope, process; morphometric properties; pediment; repose, angle of; slope, evolution; valley.

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HILLSLOPE HOLLOW

Hillslope hollows are elongate depressions within the bedrock of regolith mantled hillslopes. They have no obvious stream channel but serve as drainage lines that are integrated with the drainage network by either subsurface or surface topography. They belong to a morphological spectrum ranging from small low-relief upland depressions, a few tens of metres in length, through dells which may extend 200–300 m in length (Ahnert 1998), to valley head depressions, and DRY VALLEYs which have no active channels but are clearly of fluvial origin. Such features are found in many different countries, in a wide range of morphoclimatic and lithological conditions, and have many different modes of origin; thus giving rise to a varied and at times contradictory terminology. For example, United Kingdom usage represents a 'dell' as 'a small well-wooded stream or river valley' while in many other countries a dell is defined as 'a small dry valley with no trace of linear, fluvial erosion' and of periglacial origin (Fairbridge 1968: 250). Other forms of hollow occurring on hillslopes that are not integrated with the drainage network (e.g. individual landslide scars or fault sags) are excluded from this discussion. Reneau and Dietrich (1987) provide a literature review on hillslope hollows.

Hillslope hollows within the substrate are often filled with colluvium and other regolith material (Plate 60) and may or may not exhibit a depression in the landsurface. Because they can lack surface expression, the term 'colluvium-filled bedrock depression' has been suggested as a more accurate descriptive term for these features. In one study of 80 hillslope hollows exposed in road cuts, 37 were found to be associated with concave depressions in the slope surface, 35 occurred beneath planar slopes with no surface depression and 8 occurred on spurs. The cross-sectional form of the depression within the bedrock can be either V-shaped or broad-based (Crozier *et al.* 1990). Hillslope hollows are located headward of first-order channels or join higher order channels in positions similar to that occupied by first-order channels. Tsukamoto (1973) has used the term 'zero-order basin' to describe landsurface hollows, emphasizing their hydrological integration with the drainage network.

Plate 60 V-shaped colluvium-filled bedrock depression, in Pliocene marine sediment, Taranaki, New Zealand

The configuration of hillslope hollows in valley head settings (Figure 83) can be related to contemporary processes (Ahnert 1998; Montgomery and Dietrich 1989). The main criteria used to differentiate the various valley head forms are gradient, number of convergent hollows, and shape. There are four basic types that can be further subdivided on the basis of number of contributing hollows. Shallow gentle hollows (Figure 83a) are wide with a low gradient head and commonly a downslope topographic threshold. During prolonged rainstorms, this form of hollow generally produces saturated overland flow. Steep narrow hollows (Figure 83b), often with a head cut in the colluvial fill, are dominated by seepage and regolith landsliding. Funnel-shape valley heads (Figure 83c) usually result from the convergence of multiple steep hollows. Spring-sapping valley heads (Figure 83d) tend to be circular in shape with a distinctive low angle floor separated from convergent hollows by a marked break in slope at the spring line.

Several processes have been suggested as capable of producing hillslope hollows in different geomorphic settings (see Crozier *et al.* 1990), including landsliding, subaerial fluvial dissection, subsoil percoline erosion, gelifluction and seasonal meltwater and a combination of periglacial and fluvial processes. Cotton and Te Punga, (1955) demonstrate that hillslope hollows are products of alternating morphoclimatic regimes. They conclude that former stream channels, initially incised during the last interglacial became modified by periglacial mass movement processes in the subsequent glacial episode and eventually infilled by periglacial deposits and loess. Under present-day conditions these colluvial infills are being removed by shallow landsliding. Another cyclic infilling and evacuation process, but in this case involving landsliding as the initial hollow generating process (Figure 84),

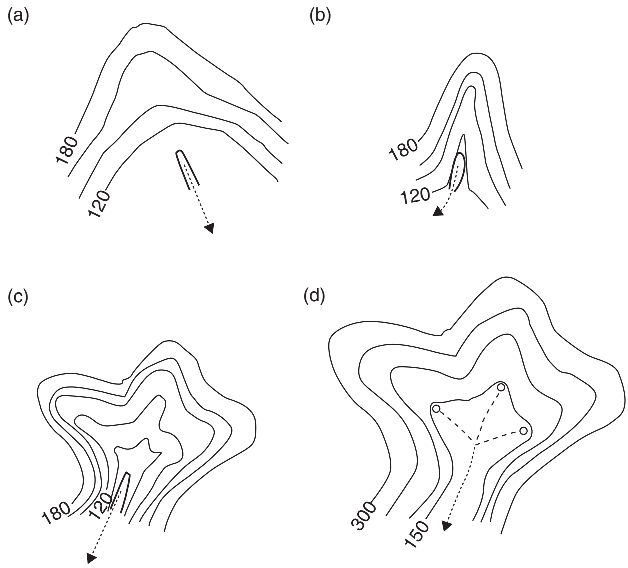

Figure 83 Contour pattern for types of valley head hollows: (a) shallow gentle hollow; (b) steep narrow hollow; (c) funnel-shaped valley head with three convergent hollows; (d) spring-sapping valley head (based on descriptions of Ahnert (1998) and Montgomery and Dietrich (1989))

has been described by Dietrich and Dunne (1978).

During rainstorms, the geometry of hillslope hollows directs both surface and subsurface runoff towards the centre line of the hollow. Accumulation of water within the hollow is a function of the contributing area and the ratio of side slope to thalweg gradients. During prolonged rainstorm events the hollows may become preferentially saturated acting as a source area for saturated overland flow. Regolith-filled hollows are also a preferential location for the generation of debris flows and debris slides. Compared to other hillslope locations, perched water tables are more readily established and accumulation of sediment surpasses the critical thickness required for failure. The scale of landsliding that occurs under these conditions is a function of the number of convergent hillslope hollows and their volume of stored sediment.

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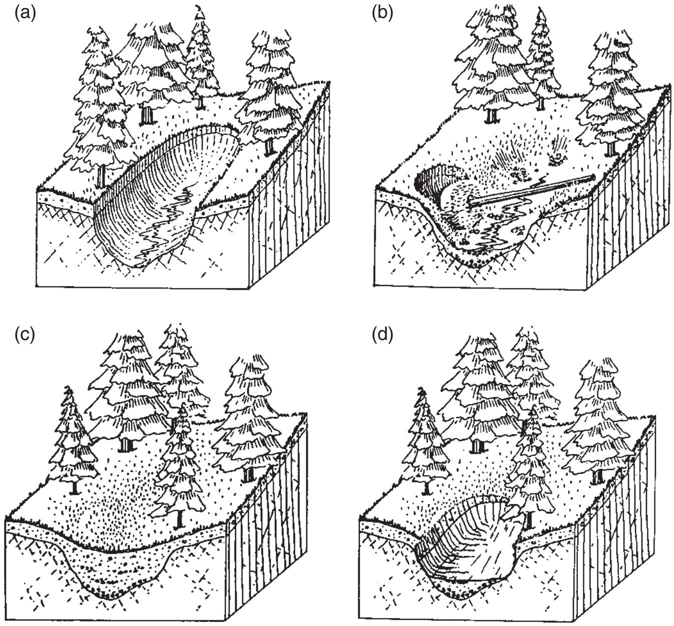

Figure 84 A model for the origin and evolution of hillslope hollows (based on Dietrich and Dunne (1978)): (a) bedrock landslide produces initial hollow; (b) peripheral debris fills hollow and is sorted by fluvial processes; (c) filled hollow becomes a site of concentrated subsurface flow and potential debris slide; (d) evacuation by debris slide

HILLSLOPE, PROCESS

The form that hillslopes take is a product of the materials of which they are made and the forces that act upon them. While the gross form of the landscape is determined by many factors, at the local scale a range of characteristic geomorphic processes shapes hillslopes. At the same time, hillslope morphology acts as a major influence on the occurrence, magnitude and nature of the processes themselves. Thus form and process tend toward a mutual adjustment. Hillslope processes are those geomorphic processes that involve the entrainment, transport and deposition of material from, over and on slopes. Their net effect is the transfer of material to lower parts of the landscape. This occurs either under the influence of gravity alone or, more commonly, with the additional incorporation of varying amounts of water. Flowing ice and wind also contribute to hillslope form, but operate at scales that are greater than the hillslope (see GLACIFLUVIAL; AEOLIAN PROCESSES; WIND EROSION OF SOIL).

A useful distinction can be drawn between two sets of processes that differ with respect to the role that water plays in the entrainment phase. First, MASS MOVEMENTs involve the entrainment of material due to the effect of gravity. The impetus for movement of material is derived from the potential energy inherent in the material by virtue of its elevation above BASE LEVEL, and the magnitude of the potential energy gradient induced by the inclination of both hillslope surface and strata. Water may play a crucial role as a preparatory or triggering factor, but it is the balance of geomechanical stresses within the mass of material that determines whether entrainment occurs (see FACTOR OF SAFETY). Once failure has occurred water is also significant in determining the nature of subsequent transport. In a second set of hillslope processes both entrainment and transport are effected directly by the kinetic energy of moving water.

The term mass movement can be applied to a broad spectrum of processes that are often classified in terms of the type of material involved (e.g. bedrock, debris or earth) and the type of movement (e.g. fall, topple, slide, spread and flow). This range of types of movement reflects an increasing significance of water in the transport phase; falls are almost exclusively gravitational, while, at the other extreme, flows almost always require the presence of water. A similar spectrum exists with respect to the importance of water for entrainment of material, i.e. the initiation of mass movement. In some instances failure may occur simply due to the effects of gravity, e.g. a rockfall. More commonly, water will have some influence on the balance of stresses within the slope material. Water is an important factor in preparing soil and regolith material for mass movement, and can often also be a triggering factor. The effects of water in various forms of physico-chemical WEATHERING produce a slow reduction in shear strength as rock is transformed into an engineering soil and rendered increasingly susceptible to gravitational stresses. More dynamically, the behaviour of water within hillslope materials frequently triggers failure. Variation in the height of groundwater tables can alter the balance of stresses by both increasing the mass of material (increased shear stress) and reducing its shear strength as a result of elevated PORE-WATER PRESSURES or reduced COHESION.

The second set of processes involves the entrainment of material through the energy imparted by flowing or impacting water. While not the direct cause of entrainment, gravity plays an important role in determining the velocity of flow, and hence its energy, and the direction of transport. The energy of a raindrop impact (see RAINDROP IMPACT, SPLASH AND WASH) is capable of detaching soil particles, which may be transported by the resulting splash. Although splash may be in all directions, the net effect is of downslope transport. Greater volumes of material may be entrained and transported by water flowing over the surface of unprotected soils (see SHEET EROSION, SHEET FLOW, SHEET WASH; OVERLAND FLOW). The mass of material that can be entrained will be controlled by the shear stress that is applied to the soil or regolith surface, which will be determined by the relationship between the HYDRAULIC GEOMETRY of the flow and the ROUGHNESS of the surface. The distance over which material of a given size can be transported is proportional to the kinetic energy of the flow. Where microtopographic convergences induce concentration into linear and turbulent flow, a threshold (see THRESHOLD, GEOMORPHIC) is reached and RILLS are formed. Once initiated, rills act to further concentrate flow. This concentrated flow can result in the formation of a GULLY, giving an example of positive feedback. Another important process involving flowing water is TUNNEL EROSION, which occurs when subsurface flow of water through the soil/regolith matrix is of sufficient velocity to initiate particle entrainment (see PIPE AND PIPING).

Although useful to illustrate the differing roles that water may play in hillslope processes, the distinction between mass movement and aquatic processes is often indistinct. The boundary between a highly fluid mass movement (e.g. an earthflow) and a sediment-laden stream may not always be easily identified. Strictly speaking, a rheologic distinction can be made between Newtonian and non-Newtonian flows; the former will involve both vertical and horizontal sorting of clasts during transport, while the latter implies matrix support and an absence of sorting. In reality the removal of material from slopes will occur as a result of a complex interaction between a range of different processes. A more important distinction can be made between those geomorphic processes that are diffuse and those that are not.

Diffusivity

Diffusivity is the property of being spread or dispersed, of not being specifically associated with any one place. Diffusive geomorphic processes, therefore, are those that are widely distributed in space. More importantly, diffusivity refers to the dissipation of energy. Thus, diffusive geomorphic processes can also be defined as those that occur where energy is dissipated over large areas (e.g. SOIL CREEP, sheet wash), while linear processes are characterized by the concentration of energy in a discrete unit of space (e.g. DEBRIS FLOW, RILLS). Consistent with the MAGNITUDE–FREQUENCY CONCEPT, low energy geomorphic processes occur frequently and with a wide spatial distribution. Conversely, high magnitude processes – those that concentrate large amounts of energy – occur more rarely and in a limited number of places. Hence, although diffuse geomorphic processes are often unspectacular, over longer periods they can accomplish large amounts of geomorphic work. Indeed, the predominance of either diffusive or non-diffusive processes has a large bearing on hillslope form (see HILLSLOPE, FORM).

Diffusivity tends to produce convex hillslope profiles. This is because of the adjustment of form to the energy available for entrainment and transport of material. On upper slopes close to drainage divides, with minimal catchment area, energy remains dissipated. Only small amounts of material can be transported. As catchment area increases with increasing distance from the divide, so too does available energy, and thus the amount of material that can be transported. The net effect of increasing entrainment and removal with greater distance from the drainage divide is the development of a convex long profile. At some distance from the drainage divide, available energy will be sufficient for the initiation of concentrated processes. This represents a threshold between diffusive and non-diffusive processes. It also represents a morphological threshold, and below this point long profiles are typically concave.

Characteristic form is therefore a reflection of process domain, and characteristic hillslope forms can be illustrated with reference to the processes that formed them. This may be on a local scale, with diffusivity implying that different processes are predominant on different parts of hillslopes (see, for example, the nine unit hillslope model developed by Dalrymple *et al.* (1969)). Available energy determines whether diffuse or concentrated processes can occur, and therefore how much geomorphic work can be done. There is thus a zone in which diffuse processes, accomplishing smaller amounts of work, predominate. These zones tend to remain constant; diffuse processes produce a characteristic low energy form, which further determines that only these low energy processes occur. Similarly, high energy processes tend to maintain the form that is necessary for their initiation until there is no longer sufficient potential energy available for their initiation (see PENEPLAIN).

However, there is a temporal element in this distinction. Given a sufficiently long period, many hillslope processes – especially mass movements – can be treated as spatially diffuse. For example, within a brief period landslides can be seen as singularities, concentrated in one particular place. Through time, however, as the sites of individual failures shift in space reflecting the availability of susceptible material, every part of the hillslope may be subjected to this process. It is important to note, however, that diffusivity in this case applies only in the sense of spatial dsitribution. The characteristic forms that are associated with the dissipation of energy through diffusive geomorphic processes will not necessarily occur. Indeed, hillslope form will generally be linked to the dominant geomorphic process, whether this is diffusive or concentrated.

At regional scales, the suite of processes that dominate will be determined by climatic and tectonic boundary conditions (see CLIMATIC GEOMORPHOLOGY, TECTONIC GEOMORPHOLOGY). Slope angle has an important control on the manifestation of gravity. Both slope angle and other MORPHOMETRIC PROPERTIES are influenced by tectonic phenomena and lithology, in addition to the action of hillslope processes themselves. Because of the role played by water in many geomorphic processes, climate is especially important, and both the amount and variability of rainfall will influence the type of processes that occur. Soils and vegetation, as products of climatic and geological phenomena, are important, particularly in their effect on slope hydrology. Mass movements dominate on steeper slopes. Aquatic processes tend to dominate in arid to subhumid climates with gentler slopes. Despite the greater availability of water in more humid environments, initiation of aquatic processes is often precluded by protective vegetation and deeper soils with greater capacity for infiltration and subsurface runoff.

Both vegetation and soil are especially susceptible to anthropogenic influence. Although anthropogenic, in many areas tillage is recognized as an important geomorphic process in its own right (see, for example, Govers *et al.* 1994). It is diffuse, with low magnitude, and has a pronounced effect on the form of hillslopes where it occurs. Because it is mechanically initiated, tillage is neither a mass movement as defined here nor an aquatic process. A mass of material is physically displaced with each application of the plough, and in this respect tillage might be considered to be analogous with a diffuse mass movement such as soil creep. Importantly, however, the mechanical disruption and displacement of soil material can also be seen as equivalent to physical weathering, providing easily erodible material for the operation of aquatic processes.

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SEE ALSO: freeze–thaw cycle; landslide; solifluction; threshold, geomorphic; unloading

NICK PRESTON

HOGBACK

A sharp, crested ridge of hard rock, with steeply dipping (> 20°) strata and steep near-symmetrical slopes. Hogbacks form as a result of slow differential erosion over time of alternating hard and soft strata. The soft rock is preferentially eroded, leaving steeply angled, slowly eroding resistant rock in place. The term is therefore derived from the resultant feature resembling a hog's back when viewed in planform. Examples of such features include Hogback Ridge, North Dakota, USA, Mount Rundle, Canadian Rockies and Gaishörndl, Austria.

STEVE WARD

HOLOCENE GEOMORPHOLOGY

The Holocene – or 'wholly recent' – epoch is the youngest phase of Earth history, which began with the end of the last large-scale glaciation on northern hemisphere continents other than Greenland. For this reason it is sometimes also known as the post-glacial period. In reality, however, the Holocene is one of many interglacials which have punctuated the late Cenozoic Ice Age. The Holocene is conventionally defined as beginning 10,000 radiocarbon (^{14}C) years ago, which is equivalent to about 11,500 calendar years. The term 'Holocene' was introduced by Gervais in 1869 and was accepted as part of valid geological nomenclature by the International Geological Congress in 1885. The International Union for Quaternary Research (INQUA) has a Commission devoted to the study of the Holocene, and several IGCP projects have been based around environmental changes during the Holocene. Since 1991 there has also existed a journal dedicated exclusively to Holocene research (J. Matthews, ed., *The Holocene*). A potted history of the Holocene can be found in Roberts (1998).

During the Holocene, the Earth's climates and landscapes took on their modern natural form. Geomorphological change was especially rapid during the first few millennia, with DEGLACIATION of the ICE SHEETs remaining over Scandinavia and Canada, and SEA LEVELs rising to within a few metres of their modern elevations in most parts of the world. Because soil formation and vegetation development lagged behind the often rapid shifts in climate, many landscapes – both temperate and tropical – experienced a phase of temporary geomorphic instability during this deglacial climatic transition (Thomas and Thorp 1995; Edwards and Whittington 2001). In river valleys, there were major changes in the discharge of both water and sediment, and many streams and rivers which experienced increased discharges at the end of the last glacial period are now underfit (see UNDERFIT STREAM). As a consequence of these changes, rates of denudation and sediment flux were frequently above the long-term geological norm at the start of the Holocene (Figure 85). A range of different sedimentary 'archives', including ALLUVIAL FAN, FLOODPLAIN and deltaic/estuarine deposits, can be used to establish long-term changes in rates of sediment accumulation and hence upstream soil erosion. On the other hand, river valleys are not closed systems, and quantitative SEDIMENT BUDGET calculations are more easily achieved by exploiting lake sequences. LAKEs act as receptacles for materials eroded from their catchments, and dated cores can be used to calculate volumes of sediment deposited per unit time (Dearing 1994).

In most coastal regions, recognizably modern shoreline configurations were achieved around

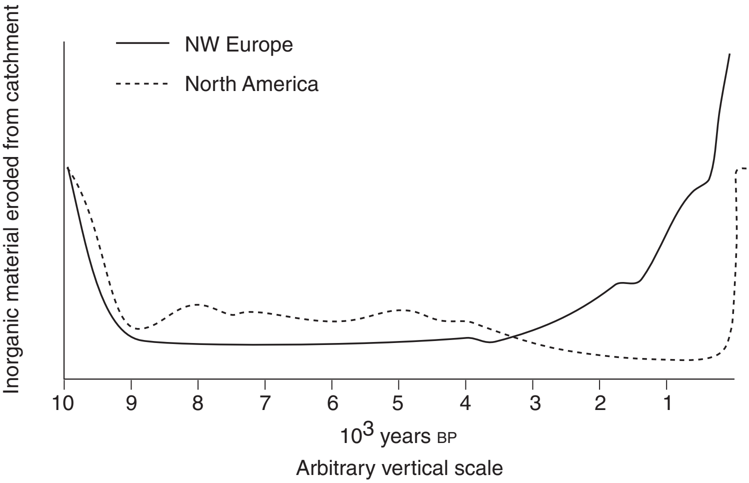

Figure 85 Generalized records of Holocene erosion based on sediment influx into lake basins (various sources, partly based on Dearing 1994)

7,000 cal. yr BP, with the main exceptions being in some high latitude regions such as Hudson Bay, where glacio-isostatic (see GLACIAL ISOSTASY) uplift has led to a continuing fall in sea levels during the Holocene. Elsewhere, rising sea levels during the early Holocene led to river valleys being drowned, with the end of this TRANSGRESSION representing the time of maximum marine incursion inland. Since then, stabilized sea levels and fluvially derived sediment discharge have led to a reversal in this trend, with the land pushing seawards at the mouths of major rivers such as the Rhône. This process has left many ancient harbour cities, such as Ephesus, Miletus and Troy in western Turkey, now stranded several km inland from the coast (Figure 86).

Various attempts have been made to subdivide the Holocene, usually on the basis of inferred climatic changes. Blytt and Sernander, for instance, proposed a scheme of alternating cool-set and warm-dry phases based on shifts in peat stratigraphy in northern Europe. In many temperate regions there is evidence of a 'thermal optimum' during the early-to-mid part of the Holocene. However, the clearest climatically induced environmental changes within the Holocene took place in the tropics and subtropics. One of the most important sources of palaeoclimatic (see PALAEOCLIMATE) and palaeohydrological (see PALAEOHYDROLOGY) information in low- and mid-latitude regions derives from non-outlet lakes, which can act like giant rain gauges. In East Africa, for example, lake levels were markedly higher and their waters markedly less saline between ~10,000 and ~6,000 cal. yr BP (Gasse 2000). On the other side, aeolian activity (see AEOLIAN PROCESSES) in regions such as the Saharan, Arabian and Thar deserts was greatly reduced and many sand dunefields (see SAND SEA AND DUNEFIELD) were inactive at this time. Rainfall in these regions increased by between 150 and 400 mm pa as tropical convectional rains moved further northwards, linked to a general strengthening of the African and Asian monsoonal system during the early Holocene.

By contrast with these climatically induced environmental changes, human impact has become an increasingly important agency in the creation and modification of landscapes during the later part of the Holocene. A critical point came when *Homo sapiens* turned to farming as the basis for human subsistence. The adoption and intensification of agriculture during the Holocene has led to widespread conversion of natural woodland or grassland into farming land, and this in turn has caused land degradation

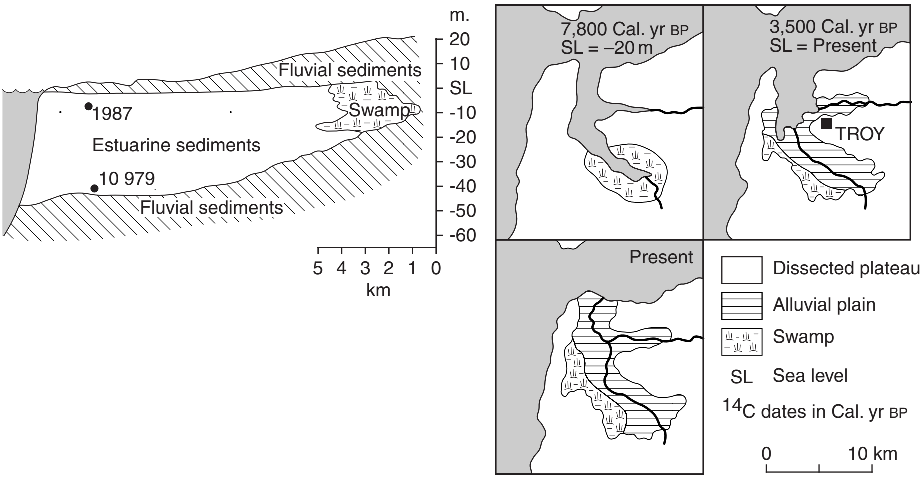

Figure 86 Geomorphological reconstructions in the vicinity of Troy, north-west Turkey, during the Holocene (based on Kraft *et al.* 1980)

through accelerated soil erosion and salinization. As a consequence, a much larger proportion of fluvial suspended sediment (see SUSPENDED LOAD) now originates from topsoil compared to bedrock sources than was previously the case. The same lake-sediment records that show high rates of flux during the major Pleistocene–Holocene climatic transition, typically show increases during the late Holocene associated with increasing human impact and land-use conversion (Figure 85). On the other hand, the dates for the onset of anthropogenically increased erosion rates vary from region to region, occurring earlier in Europe and in South and East Asia, but later in New World continents such as Australia and the Americas. At Frains Lake in the American Midwest, for example, soil erosion rates increased by two orders of magnitude to over 5 t/ha/yr^{-1} during the decade following the arrival of European settlement in 1830, but then stabilized at around 0.5–1 t/ha/yr^{-1} as forest clearance gave way to cropland (Davis 1976).

Fluctuations in climate have been superimposed upon the increasing human impact on geomorphic systems during the late Holocene. Particularly notable among these was the so-called 'Little Ice Age' from AD ~1400 to ~1850, when temperatures in Europe and the North Atlantic fell sufficiently for GLACIERs to advance down-valley in the Alps and other mountains

Plate 61 Modern and Little Ice Age limits of the Lower Arolla glacier, Switzerland

(Plate 61). This climatic deterioration brought a higher risk of geomorphological hazards including LANDSLIDEs, avalanches, glacier outbursts and other floods (Grove 2003), with comparable periods of extreme floods and droughts occurring in dryland parts of the world.

It is during the Holocene that modern boundary conditions for the Earth system have come into existence. Consequently the Holocene represents a key baseline for assessing human impact on Earth surface and atmospheric processes, in our attempts to tease apart the relative roles of natural and human agencies in rates of landscape change. It also provides the time frame over which long-term magnitude–frequency relationships (see MANGNITUDE–FREQUENCY CONCEPT) can be assessed and the return period for extreme events, such as floods, can be calculated (Benito *et al.* 1998; Knox 2000).

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NEIL ROBERTS

HONEYCOMB WEATHERING

Honeycomb weathering is a type of CAVERNOUS WEATHERING. The terms honeycomb, stone lattice, stone lace and alveolar weathering have been used as synonyms. Honeycombs are associated in

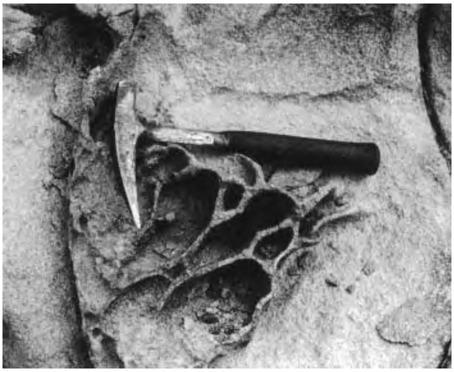

Plate 62 A small cluster of alveoles or honeycomb weathering forms developed on granite gneiss in a salty and foggy coastal environment on the south coast of Namibia, near Luderitz

particular with arid and coastal environments, though they are also a feature of some building stones in urban environments. They may also occur on Mars (Rodriquez-Navarro 1998). Many honeycombs are seemingly caused by SALT WEATHERING (Mustoe 1982; Rodriguez-Navarro *et al.* 1999). They are composed of pits, commonly some centimetres deep, that are developed so close together as to be separated by a narrow wall only millimetres thick. They are known from a wide range of rock types, including sandstones, limestone, schists, gneiss, greywacke, arkose and metavolcanics. In favourable environments they can form in a matter of decades (Mottershead 1994).

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A.S. GOUDIE

HOODOO

A common North American term for a pillar of eroded rock which is capped by a resistant rock layer, protecting a column of more erodable material beneath. They are effectively remnants of steep slopes as they are being eroded back, forming as the less resistant material is eroded away by water. The overlying hard caprock (generally a boulder or cobble) maintains the form's vertical integrity. Hoodoos are common in BADLAND morphology, and are typically formed in sedimentary rock (e.g. Bryce Canyon, Utah, USA) although examples exist in volcanoclastics (e.g. San Juan Mountains of Colorado, USA) and unconsolidated glacifluvial materials (e.g. Norway).

SEE ALSO: demoiselle

STEVE WARD

HORST

A relatively upraised fault block bounded by sharply defined and sometimes parallel reverse faults, though more commonly conjugate normal faults and opposing dips. The formation of the horst can be due to both the rifting and compressive movement of these marginal normal faults. Horsts are generally elongate ridge-like structures, with a plateau form on the uplifted horst block surface. Horst is a German term, and means retreat. Converse to horsts are graben (singular and plural). These are relatively low-standing fault blocks once again bounded by opposing normal faults, and occurring between zones of extension or compression. Half-grabens are grabens that are bounded by a normal fault on one side only. Areas of alternating uplifted and down-dropped fault blocks are thus referred to as horst and graben structures, and are associated with RIFT VALLEY AND RIFTING. In these regions, horsts are often the predominant sediment source into the down-dropped graben and any basins within. Examples of horst structures include the Black Forest and Harz Mountains in Germany, and the Vosges of eastern France. A famous graben structure is the Rhine Graben in Germany.

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SEE ALSO: fault and fault scarp

STEVE WARD

HORTON'S LAWS

In 1945, in one of the most significant twentieth-century contributions to geomorphology, the American engineer, Robert E. Horton endeavoured to express both the hierarchical arrangement and density of drainage networks in quantitative terms. In this he was explicitly following what he termed the 'ocular observation' (Horton 1945: 280) of the Scottish mathematician John Playfair (1802: 102).

> Every valley appears to consist of a main trunk, fed from a variety of branches, each running in a valley proportioned to its size, and all of them together forming a system of vallies, communicating with one another, and having such a nice adjustment of their declivities that none of them join the principal valley either on too high or too low a level.

Horton proceeded by introducing, first, the concept of STREAM ORDERING (1945: 281), in the following fashion:

> (U)nbranched fingertip tributaries are always designated as of order 1, tributaries or streams of the 2nd order receive branches or tributaries of the 1st order, but these only; a 3rd order stream must receive one or more tributaries of the 2nd order but may also receive 1st order tributaries. A 4th order stream receives branches of the 3rd and usually of lower orders, and so on. Using this system, the order of the main stream is the highest.

Somewhat unfortunately, Horton then developed an approach whereby the 'parent stream' had to be identified from source to mouth (Horton 1945: figure 7), something which is tricky and undoubtedly subjective (Figure 87a). A.N. Strahler (1957) proposed a modification of Horton's scheme (Figure 87b) which is less subjective and has been almost universally applied since 1957.

From the stream-ordering system, Horton proceeded to calculate two indices which he found to evince such regularity that they have become known as Horton's Laws (they are equally apparent whichever ordering scheme is used). The first observation was that the number of streams of different orders tended to follow an inverse geometric sequence. The ratio between each order being termed r_b, the bifurcation ratio. The second was that the average length of streams tended to increase as order rose. One set of Horton's data (and the value of r_b and r_l) are given in Table 25. If the data for stream numbers and lengths are plotted on semi-logarithmic paper, straight lines can be drawn showing roughly constant ratios throughout any one basin.

One further law – of stream slopes – was described by Horton (1945: 295), showing a geometric decrease in channel slope with increasing stream order; and yet two further laws (of basin area, which increases regularly with order) and the constant of channel maintenance were introduced by Schumm (1956). Horton's three basic 'laws' were set out (1945: 84 and figure 6) as:

Law of stream numbers: $N_o = r_b^{(s-o)}$
Law of stream lengths: $l_s = l_1 r_l^{0-1}$
Law of stream slopes: $s_c = s_1 \sqrt{r_s}^{(0-1)}$

Where: N_o is the number of streams in the drainage basin
r_b is the bifurcation ratio
s is the order of the main stream
o is the order of a given class of tributaries
l_s is the length of the main stream
l_1 is the average length of first-order streams
r_l is the length ratio
s_c is the slope of the main stream
s_1 is the average slope of first-order streams
r_s is the slope ratio

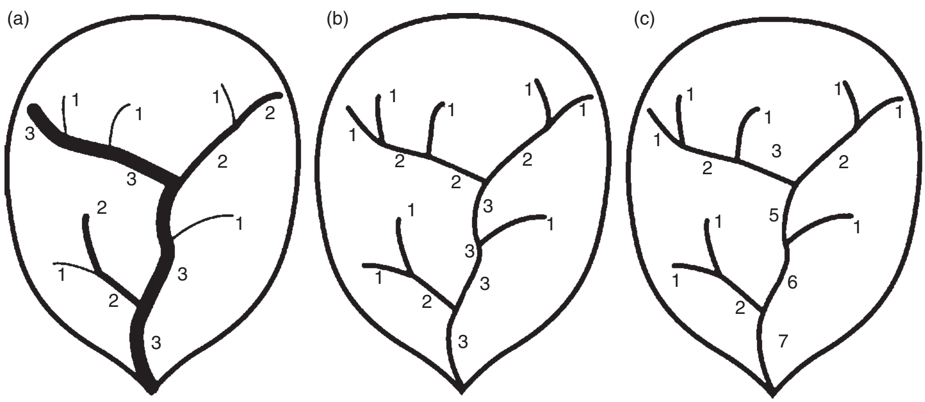

Figure 87 A drainage network with channels ordered: (a) according to Horton (1945); (b) according to Strahler (1957); (c) according to Shreve (1966)

Table 25 Drainage net, upper Hiwassee River

Order	No. of streams	Bifurcation ratio r_b	Average length, miles	Ratio of length, r_l
1	146		0.49	
		4.6		2.6
2	32		1.28	
		3.6		2.85
3	9		3.65	
		3.6		3.37
4	2		12.30	

Source: from Horton (1945: figure 7)

Schumm's additions may be stated in his words as: 'the mean drainage-basin areas of streams of each order tend to approximate closely a direct geometric series in which the first term is the mean area of the first-order basins' (1956: 606); and 'the relationship between mean drainage-basin areas of each order and mean channel lengths of each order is a linear function whose slope . . . is equivalent to the area in square feet necessary . . . for the maintenance of 1 foot . . . of channel' (1956: 607).

As Rodríguez-Iturbe and Rinaldo (1997: 6) make clear, the Hortonian laws, over the years, have been considered variously as demonstrating: that basins show a regular, evolutionary process; that basins demonstrate the development of purely topologically random networks; and that they say nothing whatsoever except that virtually all networks show these relationships. This last point is linked to the intuitively puzzling fact that Horton's Laws emerge whether basins are described by Horton's or by Strahler's methods.

However, there is a more fundamental difficulty with Horton's analysis as described in 1945: the categories of stream channels plotted on the x axis of the semi-logarithmic plots are, of course, *ordinal* numbers. As Horton's definition, quoted above and Figures 87a and b make plain, a 'second-order' basin may contain any number of 'first-order' channels between 2 and (effectively) ∞. Whilst it is quite proper to identify basins of different orders and to express numbers and averages in each class of variables and even ratios of values between classes, it is not proper to conduct the mathematical operations of multiplication or division using ordinal numbers and, in consequence, the 'regressions' shown on Horton's 1945 plots are simply spurious.

However, Horton introduced two other quantitative measures: DRAINAGE DENSITY (D_d) – the length of stream channels per unit area – and stream frequency (F_s), or the number of streams per unit area. Melton (1958) demonstrated that these two terms could be linked by a constant relationship:

$$F_s = 0.694\ D_d^2$$

This is now known as Melton's Law (Rodríguez-Iturbe and Rinaldo 1997: 8). These relationships described the degree of dissection of a landscape (see Kennedy 1978).

But the importance of the emphasis on stream frequency, involving actual counting of entities, was crucially developed by R. Shreve (1966) as the concept of basin magnitude, which is simply the number of first-order streams (or exterior links). Figure 87c shows the nature of this, true, ordering system, where magnitude is a ratio number that is susceptible of full mathematical manipulation.

Despite the difficulties with the Horton/Strahler ordering system, it is far easier to establish a sample of (say) fourth order basins than of 10 magnitude ones and a huge volume of research, especially although not exclusively from the 1950s and 1960s, focused on Hortonian relationships in generally fourth- or fifth-order basins. Church and Mark (1980) showed how this focus had tended to produce an apparent scale-dependence in the relationship between basin area and drainage density, which is actually isometric.

Nevertheless, the thrust of recent work on the FRACTAL nature of drainage basins, notably that summarized by Rodríguez-Iturbe and Rinaldo (1997), has been to substantiate the universality of the two fundamental Hortonian Laws: for example, their discussion of Optimal Channel Networks (OCNS) is shown to support both the Bifurcation Law and the Length ratio (1997: 278–279). Indeed, there is a major discussion of Hortonian networks which 'can be interpreted as signs of the fractal structure of the underlying network' (1997: 498). It is worth stressing, however, that what has endured have been the geometrically regular ratios of stream numbers and lengths between adjacent stream orders, rather than the spurious 'regression' plots which were such a feature of mid-twentieth century investigations of Horton's Laws.

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SEE ALSO: drainage density; fractal; laws, geomorphological; stream ordering

BARBARA A. KENNEDY

HUMMOCK

Small mounds of low relief, which cover the ground surface, are common where fine-grained soils overlie PERMAFROST. Most hummocks are circular, and 1 to 2 m in diameter. They are domed, with a vertical relief of up to 25 cm, but usually less than 15 cm. The ACTIVE LAYER is thickest beneath the hummock centres, and thinnest at the circumference. The base of the active layer is bowl-shaped. Segregated ice lenses, subparallel to the base of the active layer, are characteristically abundant directly beneath hummocks, and this ice-rich zone is commonly also rich in organic material. Hummocks are generally stable features, which may persist for thousands of years.

At the surface, organic material accumulates around the edges of hummocks, but the centres may be bare of vegetation (mud hummocks) or covered by peat or vascular plants (earth hummocks). The soil in hummocks is frost-susceptible, and may contain little sand. Where the clay content is low, the soil may liquefy in response to small changes in moisture content of stress, and be extruded at the ground surface. Such mudboils may occur in fields of hummocks, but in general the clay content is of the order of 40 to 50 per cent, and sufficient to prevent liquefaction.

The hummock form is maintained by soil circulation within each feature, driven by moisture redistribution during freezing and thawing (Mackay 1980). The soil circulation proceeds by upwards movement in the middle of hummocks, spreads to the circumference near the surface, and slides downwards at the edges, near the base of the active layer. The upward movement is driven by convection, due to the contrast between mud of relatively low density, and enclosing sediment, where the mud is supersaturated by melting of ice lenses. The movement at the base of the active layer is associated with heave towards the hummock centre during upfreezing at the base of the active layer, and settlement down the bowl-shaped frost table as ice lenses thaw. At the surface, soil is driven outward by heave and subsidence during freezing and thawing over an inclined plane. These three processes are constrained by the requirement for conservation of mass. Evidence for the circulation has been provided by movement of markers at the ground surface, and by involutions in the soil stratigraphy when viewed in cross section. The importance of the bowl-shaped frost table on hummock form is demonstrated by the disappearance of hummocks in the years following forest fires, when the active layer deepens, and their reappearance with subsequent vegetation regeneration as the active layer thins.

In the boreal forest, trees on hummocks are tilted, and are commonly located near the edges of hummocks. Tilting of the trees is associated with development of the ice-rich zone at the base of the active layer and accompanying heave of the hummocks. Hummocks are the complementary feature for fine-grained soil of sorted circles in coarser materials.

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C.R. BURN

HYDRATION

Hydration is the uptake of the entire water molecule by a mineral. For example, calcium sulphate (anhydrite $CaSO_4$) is hydrated to gypsum $CaSO_4 \cdot H_2O$. This results in the mineral swelling. In a confined space, hydration pressures can be up to 100 Mpa, weakening the rock. In cold climates, White (1976) felt that much freeze–thaw weathering could actually be hydration shattering, with the forces of hydration as high as 2,000 kg cm^{-3}.

Widely occurring is the conversion of iron oxide (Haematite Fe_2O_3) into iron hydroxides variously cited as being in a poorly defined crystal form as $Fe(OH)_3$, as goethite 2FeOOH or as limonite ($2Fe_2O_3 \cdot 3H_2O$). The formation of these iron hydroxides involves considerable volume increases.

Alumino-silicate minerals can become subject to hydration through the formation of hydrated aluminium oxide. HYDROLYSIS can be seen as more important than hydration because it is the products of hydrolysis which are hydrated. For the formation of the hydrated aluminium oxide from microcline, a potassium-containing feldspar:

$$KAlSi_3O_8 + H_2O \rightarrow HAlSi_3O_8 + K^+ + OH^- \quad \text{(hydrolysis)}$$

Microcline

$$2HAlSi_3O_8 + 11H_2O \rightarrow Al_2O_3 + 6H_4SiO_4 \quad \text{(hydrolysis)}$$

$$Al_2O_3 + 3H_2O \rightarrow Al_2O_3 \cdot 3H_2O \quad \text{(hydration)}$$

However, it should be stressed that it is the hydration which facilitates the physical disintegration through volume expansion, weakening the mineral structure.

In addition to this formation of new hydrated minerals, more complex layered minerals can take up water between their layers and this can also be referred to as hydration. The plate-like minerals, such as mica, can be subject to expansion and physical disintegration when water penetrates between the plates. Water can be incorporated into a clay crystal lattice and, especially in the open lattice of montmorillonite clays, involving increases in volume of around $0.5\,cm^3\,g^{-1}$.

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STEVE TRUDGILL

HYDRAULIC GEOMETRY

Hydraulic geometry of a river is the quantitative (mathematical and graphical) description of the channel cross section size and shape, fluid-flow properties and sediment-transport characteristics, in relation to the discharge being conducted by the channel. As such, every river channel, with rigid or deformable boundaries, has a hydraulic geometry. It is a descriptive device, derived from the empirical relations of regime 'theory' developed to aid canal design in India early last century (Lacey 1929). These ideas were first introduced into geomorphology by Leopold and Maddock (1953) who proposed the term 'hydraulic geometry' for this descriptor of the morphodynamics of alluvial channels.

The general equations of hydraulic geometry proposed by Leopold and Maddock (1953) necessarily are selective, reflecting relations among variables that were routinely measured or easily derived from such measurements made at US gauging stations:

$$\begin{aligned} w &= aQ^b \\ d &= cQ^f \\ v &= kQ^m \\ s &= gQ^z \\ n &= tQ^y \\ ff &= hQ^p \\ Q_{susp} &= rQ^j \end{aligned}$$

where w, d, v, s, n, ff and Q_{susp} are respectively width, mean depth, mean velocity, water-surface slope, flow resistance (Manning's n or D'Arcy Weisbach ff) and suspended-sediment load. An important missing element of this seven-variable set is bedload transport but these measured data are rarely available.

Implicit in the specification of these equations of hydraulic geometry are the following notions:

1 Discharge, Q, is the dominant independent variable in the hydraulic geometry;
2 The relations between the independent and dependent variables can be described as simple power functions;
3 As power functions, the logarithm of the dependent variables plot against the logarithm of discharge as a straight-line graph (that is, there is a linear relationship between the order-of-magnitude increases in the pairs of variables);
4 The existence of these orderly hydraulic-geometry relations implies an underlying set of processes reflecting the operation of equilibrium in the morphodynamic system;
5 Because continuity must be satisfied in fluid flow it follows from the rules of algebra that:

$$Q = wdv = (aQ^b)\ (cQ^f)\ (kQ^m)$$
and that $ack = 1$ and $b + f + m = 1$

The adjustment of channel morphology and hydraulics in relation to changes in discharge has been considered in two quite different contexts: at-a-station hydraulic geometry and downstream hydraulic geometry.

At-a-station hydraulic geometry

The at-a-station hydraulic geometry describes how the channel geometry and hydraulics of flow change as discharge increases at an individual channel cross section of a river. Thus it describes the way in which the flow fills the often effectively rigid channel boundary as discharge changes over time. An example of this type of hydraulic geometry is shown in Figure 88 for the Fraser River in western Canada. At-a-station hydraulic geometry is only defined for discharges up to the channel-filling (or bankfull) stage.

Downstream hydraulic geometry

Discharge in a river also increases as tributaries join the main stem in the downstream direction and add flow to the fluvial system. The downstream hydraulic geometry describes how this spatially increasing discharge enlarges and shapes the channel and alters the properties of the streamflow. In order to allow for comparisons between channel sections these changes are referred to as a discharge of constant return period or to a consistent relative stage. The most common reference discharge is bankfull discharge, which is often taken to be the channel-forming discharge. An example of this type of hydraulic geometry is shown in Figure 89 for Oldman River in western Canada.

Theoretical context and interpretation of hydraulic geometry

Conventional hydraulic geometry describes a partial picture of adjustments in the fluvial system but contains little information about the controls on such adjustments. When responding to changes in discharge, an alluvial stream must satisfy at least three sets of physical relations: continuity, flow resistance, and sediment transport. The first relation is definitional but the other two relations are only understood in a qualitative

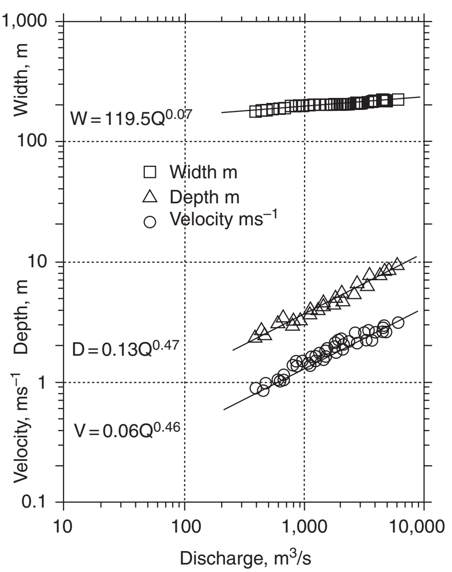

Figure 88 Hydraulic geometry for the Fraser River, Canada

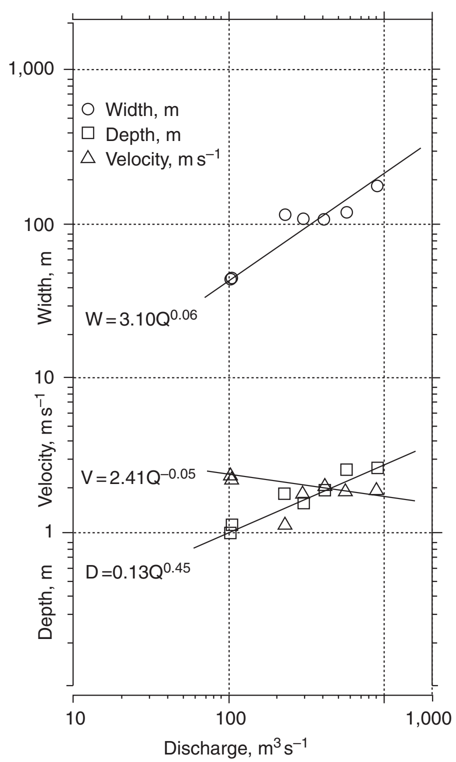

Figure 89 Hydraulic geometry for the Oldman River, Canada

sense. For this reason, and also because channels are free to adjust in ways other than by changes in width, depth and mean velocity (Hey 1988), the hydraulic geometry of alluvial channels generally is regarded as being indeterminate. Nevertheless, these physical relations do inform our interpretation of the hydraulic geometry at a qualitative level of analysis.

The primary difference between the two hydraulic geometries is that, unlike the downstream case, for the at-a-station case the boundary materials and the water-surface slope essentially remain constant as discharge changes. The principal control on the hydraulic geometry is the shape of the channel cross section formed at high discharges. Channel form in turn largely is determined by the strength of the boundary materials. If the boundary materials are cohesive and strong (mud-dominated rivers, for example), banks can develop which are steep and high. In such cases, as discharge increases, channel width will change much more slowly than flow depth and velocity. The rate of increase in flow velocity depends on the changing relative roughness of the channel. Typically, but not always, flow resistance declines as the effects of roughness elements are drowned by the increasing discharge. Thus the exponents on the velocity/discharge relation are relatively high. A different geomorphic response can be expected in the case of channels with weak non-cohesive banks (such as sand-dominated channels). In this case bank height is limited by material strength and the capacity for mean depth changes to accommodate increases in discharge is small. Change in velocity is also highly constrained by the conservative adjustment in flow depth and as a result channel width changes greatly to accommodate the discharge increases in such channels. Table 26 shows typical exponent values in the at-a-station hydraulic geometry equations for a variety of channel types.

In the case of downstream hydraulic geometry, adjustments of river channels to increases in bankfull discharge also reflect spatial changes in water-surface slope and the size of boundary-materials along the channel. A major control in this case is the balance that is struck between the impelling and resisting forces acting in the downstream direction. Although downstream increases in discharge and declining size of boundary material tend to work together to accelerate the flow, the forces for this change are almost equally countered by the decline in boundary shear stress related to declining water-surface slope. As a result, downstream hydraulic geometry is characterized by constant or declining mean velocity downstream and by the necessity for changes in discharge to be almost fully accommodated by adjustments in channel width and mean flow-depth alone. As before, the apportioning of the discharge change between width and velocity depends largely on boundary material strength. In a mud-dominated channel which can support steep and high channel banks the exponent on depth will be high and that on width low. In contrast, a sand-bed channel with weak banks will accommodate most of the increased bankfull

Table 26 Selected values of exponents in the equations of hydraulic geometry of river channels

Channel locality and type	At-a-station values			Downstream values		
	b	f	m	b	f	m
Mid-west USA (Leopold and Maddock 1953)	0.26	0.40	0.34	0.50	0.40	0.10
Mid-west USA (Carlston 1969)				0.46	0.38	0.16
Ephemeral streams, semi-arid USA (Leopold and Miller 1956)	0.25	0.41	0.33	0.50	0.30	0.20
Upper Salmon River, Idaho (Emmett 1975)				0.54	0.34	0.12
R. Bollin Dean, coarse-bed cohesive banks (Knighton 1974)	0.12	0.40	0.48	0.46	0.16	0.38
British gravel-bed rivers (Charleton *et al.* 1978)				0.45	0.40	0.15
Columbia River, Canada, canal-like anastomosed sand channels, cohesive banks (Tabata 2002)	0.10	0.66	0.24			

Notes: $W = aQ^b$, $D = cQ^f$, $V = kQ^m$

discharge downstream by widening the channel and the exponent for depth will be correspondingly low. Table 26 shows typical exponent values in the downstream hydraulic geometry equations for a variety of channel types.

Comprehensive reviews of all aspects of the hydraulic geometry of natural stream channels are available in fluvial geomorphology texts such as those by Knighton (1998), Richards (1982) and Leopold *et al*. (1964).

Limitations of hydraulic geometry

The power of conventional hydraulic geometry is its facility to generalize the process of channel adjustment in terms of simple functions so that the morphology of channels can be compared readily, one with another. This facility is also its primary limitation: it ascribes to the channel adjustment process the simple behaviour of simple functions which in detailed reality, however, may be very complex. Given the typical statistical noise in measurements of hydraulics and morphology of natural channels, power functions provide a simple and robust model for the hydraulic geometry but there is no independent theoretical justification for their use. The power-function model is merely a convenient approximation of reality.

Geomorphologists who recognize the limitations of power functions and have suggested hydraulic geometries based on alternative linearizing models (such as the log-quadratic model) include Richards (1973), Knighton (1975) and Ferguson (1986). Still others have taken multivariate statistical approaches to characterizing hydraulic geometry (Bates 1990; Rhoads 1992).

All these mathematical models of hydraulic geometry imply, however, that the channel adjustment process is continuous when in reality it is often markedly discontinuous. For example, many channels exhibit channel-in-channel morphology reflecting the capacity of low flows to shape the basal boundary (long duration offsetting low flow magnitude). Others exhibit within-channel benches reflecting the unequal effectiveness of higher discharge ranges of the hydrologic regime in shaping the channel (Woodyer 1968). Further, although low discharges essentially flow over rigid boundaries in many channels, increases in discharge at a section eventually lead to the onset of sediment entrainment and channel scour and this fundamentally alters the morphodynamics of the channel. Indeed such scour-related discontinuities in channel adjustments to changes in discharge may be an important part of the adjustment regime but may be completely obscured by the use of hydraulic geometry (Hickin 1995).

In the case of downstream hydraulic geometry the point-source addition of tributary flow and sediment to the fluvial system is a fundamental spatial and process discontinuity that is essentially a step-function process, not the continuous adjustment approximated by power functions.

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SEE ALSO: channel, alluvial; fluvial geomorphology

EDWARD J. HICKIN

HYDRO-LACCOLITH

A mound of ice formed by frost heaving of frozen underground water, resembling a laccolith in section. The term hydro-laccolith is synonymous to the terms ice laccolith and PINGO. However, they differ from pingos in that they are seasonal forms (whereas pingos are perennial), and differ from ice laccoliths in that they do not form within the active layer of permafrost ground. Hydro-laccoliths range in size between 1 and 10 m diameter, and are usually less than 2 m in height.

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SEE ALSO: periglacial geomorphology

STEVE WARD

HYDROCOMPACTION

The compaction and reduction in volume of soils and sediments that occurs when their moisture content is increased. It is also known as ‘collapse compression’, ‘hydrocompression’, ‘hydroconsolidation’ and ‘saturation shrinkage’ (Charles 1994). The process causes ground subsidence when unconsolidated sediments of low density are wetted, as for example by the application of irrigation water. It is a feature of arid and semi-arid lands where materials such as wind-blown loess or certain alluvial sediments above the water table are not normally wetted below the root zone and have high void ratios. When dry, such materials may have sufficient strength to support considerable effective stresses without compacting. However, when they are wetted, their intergranular strength is weakened because of the rearrangement of their particles. The associated subsidence may create fissures in the ground and is a process that needs to be considered during the construction of canals, pipelines, dams and irrigation schemes (Al-Harthi and Bankher 1999).

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A.S. GOUDIE

HYDROLOGICAL GEOMORPHOLOGY

Hydrological geomorphology is literally the interface between hydrology, the science of water and geomorphology, the study of landforms and their causative processes. It is particularly surface water hydrology that interacts with geomorphology although recently there has been an increasing convergence between research in geomorphology and in groundwater hydrology and hydrogeology (Brown and Bradley 1995).

When geomorphology emerged, more than a century ago, with a focus upon the morphology of landscape and the study of landforms, the attractive CYCLE OF EROSION promulgated by W.M. Davis provided a focus for many approaches to geomorphology adopted in the first half of the twentieth century. In the second half of the century, investigation of processes prevailed to a much greater extent (Gregory 2000) and to achieve this in relation to research on land areas, involving the study of flowing water, there was a need to take a much greater interest in hydrology. At first a period of familiarization saw the publication of books written by geographers (e.g. Ward 1966) and geomorphologists (e.g. Gregory and Walling 1973) but then progress was made towards the research interface between geomorphology and hydrology where the geomorphologist could make significant contributions. Hydrology was for long the science of water, with comparatively little attention given to water quality, but increasing attention given to landscape-forming processes, and to hydrological influences upon those processes, naturally led to original and innovative contributions being made by geomorphologists at the interface of hydrology with geomorphology.

Many interface investigations have accounted for the growth of hydrological geomorphology and these include at least four types. In addition to relationships between drainage basin characteristics and basin hydrological response, geomorphologists have made particular contributions in the investigation of *runoff producing areas* and the dynamic ways in which such areas contribute to the generation of stream hydrographs, including networks of subsurface pipes as well as headwater drainage systems and the modelling of their role in runoff production (Beven and Kirkby 1993). Such investigations often employed results from small experimental catchments which were also useful in relation to research on *sediment dynamics*. Geomorphological interest in the sediment area arose from the requirement for rates of erosion or denudation to relate to landscape development. In the case of suspended sediment production and transport, whereas simple rating curves relating discharge and suspended sediment concentrations had previously been used for analysis, it was demonstrated how analysis of sediment hydrographs could be employed to advance understanding and explanation of the mechanics of erosion. Later research by geomorphologists could therefore focus on the mechanics of production of such sediment in relation to the range of sediment sources and sediment producing areas. Similar contributions, made by geomorphologists to investigate and refine relationships employed to model bedload and solute transport, benefited from studies of the generation of solutes from catchment areas and of the entrainment of bedload in different channel situations.

Contributions at the level of the drainage basin have arisen first because of interest in the *temporal changes* that have occurred. As such changes cannot be based entirely on continuous hydrological records used in hydrology, other techniques are necessary to reconstruct past hydrological changes and these are used in geomorphology. The approach of palaeohydrology (Schumm 1965), which has been defined as 'the science of the waters of the earth, their composition, distribution and movement on ancient landscapes from the beginning of the first rainfall to the beginning of continuous hydrological records' (Gregory 1996), has been developed significantly so that the broad picture of past hydrological changes has been reconstructed for different parts of the world (Benito and Gregory 2003). Such reconstructions can provide potentially useful background for studies of global change and of basin management. A particularly valuable and successful approach has been the investigation and analysis of PALAEOFLOODS based upon the recognition of remnants of flood deposits, often as slackwater deposits, because this affords information on flood frequency which extends well beyond the period of instrumental record but may significantly affect the way in which flood frequency is analysed and relationships are established (Baker 2003). Partly as a result of the results obtained from investigations of temporal change, geomorphological contributions have become significant in relation to *river basin management*. Approaches are increasingly required to be integrated and should therefore include consideration of the range of human impacts throughout the basin (Downs *et al.* 1991) but in addition there is a need for sustainable approaches both at the basin level (NRC 1999) and in relation to the restoration of specific river reaches (Brookes and Shields 1996).

A number of the salient contributions made by geomorphologists have illuminated understanding of particular components of the hydrological cycle so that aspects of a hydrological geomorphology have emerged. This has, paradoxically, led to the fudging of the original definition of geomorphology. No longer focused primarily upon landforms, geomorphology is involved in contributions to hybrid fields where some of the most innovative research occurs and where multidisciplinary approaches can be optimized. Thus hydromorphology was suggested by Scheidegger (1973) as the geomorphological study of water and its effects, which includes coastal as well as fluvial hydrogeomorphology, in which there is a range of ways in which applications can be made (Gregory 1979). Such multidisciplinary fields also include biogeomorphology and, although there is no precise definition of it, hydrological geomorphology is an area of interaction which continues to offer promising research and applied opportunities.

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KENNETH J.GREGORY

HYDROLYSIS

Hydrolysis is a chemical reaction of a compound with water. As opposed to HYDRATION where the water is absorbed into the compound, in hydrolysis (or 'splitting by water') both the water and the compound split up and recombine. The water is thus a reactant and not merely a solvent. For example, the reaction between a potassium-containing feldspar and water:

$$\underset{\text{feldspar}}{KAlSi_3O_8} + \underset{\text{water}}{H_2O} \rightarrow \underset{\text{silicic acid}}{HAlSi_3O_8} + \underset{\text{potassium}}{K^+} + \underset{\text{hydroxyl}}{OH^-} \qquad (1)$$

The mineral releases potassium and the water splits into OH^- and H^+, the H^+ combining with the aluminosilicate from the mineral. The production of hydroxyl (OH^-) ions in solution means that the pH of the water rises. This is illustrated by grinding a mineral to powder in water and measuring the pH or 'abrasion pH'. For the more reactive minerals, this is between 8 and 11, with 8 for calcite and feldspars 8–10.

This reaction can take place in pure water at neutral pH (7). However, if the water is acidified by additional H^+ ions so the pH is below 7, the weathering reaction is accelerated. The prevailing form of acidification is by carbon dioxide:

$$CO_2 + H_2O \rightarrow H_2CO_3 \rightarrow H^+ + HCO_3^- \qquad (2)$$

with, for calcite:

$$CaCO_3 \rightarrow Ca^{2+} + CO_3^{2-} \rightarrow Ca^{2+} + 2(HCO_3^-) \qquad (3)$$

The mineral has combined with the constituents of water, giving a free mineral ion in the water (Ca^{2+}) with one source of HCO_3^- from OH^- in the water and CO_2 and the other source of HCO_3^- from the H^+ from the water in Equation 2 combined with the CO_3^{2-} from the calcite.

Hydrolysis is thus a fundamental weathering process and it can be readily appreciated that as there are many organic sources of CO_2 through respiration and decomposition, then many such reactions are biologically originated.

STEVE TRUDGILL

HYDROPHOBIC SOIL (WATER REPELLENCY)

A soil that resists wetting by water for periods ranging from a few seconds up to days or even weeks. The reduced affinity for water is caused by a coating of long-chained organic molecules on soil particles and/or by the presence of hydrophobic (water repellent) interstitial matter. Such matter is released from a wide variety of plants through mechanical wear from leaf surfaces, decomposition of litter, release via roots and vaporization followed by condensation onto soil particles during burning, or from soil fungi and micro-organisms. The effect is temporally variable and is usually most pronounced after prolonged dry spells. Although mostly associated with semi-arid and areas with Mediterranean-type climates, it is now known to occur in a wide range of climates, including temperate and arctic-alpine environments. The potential geomorphological impacts include the restriction of soil water movement to preferential pathways; increased OVERLAND FLOW; enhanced streamflow responses to rainstorms; enhanced total streamflow; enhanced splash detachment by raindrop impact (see RAINDROP IMPACT, SPLASH AND WASH); increased SOIL EROSION by both wind and water; and increased erosion by dry creep (movement by loose, dry surface material on steep slopes). In contrast, water-repellent organic material in well-developed soil aggregates can help to stabilize them, thereby reducing soil ERODIBILITY. These impacts, however, have been largely inferred rather than demonstrated under field conditions.

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SEE ALSO: fire

RICHARD A. SHAKESBY

HYPERCONCENTRATED FLOW

A flowing mixture of water and sediment transitional between a debris flow and muddy streamflow. The terms hyperconcentrated flow, hyperconcentrated flood flow, and hyperconcentrated streamflow are synonymous. The term was originally used for streamflow with sediment concentrations between 40–80 per cent by weight or 20–60 per cent sediment by volume. Rheologically the fluid appears to be slightly plastic but flows like water (Pierson and Costa 1987). Such flows are gravitationally driven, non-uniform mixtures of debris and water. They possess fluvial characteristics yet are capable of carrying very high sediment loads. Hyperconcentrated flows show clast support from grain-dispersive forces, dampened turbulence and buoyancy (implying yield strength). Sediment deposition appears to be by rapid grain-by-grain settling at the base and margins of the flow. Resultant deposits are usually either massive or display weak, near-horizontal stratification. Hyperconcentrated flows are common in volcanic environments where eruptions release large volumes of water from crater lakes or from melting of ice and snow, and when debris flows evolve downstream into hyperconcentrated flows.

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VINCENT E. NEALL

HYPSOMETRIC ANALYSIS

Hypsometric analysis is the study of the distribution of topographic surface area with respect to altitude. The area-altitude relationship is described by the hypsometric curve (hypsographic curve) that is expressed by the function $y = f(x)$. In its absolute formulation, this curve is obtained by plotting on ordinate elevations and depths, from the top of the highest mountain to the maximum depth of abyssal trenches, and on abscissa the values of topographic surface areas. It is a cumulative curve: the abscissa of any point on it expresses the total area lying above the elevation of the corresponding ordinate.

The *absolute hypsometric curve* can be constructed for any area of land, from a small portion to the entire planet. Its use, however, is unsatisfactory when it is necessary to compare areas of different sizes and relief. To overcome this difficulty, percentage hypsometric analysis can be used, as it affords a method for expressing the area-altitude relationships in a dimensionless form (Langbein 1947).

The *percentage hypsometric curve* was used by Strahler (1952) to analyse erosional topography of drainage basins that are the basic geomorphic unit. This curve is represented by the function $y = f(x)$, but x and y are dimensionless parameters: x is the ratio of the area a above a given contour line and the whole basin area (A) and y is the ratio of the height h between the basin mouth and the contour that defines the lower limit of the area *a* and the total height range in the basin (*H*). Obviously, x and y vary between 0 and 1. These curves can be compared irrespective of true scale as they express merely the way the landmass is distributed from base to top.

Integrating the function between the limits of $x = 0$ and $x = 1$ (or simply measuring on the graph the area under the curve) the *hypsometric integral* is obtained. It is expressed in percentage units and represents the ratio of the landmass volume of a given drainage basin to the volume of the reference solid with base equal to the basin area and height equal to the total height range in the basin. In other words the hypsometric integral measures the percentage volume of earth material remaining after the erosion of an original landmass having volume equal to the reference solid.

In their classical interpretation, Strahler's hypsometric curves and integrals identify quantitatively the stages of the Davisian geomorphic cycle. Convex curves, with hypsometric integrals higher than 0.60, indicate the inequilibrium stage of 'youth'. Smoothly S-shaped curves that cross approximately the centre of the diagram and have integrals ranging from 0.60 to 0.40 express the equilibrium stage of 'maturity' or the 'old stage'. Strongly concave curves with very low integrals result only where monadnock masses are present.

Further studies delineated another interpretation of the area-altitude analysis: the hypsometric curves express not only the stage of the 'geomorphic cycle' but also the complexity of denudational processes and the rate of the geomorphological changes in drainage basins. Such changes take place through subsequent stages of dynamic equilibrium between tectonic uplift and DENUDATION (Ciccacci *et al.* 1992); therefore each basin is marked by a hypsometric curve which is mostly a function of the denudational process type. Convex curves with a high integral refer to basins in which stream erosion is the most vigorous denudational process. Concave curves with a low integral mark basins mainly affected by intensive slope processes. Finally, hypsometric curves with an integral close to 0.5 are characteristic of basins where stream erosion balances the effectiveness of slope processes.

Actually, the classic interpretation of the hypsometric curves matches the morphodynamic characters of drainage basins in tectonically stable regions; the same interpretation is rather unsuited to explain the plano-altimetric configurations of regions affected by recent or active tectonics (Ohmori 1993; D'Alessandro *et al.* 1999).

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ELVIDIO LUPIA-PALMIERI

I

ICE

Ice is the solid phase of water, i.e. a chemical compound of two positive univalent ions of hydrogen and one negative bivalent oxygen (H_2O). Water and ice are responsible for most of the processes leading to landscape sculpturing. In water an isosceles triangle is formed by the three nuclei of the H_2O and the apex angle is equal to $105^\circ 3'$. The oxygen nucleus is at the apex and the hydrogen nuclei (protons) at the other two corners of the molecule. Two of the eight electrons of the oxygen atom rotate close to the nucleus and another two have eccentric orbits which also contain the electrons from one of the hydrogen atoms. The oxygen nucleus and one proton are enclosed by each of these orbits. The other four electrons circle in two other eccentric orbits. For this reason, four eccentric orbits radiate tetrahedrally from the oxygen nucleus, which is completely screened by the electron orbits. Some of the positive charge of the protons is not screened completely. The eccentric electron orbits provide an excess negative charge in the direction of the two orbits without protons. Small negative charges at the oxygen end of the molecule and equally small positive charges at the hydrogen end determine that the H_2O molecules are slightly charged. Thus, a weak electrostatic bond is formed between the hydrogen end of one molecule and the oxygen end of another (the hydrogen bond) and it is for this reason that each molecule of water is surrounded by four others in a regular tetrahedral spatial arrangement. With respect to any particular molecule, the hydrogen bond is more than ten times weaker than the covalent bond. Therefore, mixtures of molecules are easily formed and broken by the reduction or addition of energy. In the liquid phase, the molecules are in motion and can be of variable distances, one to another, but, in ice, stable hydrogen bonds are created between the surrounding molecules, thereby generating the hexagonal crystal structure (Figure 90a, b). This is because in negative temperatures the energy of the system is lower. It gives pure ice a symmetrical lattice arrangement and a lower density ($0.91668\,\mathrm{g\,cm^{-3}}$) than pure water ($0.999984\,\mathrm{g\,cm^{-3}}$ – at the temperature 0 °C and the normal atmospheric pressure 1013 hPa). The molecules within the ice crystal are organized in layers of hexagonal rings. The atoms in a ring are not in one plane but two (Figure 90b). Spacing between two layers (0.276 nm) is much larger than between two planes of atoms (0.0923 nm). Adjacent layers form mirror images of each other. The optic axis (also called the *c-axis* or *principal axis*) of the ice crystal is perpendicular to the *basal plane*, i.e. the plane of a layer of hexagonal rings of atoms. Within the hexagonal lattice, three crystallographic *a-axes*, separated by 60° from each other, form the basal plane (Figure 90a, c).

A dozen crystallographic kinds of ice are known in nature and laboratory experiments but only two are observed in the Earth's natural conditions. The most common is the hexagonal crystalline ice which exists in temperatures between 0 °C and *c.* −70 °C and pressures up to *c.*210 MPa. Ice in the cubic crystalline form has been found at very low temperatures (below *c.* −70 °C) and very low pressures in the upper parts of the troposphere. Other ice polymorphs can appear in very high pressures (higher than 800 MPa) and at various positive temperatures. Such 'hot ices' can reach density higher than $1.3\,\mathrm{g\,cm^{-3}}$ and might be present in the lithosphere as films of solid water, only a few molecules thick, chemically bonded

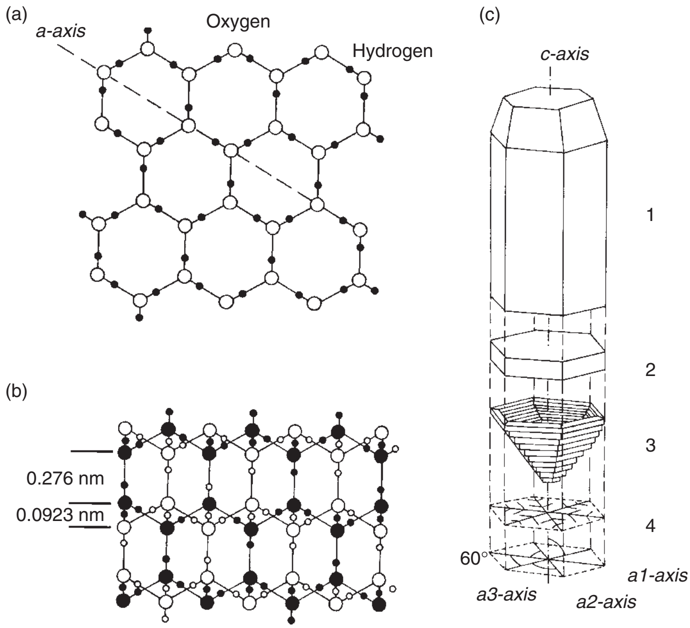

Figure 90 Structure of ice and forms of crystals (modified from Jania 1988). Models of the ice lattice structure: (a) basal plane, (b) view perpendicular to the c-axis. Types of the hexagonal ice crystals (c): 1 – column (prism), 2 – plate, 3 – cup, 4 – snow star

with other minerals. It is thought that these may affect certain properties of rocks, e.g. their susceptibility to weathering after exposure at the land surface.

The increase of volume of freezing water, by *c.*10 per cent, is responsible for important geomorphic processes (e.g. MECHANICAL WEATHERING; SOLIFLUCTION). The higher specific volume of the 'common' hexagonal ice compared with water, especially in respect of the relatively more dense sea water, enables ice to flow. Floating ice can be also a geomorphic agent (e.g. ICEBERGS). *Drift ice deposits* resulting from melting of debris-rich icebergs are recorded not only in deep sea deposits but also in shallow bays and proglacial lakes.

Six isotopes of the oxygen and three isotopes of hydrogen have been noted in natural conditions. Their various combinations result in thirty-six isotopic kinds of water and ice. The most important are: $H_2{}^{16}O$ (common water); $HD^{16}O$ ('heavy water' with deuter – D) and $H_2{}^{18}O$ (water with 'heavy' oxygen – ^{18}O). The last two are minor admixtures (0.03 per cent and 0.2 per cent, respectively) of ocean waters which are the most important source of water vapour in the atmosphere and consequently of snow and ice on land. Molecules of water with the heavier stable isotopes evaporate less rapidly and condense slightly faster from the vapour. The isotopic composition of snow precipitation depends on the temperature of evaporation of the sea water and the temperature of condensation (re-sublimation). The proportion of contents of heavy isotopes is a fraction of their concentration relative to the 'light' isotopes ($^{18}O/^{16}O$ or D/H) in a sample of snow (ice). In respect of its deviation (δ) from the ratio in the 'standard mean ocean water' (SMOW) it is expressed in parts per thousand (‰). The average annual values of $\delta^{18}O$ in snow samples in the polar areas correlates closely (in linear regression) with mean annual air temperatures there. It follows that the stable isotope content in samples from deep ice cores is a good source of data on air temperatures in the past (paleotemperatures).

Despite the fact that the hexagonal crystalline ice (Figure 90c) is the most common on the Earth, perfectly symmetrical hexagonal ice crystals are in the minority in the natural environment. Varied factors, which influence crystallization of water, determine that ice crystals are often incomplete or defective. First, natural waters are very rarely chemically pure. They have dissolved admixtures which lead to the phase change (crystallization) at temperatures lower than 0.16 °C – the figure usually regarded as the freezing temperature. Water in a supercooled state (i.e. being in the liquid form below the freezing point temperature) rarely freezes spontaneously. It is able to freeze only when crystallization centres are available within the liquid (e.g. small particles of ice or other minerals). When water freezes, H_2O molecules arrange themselves in a lattice of crystals. When cooling is slow in still waters, other molecules are removed to outside the walls of the ice crystals. Fast crystallization and turbulent movement of water cause incorporation of molecules of other chemical species into the lattice of ice crystals and their structure becomes defective. In general, optic axes of ice crystals are oriented perpendicularly to the 'cooling front' or the freezing surface. Vertical ice crystals are characteristic of lake ice and *naled ice*, whereas *rime ice* may have c-axes oriented toward the wind direction. *Glacier ice* and the anchor and frazil river ice usually have randomly oriented crystal optic axes.

Defects within the crystal lattice are regarded as factor facilitated recrystallization and ice crystals become reoriented when new directions of shear stress are applied. Under high cryostatic pressure, atmospheric gas molecules (e.g. N_2, O_2, CO_2) from air bubbles, trapped between glacier ice crystals, are incorporated into their lattice in the form of clatrates. When such ice crystals appear on the surface or in front of a glacier, where pressure is lower, these gas molecules are released from the lattice to either meltwater or the atmosphere.

Ice under natural conditions may variously be regarded as a mineral, sediment or a rock (Shumskyi 1955: 15–16). The hardness of ice varies with temperature. At 0 °C, its hardness is 1.5 on the Mohs' scale (as is talc and gypsum), whereas at −40 °C, the ice hardness is as much as 4 (equivalent of the hardness of fluorite). It is clear that glacier erosion would be impossible without the regelation process at the glacier bed, i.e. melting (under pressure) at the proximal side of subglacial obstacles and refreezing in their lee side.

Ice is common on the planet Earth and is also present on Mars, Pluto and satellites of Jupiter (e.g. Europa, Callisto) and Saturn. However, the presence of water in vapour, liquid and solid forms makes the Earth a planet unique to the Solar System. An irregular envelope containing ice encircles the Earth and is called the cryosphere. The term, introduced by Dobrowolski (1923), derives from the Greek (*kryos* – cold). The cryosphere exists only in an intermingled state with the lithosphere, hydrosphere and atmosphere, where water appears in the form of ice (snow cover, glaciers, sea and fresh waters ice, ground ice and ice in the atmosphere). The total mass of all ice on the Earth is estimated at 2.5×10^{16} metric tons. This is present on the surface or in the near-surface layers of the crust of our planet and occupies an area of *c.* 73.4×10^6 km^2. It covers 14.2 per cent of the total area of the globe with annual fluctuations between 10.5 and 17.9 per cent. Ice, therefore, can hardly be other than a vastly important geomorphic agent, especially in polar and temperate climates. Considering all forms of natural ice, GLACIERs are the most important agents of relief changes. Total volume of glaciers and ice sheets constitutes about 97.7 per cent of all natural forms of ice on Earth, while subsurface ice is only 2.1 per cent (Kotlyakov 1984: 347–348).

Atmospheric ice, in the form of stars, plates or columns (Figure 90c), forms a snow cover when deposited on the ground. This mixture of ice crystals and their inclusions of air and water (in warmer environments) normally has a density between 0.05 and 0.5 g cm^{-3}. A seasonal snow cover affects slope processes, including rapid and spectacular mass movement as avalanches (see AVALANCHE, SNOW).

In areas where accumulation of snow exceeds its melting, snow is transformed into glacier ice. The chain of processes leading to the metamorphism of a fresh snow cover into this specific form of polycrystalline 'ice rock' differs in warm (temperate) environments (with presence of water within the snow cover) and in dry cold ones. Infiltration of meltwater into the snowpack accelerates the transformation. In cold conditions sublimation plays an important role in the rounding of crystals and their growth. Despite the environmental differences, four steps of transformation of snow into polycrystalline glacier ice are usually distinguished: the first is from fresh snow (porosity *c.* 95 per cent) to old snow (density 0.3–0.5 g cm^{-3}) due to setting, compaction and rounding of crystals (step 1); further densification and compaction by pressure of overlying layers leads (step 2) from old snow (porosity *c*.50 per cent) into coarse-grained snow called *firn* (density 0.55 g cm^{-3}) about a year later. Firn has a low porosity, but is still permeable. Transformation from firn to the white (new) glacier ice (step 4) takes a longer time in cold climates, where intergranular movements (due to compaction), combined with sintering (recrystallization by way of sublimation) and plastic deformation, dominate. In warmer climates, owing to liquid water infiltration, metamorphosis is faster; this is due to refreezing and, of course, settling. In extremely cold and dry conditions, such processes takes hundreds or thousands of years (*c*.4,000 years at the Vostok Station, East Antarctica). In temperate and subpolar climates, it lasts decades. New glacier ice has a density higher than 0.83 g cm^{-3}, and, owing to closure of the pores into air bubbles between ice crystals, very low permeability. At this stage, crystals usually have dimensions of a couple of millimetres and their c-axes are randomly oriented. The last step in the growth of crystals and densification of glacier ice occurs through dynamic metamorphism and requires the deformations caused by *glacier flow* (see GLACIER). The old glacier ice may have densities of as much as 0.9 g cm^{-3} and crystal sizes can have diameters larger than 10 cm (Plate 63). The optic axes of dynamically transformed glacier ice crystals reveal a predominant orientation, which is perpendicular to the terminal stress field. However, the *basal ice* of glaciers possesses a different structure. Most glaciers have an irregular bed with resistant rock protuberances. Melting develops when the sliding sole of a glacier passes such an obstacle in its bedrock (in thermal conditions close to the pressure melting point of ice). On the lee side of the obstacle, water refreezes, forming a layer of REGELATION ice and basal debris can be incorporated into this. The formation of debris-rich basal ice is observed in many temperate and polythermal glaciers. The isotopic composition of basal ice layers suggests an accretion of ice crystals from supercooled water from the subglacial drainage (Titus *et al.* 1999: 43). The debris content incorporated into basal ice (a basal moraine) is different, is very variable and depends on a number of factors. The basal debris zone tends to be thicker in the cold-based ice sheets of Greenland and Antarctic (up to16 m) than in temperate or polythermal glaciers, where its thickness varies from 0.4 m to a couple of metres. Debris concentration in the thick dispersed basal debris zones reaches 7–12 per cent (by volume), whereas in the thin basal zones (<1 m) of warm-based glaciers, the concentration can exceed 50 per cent (Menzies 2002: table 6.2). In contrast, the englacial concentration of debris is generally very low (<1 per cent). Glacier sediment discharge depends on the debris concentration in the basal ice zone and the basal sliding velocity.

An ICE SHEET is a shield-like, broad ice mass of continent scale (larger than 50,000 km^2). It can be thick enough (thousands of metres) to cover any irregularity in topography of its bed completely. Ice flow is generally organized in a quasi-radial pattern from one or more centres: *ice domes*. The Antarctic Ice Sheet is the contemporary example ice mass of such dimensions: *c*.13×10^6 km^2, with volume of 30.11×10^6 km^3. The ice cover of Antarctica has five ice domes which reach an elevation of 4,000 m a.s.l. The maximum thickness of ice (4,776 m) has been found in East Antarctica and its mean thickness is 2,160 m (Drewry 1983: 4). The Greenland Ice Sheet covers *c*.1.75×10^6 km^2 and has a mean ice thickness of 1,790 m (volume: 2.74×10^6 km^3). Two well-pronounced ice domes are distinguished in Greenland and the northern dome reaches an elevation of 3,236 m a.s.l. (Van der Ween 1999). An ice dome has a

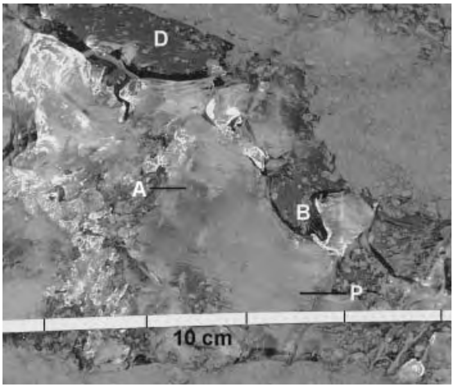

Plate 63 Large crystals of the glacier ice from bottom layers of Horn Glacier, Spitsbergen (intervals of 10 cm are marked). Note inclusions of air bubbles (A), cross section of the crystal basal planes (P), plastically deformed crystal boundaries (B) and presence of debris admixtures (D). Photo by Jacek Jania

convex shape and ice flow is predominantly vertical (downward) within it.

An ICE STREAM is a zone within an ice sheet or ice cap in which the ice flows significantly faster than on its sides. Theoretically, an ice stream has no rock boundaries; however, the course of the majority of ice streams is determined by their bedrock topography. Antarctic ice streams drain up to 90 per cent of mass accumulation in the interior (Paterson 1994: 301). The best known ice streams are on Siple Coast in Antarctica (Ice Streams A, B, C, D and E) and Jakobshavn Glacier, West Greenland. The measured velocities of Ice Stream B exceed 800 m a^{-1} whereas on the neighbouring ice sheet area they are less than 10 m a^{-1} (Van der Ween 1999). The surface velocity of Jakobshavn Glacier reaches almost 8 km per year near its terminus (Fahnestock *et al.* 1993: 1,532). In both cases, basal sliding is responsible for the fast flow of ice. High basal velocities of ice streams are thought to produce intense erosion of glacier beds.

In PERMAFROST areas several kinds of ground ice are present and ice bodies have a different origin, size, shape and internal structure. They might reasonably be classified into two genetic groups: (1) those originated from ground water and (2) those from water migrating from the surface to the ground. Within the first group, the largest ice masses are called hydro-laccolites. Their thickness can be as much as 20–40 m and their diameters can be dozens or even hundreds of metres (e.g. PINGO). Smaller forms are usually lenticular. *Ice lenses* appear in relatively dry ground or in sediments which have low permeability (e.g. tills). Owing to mineralization of water, which migrates within the sediments and to the higher than atmospheric pressure within the ground, crystallization of ice within the ground takes place in temperatures below the freezing point. Ice lenses grow upwards and have a horizontally laminated structure with inclusions of mineral particles. Ice wedges (see ICE WEDGE AND RELATED STRUCTURES) are the most common forms of the various ice bodies which result from water infiltration.

Ice jams on rivers most often take place in the winter and spring season and are a notable feature of large rivers which flow northwards in the northern hemisphere (Siberia, Canada). They are often caused by the earlier melting of snow in the southern parts of drainage basins and blockage of water by the longer preserved river ice cover in the northern reaches. Jamming of ice transport occurs as it lodges beneath the superficial ice cover of the frazil and anchor ice and thereby drainage blockage is typical for winter periods. The impediment may be so severe that wide areas of the adjacent floodplains become inundated. Detached drifting ice blocks may locally cause erosion of the floodplain sediments.

Specific ice masses originate in caves. The *ice cave* usually has a thick layer of laminated infiltration ice and icicles (ice stalactites and stalagmites). Ice caves are common in KARST areas within a permafrost zone. They are also known from locations where ground is not perennially frozen. In such cases, the very specific microclimate of the cave is responsible for the formation of annual infiltration ice layers: there is only one entrance located in the upper part of the cave and during winter, cold air flows down through the cave, while lighter warm air in spring and summer cannot penetrate into the system. Percolated meltwaters from snow or precipitation are frozen on the floor of the ice cave.

Tunnels formed in glaciers by englacial and subglacial drainage systems are called glacier caves. Vertical tunnels which transfer water from the glacier surface to subglacial channel are termed *ice shafts* or MOULINs. Ice shafts develop on planes of discontinuity within the glacier ice as crevasses and shear planes.

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SEE ALSO: glacier; ice sheet; ice stream; ice wedge and related structures; permafrost

JACEK JANIA

ICE AGES (INTERGLACIALS, INTERSTADIALS AND STADIALS)

Ice ages have occurred throughout geological time: as, for example, during the Precambrian, the Ordovician, Permo-Triassic and the Cenozoic (Tertiary), of which the Quaternary is a part. During the Permo-Triassic ice age, the ice sheets were centred on the Antarctic continent, located across the southern polar area. Its principal glacial deposit, the Dwyka tillite, is found in South America, southern Africa, southern India and south Australia, a distribution that shows the fragmentation of the super continent (Gondwanaland) where glaciation occurred by 'continental drift'. The Cenozoic ice ages commenced when Antarctica experienced major glaciation about 34 Ma ago (Ma = millions of years). But, in a geomorphological sense, the term 'ice age', incorporating many ice ages and interglacials, most appropriately refers to the Quaternary ice ages, when major geomorphic changes occurred.

Evolution of the present boundary conditions of the Earth's climate, namely the location of the continents, seaways and mountain ranges, closely match the evolution of the climate system and ice ages. As the southern continents 'drifted' away from Gondwanaland, India collided with Asia to produce the largest landform on Earth, namely the Tibetan Plateau, a feature so high that it modulated the circulation of the high level westerlies that led to waves on the circum polar vortex. These drew cold arctic air to Canada and Europe. Cold continents adjacent to warm oceans were the ideal combination for initiating the major Quaternary ice ages. Once the closure of the Straits of Panama (about 4 Ma) had occurred, a zonal circulation of ocean water between the Atlantic and Pacific ceased, to be replaced by meridional warm water circulation in the Atlantic – the provider of precipitation to grow the ice sheets – and the start of the intensification of North Atlantic Deep Water formation, the driver of oceanic thermohaline circulation.

The standard definition of the Quaternary is that it commenced 1.8 Ma ago, a view tied to a stratotype (type-site) at Vrica, near Crotone, Calabria, Italy. This horizon lies at the top of the Olduvai magnetic reversal Subchron. But, an alternative view would place the start at about 2.5 Ma, because it coincides with the first record in marine sediments in the North Atlantic. This shows that ice sheets had grown large enough to reach tidewater lands where they launched icebergs into the ocean. When these melted, their load of ice rafted debris (IRD) was released onto the ocean floor. In China, it coincides with the contact between the 'Red Clay' and the first, Wucheng Loess; and also coincides with the base of the earliest, pre-Nebraskan, glaciation of North America.

The Quaternary ice ages are characterized by lowered temperatures when the snowline in mid-latitude regions was lowered by up to 1 km. This led to the growth of large mid-latitude ice sheets in Canada and Scandinavia, while the water they abstracted from the oceans lowered sea level by up to 135 m. But how did these large ice sheets grow, because they have no modern analogues? Three main theories have been proposed. (1) *Highland origin and windward growth*: based on the former Laurentide ice sheet in Canada and then applied to the Fenno-Scandian ice sheet. It is suggested that ice grew initially in the highland of Ungava and Quebec, then spread towards its source of precipitation from the Gulf of Mexico, before forming a large ice dome up to 3 km thick or more, centred over Hudson Bay. (2) *Instantaneous glacierization* involved rapid vertical growth of ice from large snow patches distributed over wide areas, including the low ground of Keewatin west of Hudson Bay. It led to a thinner multi-domed ice sheet, capable of rapid response to climate change. (3) *Marine-based ice sheets* are conceptually based on the West

Antarctic ice sheet and involved mid-latitude ice sheets centred on shallow sea areas such as Hudson Bay or the Baltic Sea, where large ice domes grew. These, drained by ice streams, were buttressed by floating ice shelves which, on destabilization, no longer provided support, so that the ice domes collapsed and surged into the ocean. Whatever the mode of ice sheet growth, it is clear that they were large enough to reach the tidewater regions of the North Atlantic Ocean on several occasions when they launched armadas of icebergs. On melting, these released the glacial material they were carrying which was deposited in the ocean as discrete layers of ice rafted debris (IRD). Such episodes are known as Heinrich Events and they occurred on an irregular 5 to 7 ka cycle.

South of the ice sheets vegetation zones were compressed towards the equator. Accompanying this was a migration of the fauna and flora and in geomorphic processes. Mid-latitudes experienced periglacial climates and atmospheric circulation was more vigorous as the consequence of steeper poleward climate gradients. Low latitudes experienced widespread aridity. Ice ages were diversified by brief climatic ameliorations called interstadials separated by colder stadials. Interglacials supported no mid-latitude ice sheets, had relatively high sea levels and a fauna and flora similar to the present (Holocene) interglacial. During the past million years or so, ice ages have commonly lasted about 100,000 years. But the length of interglacials appears to have been variable and is controversial. The interglacial about 400,000 years ago may have lasted for the best part of 60,000 years; but duration of the 'last interglacial' about 125 ka is controversial: according to some, it lasted about 10,000 years, but others believe it was twice as long. The current interglacial has already lasted 11,600 years, during which time several minor fluctuations in climate have occurred, such as the Little Ice Age.

Figure 91 shows that some fifty ice ages and fifty interglacials occurred in the last 2.5 Ma. This contrasts with earlier, classical, views that maintained that only four ice ages occurred in the 'Great Quaternary Ice Age', a view based on the record of Alpine glacial advances shown by four outwash terraces in Bavaria by Penck and Bruckner in 1910. Their view was reinforced in America and elsewhere by four major groups of glacial deposits.

Their paradigm lasted for over sixty years. But evidence of greater complexity came from an unexpected source, namely marine deposits of the deep open ocean, where long sequences recorded the ice ages of the entire Quaternary. These consist of muds composed of microfossils of plankton (planktic) and benthos (benthic) which secreted their calcareous shells in oxygen isotopic equilibrium with the ocean water they inhabited. Initially it was believed that variability in the ratio of ^{18}O and ^{16}O in foraminifera microfossils was primarily an indicator of the ocean temperatures in which the organisms grew, and only to a lesser extent was the isotopic composition of the ocean involved. Subsequently it was shown that the isotopic composition, not the temperature, of the ocean was the primary control on oxygen isotope

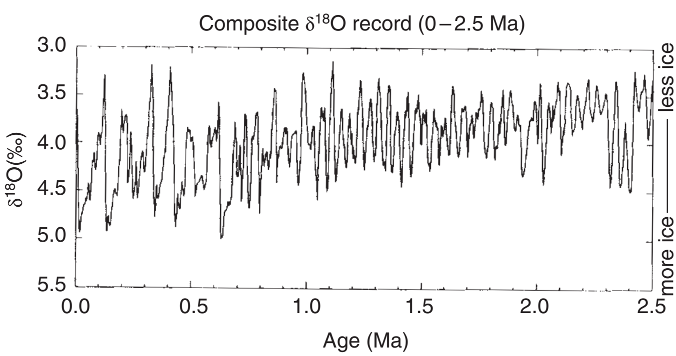

Figure 91 Composite record of the ice ages for the last 2.5 million years, based on oxygen isotope variability (from Crowley and North 1991)

ratio. Furthermore, because bottom water temperatures remained constantly cold, then benthic foraminifera would only record the oxygen isotopic composition of the ocean. Thus, because the primary control on this was the volume of isotopically light ice locked up on the continents during ice ages, then the benthic signal would provide an ice volume indication: that is, a record of the ice ages and interglacials. Moreover, because this was altered as the sea rose and fell as continental ice volume changed, the benthic signal was also one of sea-level change. Recently, it was shown that bottom water temperatures had varied; thus ^{18}O benthic ratios could no longer provide precise ice volume or sea-level information. Fortunately this was resolved by temperature determinations based on Magnesium/Calcium ratios in Ostracoda, that allowed discrete ice volume and temperature parameters to be identified separately.

Oxygen isotope signals are subdivided into *oxygen isotope stages* and *sub-stages*. Interglacials are numbered backwards in time with odd numbers, starting with the present (Holocene) interglacial as stage 1. Similarly ice ages are numbered backwards with time with even numbers, starting with the last ice age or Last Glacial Maximum as stage 2. Considerable confusion has been caused by the designation of an oxygen isotope stage 3 in the middle of the last 100 ka cycle. This arose because it was once thought that the 41 ka tilt frequency was the main ice age pacing (below). So in reality, the present (Holocene) interglacial is stage 1 and the last interglacial centred on ~125 ka is sub-stage 5e, the warmest part of stage 5, when conditions more or less corresponded with the present. Stage 3 is not an interglacial but merely a general climatic amelioration punctuated by millennial scale interstadials (below) during the last glacial cycle.

A means of providing a chronology for the ice ages and interglacials was provided by fixed points in the cores where changes in the Earth's magnetism occurred. Of these, the most important reversal is the Matuyama/Brunhes reversal at 0.78 Ma (780 ka). By assuming constant sedimentation, ages were provided for the ice age and interglacial boundaries. This was reinforced when it was found that the predicted age of the 'last interglacial' of 125 ka was supported by Uranium 234–Thorium 230 ages on uplifted coral reefs that showed a relatively high sea level at that time. Since then, the same has been confirmed for the age of the interglacial at about 300 ka, oxygen isotope stage 9. The last interglacial shoreline is found throughout the world as coral reefs, raised beaches or shore-platforms. Detailed records of sea-level fluctuations have been established in regions with uplifted coral reefs, such as Barbados, Tahiti and the Huon Peninsula of New Guinea.

Inspection of oxygen isotope signal reveals a number of cycles or pacings (Figure 91). These are the well-known pacings caused by the orbital movement of the Earth around the sun, and occur at 100 ka (eccentricity of the orbit), 41 ka (variation of the ecliptic or tilt of the Earth's axis), 23 ka and 19 ka (precession of the equinoxes that varies the distance between the Earth and the sun at the summer solstice). Relating the changing amount of solar energy received at given latitudes to the oxygen isotope (ice volume) record, however, had to await confirmation of these pacings in high sedimentation marine cores from the Indian Ocean. The astronomical and oxygen isotope records are not an exact match, because of a time-lag between orbital forcing and the climatic response as ice sheets grew or decayed to change the oxygen isotopic composition of the ocean. By using phase relationships (leads and lags), the oxygen isotope record has been 'tuned' by the predicted orbital changes. This controversial method is illustrated by the debate over the commencement age and duration of the 'last interglacial' (oxygen isotope sub-stage 5e) which Uranium–Thorium ages on stalagmite shows has been underestimated. It may be resolved because the shorter estimate is based on orbital calculations for 65°N (the 'sensitive latitude' for mid-latitude ice sheets), whereas the longer one is more consistent with orbital calculations for 65°S.

Despite ongoing debate about the precision of ice age and interglacial timing, it would seem that the origin of the ice ages has been discovered. Changes in temperature caused by changes in the orbit of the Earth seem to be the pacemaker of climate change. But how was orbital forcing transformed into actual climate change? Several questions remain before the matter can be settled. The earlier part of the Quaternary record displays a strong 41 ka tilt pacing with no trace of the 100 ka eccentricity pacing that only becomes distinct after 700 ka (Figure 91). This was when the large mid-latitude ice sheets grew and extended south to Ohio, London, Berlin and Kiev for the first time. What was the cause of the later

Quaternary 100 ka cycles? Perhaps further uplift of the Tibetan Plateau intensified the climate forcing? More likely is that continental erosion had progressively removed extensive layers of weathered rock to reveal more resistant bedrock. Low gradient ice sheets occur on soft deformable beds, but more resistant rocks support ice domes with steep marginal profiles. These depressed the Earth's crust, while continuing to receive more snow at their summits, thus prolonging the ice age. This may account for the transition from the 41 ka to the 100 ka world. Not unrelated to this is the large mismatch between the forcing provided by the 100 ka pacing and the response of the climate system (Figure 92). While there is a predictable and strong response of the climate system in the tilt and precessional bands, the disproportionate response in the eccentricity band remains unexplained. Could it be because of the development of large ice domes (above), or perhaps the real role of eccentricity is to modulate the tilt and precession forcings?

Amplification of orbitally forced climate change occurred by enhanced or decreased greenhouse gas concentrations of carbon dioxide and methane in the atmosphere. These are clearly implicated in climate change as is shown by measurements of past atmospheres preserved within bubbles of air trapped in ice layers of the Antarctic and Greenland ice sheets (Figure 93). Higher concentrations of greenhouse gases correspond with interglacials, while they are lowered during the ice ages. Oxygen isotope or Deuterium analyses of ice from cores through the Greenland and Antarctic (Figure 93j) ice sheets, provide a record of temperature changes at the surface of the ice sheet. But the age of the ice layers and the bubbles of air within them is not the same because air bubbles provide a greenhouse gas record at the time when they were sealed by the weight of overlying snow and firn. Differences of up to sixty years occur in high snowfall regions such as Greenland, but up to 1,200 years in the centre of Antarctica where only about 2 cm of snow falls

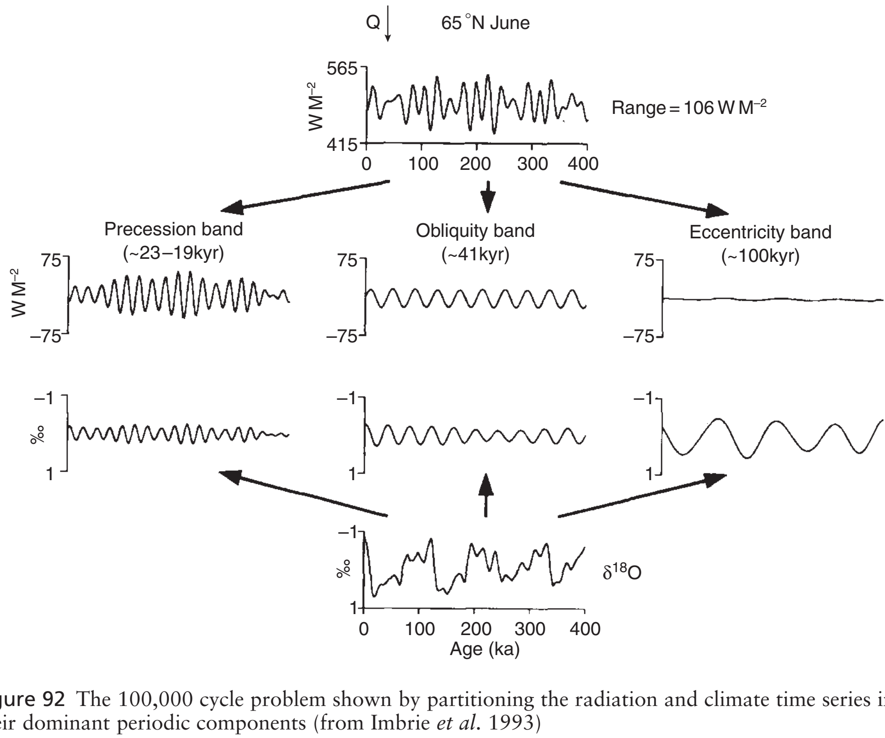

Figure 92 The 100,000 cycle problem shown by partitioning the radiation and climate time series into their dominant periodic components (from Imbrie *et al.* 1993)

annually. This may conceal important phase relationships between temperature and greenhouse gases. Fortunately the sharp methane spikes in the record allow good correlation between interhemispheric ice cores. The record of temperature and greenhouse gases now extends back more than 400,000 years at Vostok Station, Antarctica, through four major ice ages and five interglacials; whereas the record at Greenland Stations only extends back beyond the last interglacial.

Not only do palaeotemperature records from ice sheets reveal the main orbital pacings, they also display strong millennial ones (Figure 93). Moreover, unlike the sinusoidal shape of the

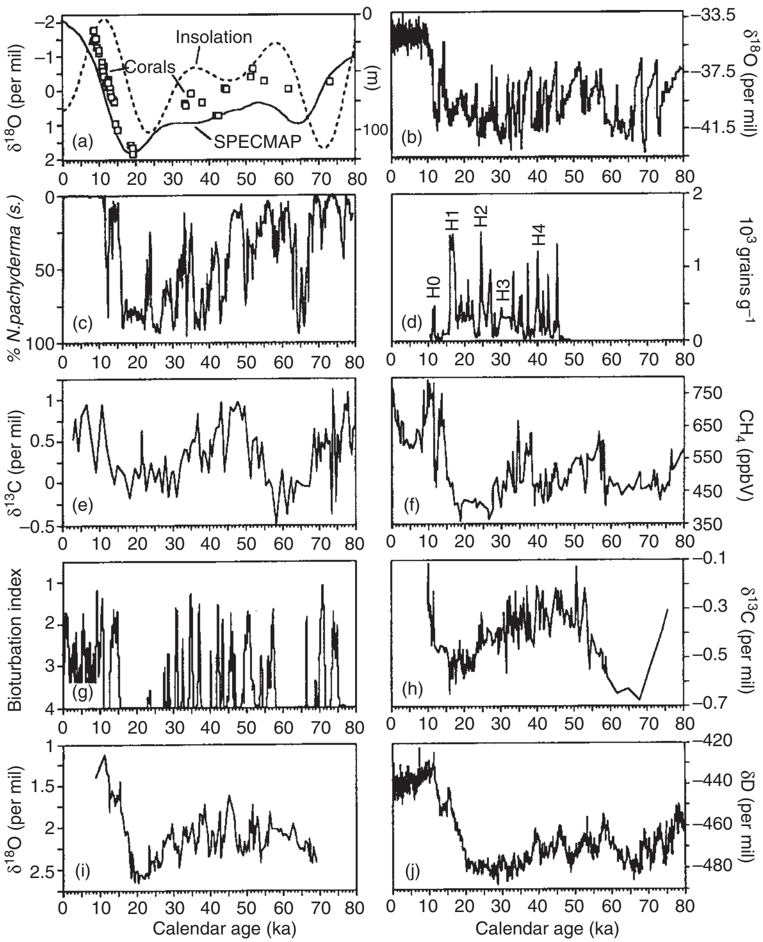

Figure 93 Millennial pacings (Alley and Clark 1999). (a) marine oxygen isotope signal (SPECMAP), insolation (orbital forcing) and sea level from dated corals; (b) Greenland ice sheet project (GRIP) temperatures; (c) sea surface temperatures from *N. pachyderma*, off western Ireland; (d) ice rafted debris and Heinrich Events; (e) tropical Atlantic 13C variability; (f) methane concentrations in the Greenland Ice Sheet Project 2 (GISP2); (g) changes in sediment type, Santa Barbara Basin, California; (h) north-east Pacific 13C variability; (i) southern ocean sea surface temperatures from oxygen isotopes; (j) deuterium (temperature) variability, Vostok Station, South Pole

orbital pacings, the millennial ones are square-waved in shape, and show abrupt changes in temperature. Some of these represent changes in temperature of up to 10 °C in less than a decade. Matching millennial records from high sedimentation marine core records also correspond with sea surface temperature variability (Figure 93c). These are estimated from the palaeoecological requirements of planktic fossils, notably the polar *Neogloboquadrina pachyderma* (with sinistral coiling). These millennial records are joined by other records: for example, continental pollen records in Europe, by the monsoon record from the Arabian Sea, sea surface temperatures from the Sulu and China Seas, changes in biological productivity or ocean ventilation inferred from ^{13}C millennial variability (Figure 93h), and changes in water masses from offshore California (Figure 93g).

Such millennial pacings are global in character and they show that the Earth's climate changed on short and abrupt timescales. The principal pacing, that underpins all the major changes in climate over the past 110,000 years, is that at 1,450 (sometimes stated as 1,500) years. It corresponds to pacings in the production rate of the cosmogenic isotope Berylium 10 (^{10}Be) also recorded in ice sheets. ^{10}Be is produced in the atmosphere by cosmic ray bombardment. Higher quantities are produced when the solar magnetic field is at its weakest and not shielding the Earth's atmosphere from cosmic ray bombardment. Conversely, lower quantities are produced when the solar magnetic field is at its strongest and shielding the Earth from cosmic bombardment. Therefore, lower quantities of ^{10}Be correspond to a stronger and warmer Sun and vice versa. Thus variability in solar output appears to control the 1,450 climate cycle. The 1,450 pacing occurs throughout the last glacial cycle as well as the current (Holocene) interglacial. Its amplitude is greater during the unstable geographical configuration of the ice ages because of large ice sheets, low sea levels and an isostatically depressed crust. But its interglacial mode is manifested in cycles of sea ice formation in the North Atlantic, the record of alpine glacier fluctuation, and in high sedimentation continental records throughout the world.

Cycles of ice core interstadials are known as Daansgard–Oeschger cycles, named after the Danish and Swiss scientists who discovered them. Some of them correspond to the less perfect record of interstadials on the continents, where evidence from pollen and insect faunas show climatic amelioration during ice ages. The Bølling–Allerød interstadial, between about 15 and 13 ka ago, is well known and was followed by the Younger Dryas stadsial. This dramatic reversal to cold conditions just before the Holocene may have been forced by a major meltwater flux to the North Atlantic from glacial Lake Aggasiz in North America, or it may be the result of solar forcing. Cycles of Heinrich Events (Figure 93) are known as Bond cycles, and they contain packages of Dansgaard–Oeschger interstadials. Heinrich Events may have been caused by mechanisms internal to the ice sheets, but their synchronous nature around the North Atlantic suggests they may have been forced by changes in climate.

Comparison of Greenland and Antarctic ice cores shows broad similarities, but also some important differences. One school of thought believes that deep water thermohaline ocean circulation in the Atlantic carries climate signals between the two hemispheres and is a prime control on the global climate system, with the signals originating in the North Atlantic. Others see the climate signal originating in the tropical Pacific Ocean, its global transmission taking place as millennial scale phenomena similar to El Niño events.

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ICE DAM, GLACIER DAM

Glacier dams are barriers of ice that act as a seal for water impounding glacial lakes. Water may be impounded within the GLACIER (*englacially* and *subglacially*), on the surface of the glacier (*supraglacially*), or in lakes formed at the glacier's edge (*marginally* or *proglacially*). Subaerial glacial lakes (marginal or proglacial lakes) are confined by ice on one side and by topographical barriers on the other. Subglacial lakes may form a cupola above the glacier bed at subglacial geothermal areas, accompanied by a depression in the glacier's surface, or be situated in a bedrock hollow beneath a relatively flat dome-shaped glacial surface. Supraglacial lakes are isolated in depressions on the glacier surface.

In general, ice-dammed lakes of every type can drain in episodic bursts. The glacier's surface and the water pressure potential slope towards the lake. The ice-dammed lakes receive water inflow and are gradually made to expand. Basal water pressure will increase and the lake level will be raised. Eventually, the hydraulic seal of the ice dam will be ruptured at the glacier base, the hydraulic seal opened, and seepage causes enlargement of the drainage system, initiating a flood under the surrounding ice. After discharge has begun, pressure from the ice constricts the passageway, and water flow at an early stage in the flood correlates primarily with enlargement of the ice tunnel due to heat from friction against the flowing water and to thermal energy stored in the lake. Increasing as an approximate exponent of time over a matter of hours or days, the discharge falls quickly after peaking. The recession stage of the hydrograph sets in when tunnel deformation begins to exceed enlargement by melting. Fluctuations in the thickness of the blocking ice, due to climatic variations or surges, may modify the outburst cycle or even stop bursts completely.

Occasionally, glacial outbursts are triggered by flotation of the ice dam. Rather than initial drainage from the lake being localized in one narrow conduit, the water is suddenly released as a sheet flow, surging downhill and propagating a subglacial pressure wave, which exceeds the ice overburden and lifts the glacier in order to create space for the water. In this instance, discharge increases faster than can be explained by conduits expanding due to melting.

Marginal lakes at subpolar glaciers, where the ice barrier is frozen to the bed, are typically breached as water spills over the top of the dam into supraglacial channel that melts into a bigger breach – commonly at the juxtaposition of the glacier and a rock wall.

Glacial outbursts (*jökulhlaup* in Icelandic) can have pronounced geomorphological impact, since they scour river courses and inundate floodplains. Outbursts result in enormous erosion, for they carry huge loads of sediment and imprint the landscape, with deep canyons, channelled SCABLANDs, ridges standing parallel to the direction of flow, sediment deposited on outwash plains, coarse boulders strewn along riverbanks, kettleholes where massive ice blocks have become stranded and melted, and breached terminal moraines. Some modern outbursts have produced flood waves in coastal waters (TSUNAMIs). In the North Atlantic, outburst sediment dumped onto the continental shelf and slope has been transported far away by turbid currents. Outburst floods wreak havoc along their paths, threatening people and livestock, destroying vegetated lowlands, devastating farms, disrupting infrastructure such as roads, bridges and power lines, and threatening hydroelectric plants on glacially fed rivers.

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HELGI BJORNSSON

ICE SHEET

An ice sheet is a large dome-shaped ice mass ($>50{,}000\,km^2$ in area) that exhibits a generally radial flow pattern which is predominantly unconstrained by the underlying topography. If the ice mass is less than $50{,}000\,km^2$, it is usually referred to as an ice cap. Ice sheets are comprised of a single ice dome or series of coalescent domes that represent the highest parts of the ice sheet surface. In the interior regions of the three present-day ice sheets in East Antarctica, West Antarctica and Greenland, ice thickness often exceeds several thousand metres. These central accumulation areas (or dispersal centres) are marked by ice divides which delineate neighbouring catchment areas (analogous to a drainage divide in fluvial systems).

Towards the margins of an ice sheet, the underlying topography becomes increasingly more important in channelling flow and fast moving outlet GLACIERS and ICE STREAMS may develop in deep subglacial troughs. Other ice streams may arise irrespective of the bedrock topography. Thus, flow within an ice sheet is generally divided into slow 'sheet' flow, typical of ice domes, and fast 'stream' flow, which occurs in outlet glaciers and ice streams. This may be an oversimplification because some ice streams and outlet glaciers have smaller tributaries of intermediate velocity extending well into the ice sheet interior. Ice streams and outlet glaciers are known to be a key control on overall ice sheet stability because of their capacity to rapidly drain large volumes of ice.

Once established, ice sheets respond to climate forcing and also influence the climate system over a global as well as local scale. They represent a major obstacle to atmospheric circulation and produce some of the largest regional anomalies in albedo and radiation balance (Clark *et al.* 1999). Ice sheets also store and release considerable amounts of freshwater and are one of the main regulators of continental water balance and global sea-level change. For example, during the Last Glacial Maximum (18,000–21,000 yr BP), global SEA LEVEL was around 120 m lower as continental ice sheets developed in the high to mid-latitudes of the northern hemisphere, covering large parts of North America (e.g. Laurentide Ice Sheet) and Europe (e.g. British and Scandinavian Ice Sheets). It has been discovered that rapid discharges of ICEBERGS and meltwater (see MELTWATER AND MELTWATER CHANNEL) associated with these former ice sheets exerted a profound and often abrupt impact on ocean circulation and climate (cf. Clark *et al.* 1999). There is also compelling evidence that some former ice sheets (e.g. Laurentide Ice Sheet) were characterized by relative instability during the last glacial cycle, indicating that even the largest ice masses are highly dynamic systems (Boulton and Clark 1990). A major challenge for contemporary ice sheet research lies in predicting the response of the potentially susceptible West Antarctic Ice Sheet to future changes in sea level and climate, particularly global warming.

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SEE ALSO: glacier; glacial deposition; glacial erosion; glacial theory; ice

CHRIS R. STOKES

ICE STAGNATION TOPOGRAPHY

Often also referred to as 'dead-ice topography', this is undulatory to hummocky terrain composed of a wide range of glacigenic sediments that accumulate on the surface of a melting glacier by the processes of melt-out, mass movement and glacifluvial reworking (Boulton 1972). Hummocks are often randomly distributed and of different dimensions, giving a chaotic appearance, but some linearity may be imparted by former ice structures. Although the term ice stagnation topography has been traditionally associated with deglaciated terrain, large areas of such supraglacial landform assemblages often still contain buried glacier ice, sometimes many thousands of years in age. In glaciers that carry relatively large debris loads, englacial and supraglacial sediment-landform associations develop over long periods of time and are subject to numerous cycles of reworking, re-mobilization and topographic inversion before they are finally lowered to the glacier substrate. The wide range of processes involved in the production of ice stagnation topography, including mass flowage, meltwater reworking, lacustrine sedimentation folding and faulting, result in the occurrence of a variety of sediment assemblages.

Although the term 'hummocky moraine' has been used to describe a wide range of landforms, it has most recently been restricted to moundy, irregular topography deposited by the melt-out of debris-mantled glaciers and therefore also relates to stagnating ice but is not strictly synonymous with the term ice stagnation topography. For example ice stagnation topography often encompasses large areas of glacifluvially reworked materials or supraglacial KAME and kettle (see

KETTLE AND KETTLE HOLE) topography and even complex ESKER systems and ice-walled lake plains. In some circumstances the melt-out process results in the partial preservation of englacial structures, specifically alternating debris-rich and debris-poor ice layers, and the production of transverse linear elements called 'controlled moraine'. The preservation potential of controlled moraines is poor but they are very striking elements of freshly deglaciated terrain at high latitudes where glacier snouts are melting down very slowly and supraglacial sediment reworking is locally restricted.

The processes of differential ablation, multicyclic debris reworking and topographic inversion and complex meltwater drainage development on a debris-rich glacier will result in the production of a thick supraglacial debris mantle that can decouple glacier response to climate forcing. Ice wastage or stagnation can therefore lag behind climatic inputs by decades or even thousands of years. As meltwater systems gradually open up large subglacial and englacial drainage conduits so the developing ice stagnation topography may become perforated by numerous moulins, ponds and lakes. Referred to as glacier karst (Clayton 1964), these water-filled depressions often coalesce to produce large supraglacial lakes.

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DAVID J.A. EVANS

ICE STREAM

An ice stream is a zone of an ICE SHEET that flows much faster than the surrounding ice. Typical ice sheet velocities are of the order of tens of $\mathrm{m\,a^{-1}}$ but ice stream velocities range from hundreds to several thousands of $\mathrm{m\,a^{-1}}$. Ice streams display a broad range of characteristics and behaviour but they are generally large features with widths of tens of km and lengths of hundreds of km. Their large size and profligate ice flux dominate ice sheet discharge and it is for this reason that they are viewed as a critical control on ice sheet mass balance and stability.

A defining characteristic of ice streams is that they are bordered by slower moving ice. This creates heavily crevassed lateral shear margins that aid their identification. If the fast-flowing ice is bordered at the surface by rock walls, it is usually referred to as an outlet glacier. It should be noted, however, that many ice streams show characteristics of both along their length. To add to their complexity, some ice streams appear to be fed by numerous smaller tributaries that penetrate up to 1,000 km into the interior of the ice sheet (Bamber *et al*. 2000).

Ice streams have been investigated in the East Antarctic, West Antarctic and Greenland Ice Sheets and have also been hypothesized in many palaeo-ice sheets. They occur in a variety of settings and their fast flow may be achieved through a variety of flow mechanisms. At one end of the spectrum, some ice streams occupy deep bedrock troughs and they are often characterized by steep surface slopes and high driving stresses. A large component of their fast flow arises from rapid basal sliding but, under exceptionally high driving stresses, thermally enhanced deformation of a thick basal ice layer may be the dominant flow mechanism (cf. Iken *et al.* 1993). In contrast, some ice streams have low surface slopes, low driving stresses and do not appear to be constrained by the underlying topography. Their behaviour is more enigmatic but their fast flow appears to be related to a metres thick layer of soft, saturated sediment that deforms beneath the ice stream and/or provides an effective surface for basal sliding (Engelhardt and Kamb 1998). Other ice streams show characteristics of both of these end-member behaviours and many more ice streams remain to be studied.

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SEE ALSO: glacier; ice; ice sheet

CHRIS R. STOKES

ICE WEDGE AND RELATED STRUCTURES

Ice wedges are among the most common forms of ground ice within continuous PERMAFROST. They are commonly 1.0–1.5 m wide at the top and up to 4 m deep (Plate 64), although extreme examples up to 4 m wide and 10 m deep are reported from Siberia. In plan the linear ice wedge structures link to form polygonal patterns with a mesh size of several tens of metres, termed ice-wedge polygons or tundra polygons. The surface expression is a linear furrow due to differential settlement of the active layer immediately above the ice wedge.

Ice wedges result from thermal contraction cracking of the permafrost. Low winter air temperatures, and lack of insulating snow, lead to rapid cooling of the ground and the development of tensile stresses. Lachenbruch (1962) showed that cracking is most likely when the ground temperature is below approximately −20 °C, though slightly lower temperatures may be necessary in sands and gravels than is the case in silts and clays. Entry of snow and hoarfrost prevents crack closure as the permafrost warms during the following summer, leaving thin (often less than 1 mm; Mackay 1992) veins of ice penetrating into the permafrost. These provide lines of weakness for further cracking in subsequent years, leading to progressive widening into ice-wedges over time. Wedge growth associated with individual cracking events is generally less than 1 mm, and cracking may occur only periodically, so that a 1-m wide ice wedge may take thousands of years to form (Harry and Gozdzik 1988).

Epigenetic ice wedges develop below a stable ground surface and are younger than the adjacent frozen host sediments, while syngenetic ice wedges form contemporaneously with the slow

Plate 64 Upper portion of an ice wedge, Ellesmere Island, Canada. Note the active layer above the wedge is approximately 0.75 m deep at this location

accumulation of host sediments in subaerial permafrost environments. Where aeolian sediment transport is active, sand or silt – rather than snow and hoarfrost – may enter open thermal contraction cracks, resulting in the formation of sand-filled wedge structures termed sand wedges (Murton *et al*. 2000).

A mean annual air temperature value of between −6 °C and −8 °C has been widely used as the warm limit for the development of thermal contraction cracking and the formation of ice wedges (Péwé 1966), although variations in seasonal extreme temperatures may lead to a range of limiting mean annual temperatures. Thawing of permafrost and contained ice wedges is associated with slumping of sediments to fill the resulting voids, thus preserving the wedge forms within the host sediments. Such 'pseudomorphs' or 'casts' are important stratigraphic markers in sedimentary sequences, providing evidence for the former existence of permafrost (Svensson 1988). Host sediments may be upturned against the sides of casts, marking former compression during summer ground warming, or there may be downward slumping of adjacent sediments into the cast.

Ice-wedge casts and sand wedges may be preserved within sedimentary units (intraformational), between sedimentary units (interformational) or they may penetrate downwards from the present ground surface (supraformational). Intraformational casts suggest episodic sediment accumulation in a permafrost environment and are common in fluvial gravel trains (Seddon and Holyoak 1985). Interformational wedge casts indicate a major change in sedimentary environment separated by a phase of permafrost. Supraformational casts may be visible in aerial photographs as polygonal 'crop marks' (Svensson 1988).

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CHARLES HARRIS

ICEBERG

Icebergs are floating pieces of glacier ice that are found in marine and lake waters. They are produced when ice breaks off from the margins of glaciers and ice sheets that terminate as ice cliffs in the sea and lakes. Icebergs vary in length from metres to kilometres – those less than 5–10 m in length are referred to as bergy bits. Larger icebergs are usually flat-topped or tabular in shape, whereas smaller bergs, and those in the later stages of melting and breakup, are of irregular form. The ice in icebergs comes originally from snowfall on the glacier surface and is transformed into glacier ice as a result of pressure during burial. This ice may be hundreds or even thousands of years old by the time it calves into the sea. Icebergs are less dense than water because of the presence of air bubbles that are trapped within them during formation. This density difference makes them buoyant, and about 80–90 per cent of their bulk is hidden below the water surface. Iceberg keels can reach hundreds of metres deep. Once icebergs have broken off from their parent glacier, they drift under the influence of ocean currents and, to a lesser extent, wind.

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JULIAN A. DOWDESWELL

ICING

Sheets of ice formed on the ground or on lake or river ice by the freezing of successive flows of water discharging from the ground or rising up through fractures in an antecedent ice cover. Icings are not restricted to PERMAFROST areas, but are common in the discontinuous permafrost zone, especially in carbonate terrain. Icings may be a serious hazard for road traffic, occurring persistently at the base of road cuts and at creek crossings. Most icings melt during the first few weeks of summer, but some extensive features may persist for years.

River icings are the most extensive features, and may extend over tens of km^2. The spatial extent of the icing depends on the discharge rate of water to the ice surface, water temperature, air temperature and channel slope (Hu and Pollard 1997). Icings are characteristically layered, with each bed the product of a discharge event. The basal portion of these layers may be discoloured by the solutes expelled and concentrated during freezing. Icings may extend outside the channel used by the river in summer, and may leave a dusting of precipitate on the local vegetation as they melt.

Icing blisters form if hydraulic pressure lifts an overlying ice layer. In continuous permafrost, such blisters have been observed at the edges of pingos and in residual ponds of drained lake beds, where water is expelled during permafrost aggradation (Mackay 1997). Such features are similar to frost blisters, but lack overlying ground.

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C.R. BURN

ILLUVIATION

Illuviation is a process by which material removed from one horizon is deposited in another horizon of a soil (Foth 1984; Ritter 1986; Soil Science Society of America 1987). Usually the direction of transport is from an upper to a lower horizon. The lower horizon, an illuvial horizon, is considered a zone of accumulation and concentration for material. The material can be precipitated from solution or deposited from suspension. Depending on the conditions of soil formation, different illuviated constituents are present. For example, an acid soil formed under forest vegetation may exhibit a dark-coloured illuvial horizon containing quantities of sesquioxides and mineral-organic complexes as well as clay minerals and clay-size material. These materials have been eluviated or transported from an overlying horizon that may be significantly depleted in these constituents. A soil in a prairie region may exhibit only clay increases in the illuvial horizon or zone of accumulation. For contrast, see ELUVIUM AND ELUVIATION.

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CAROLYN G. OLSON

IMBRICATION

Imbrication is the orientation of an assembly of rocks or pebbles in one predominant direction (usually upstream), as a result of flow movement. Imbrication is common on gravel beaches, on bars and outwash fans in braided rivers, and glacial tills. Additionally, imbrication is often a key for interpreting facies. The term imbrication also refers to the near parallel overlapping and orientation of a series of lesser thrust faults, directed towards the source of stress.

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STEVE WARD

INHERITANCE

Landforms and landscapes have different lifetimes, which generally increase proportionally to the size of a landform. Therefore, many small landforms are formed and destroyed relatively quickly, over timescales of 10^3–10^4 yr, especially if the substratum is soft. By contrast, landforms built of hard, resistant rocks such as certain INSELBERGS could be very durable and may survive for much longer, over timescales of 10^5–10^7 yr. Likewise, regional landscapes such as extensive PLANATION SURFACES may have considerable lifetimes, measurable in millions of years. With such protracted histories, the bearing of present-day environmental conditions on their appearance and development is limited, and inheritance of the past becomes crucial.

Inherited landscapes are those that were formed in the past, under influences of a different external environment and/or within a tectonic regime different from the recent one, and have retained their characteristic features up to now, because later processes have proved to be inadequate to obliterate earlier sets of type landforms. Hence, there is no universal, worldwide applicable boundary between 'inherited' and 'contemporary'. In high latitudes of northern Europe and North America everything that predates Pleistocene continental glaciation is usually considered 'inherited', so that inherited landscapes are older than 1–1.5 Ma. By contrast, in Australia and South America some landscapes are likely to be inherited from the period preceding the breakup of Gondwana in the Mesozoic (Ollier 1991; Twidale 1994). In Central Europe, in turn, major change in the course of geomorphic evolution is often associated with growing tectonic instability from the Palaeogene/Neogene boundary onwards. Therefore, 'inherited' landforms would be those which originated within a general regime of tectonic quiescence and are older than ~20 Ma. However, in the specific context of Holocene morphogenesis, Pleistocene periglacial landscapes will also be regarded as inherited. It has to be emphasized that, strictly speaking, each landscape is recent in the sense that current processes do some action on it; yet the rate of change could be so negligible that the origin of gross features of such a landscape can be traced well back beyond the present day. In other situations, inherited forms may occur very close to relief features of recent origin, forming a palimpsest of landforms of contrasting origins and ages (Starkel 1987; Brunsden 1993a). In the case of the Fennoscandian Shield, where exhumed elements play a significant part, the range of ages of inherited landscapes spans from the Precambrian through the Jurassic, Cretaceous, late Tertiary up to the Pleistocene (Lidmar-Bergström 1995).

The phenomenon of landform and landscape inheritance has been recognized in geomorphology for a long time, though the very term 'inheritance' has not always been used. A German geomorphologist, S. Passarge (1919), made a distinction between *Vorzeitformen* (i.e. inherited) and *Jetztzeitformen* (i.e. moulded at present). Realization that the history of environmental change is very complex itself has led to the concept of 'generations of relief' (Büdel 1977). Accordingly, landscapes observed nowadays comprise units inherited from various epochs of the past, each recognizable through its characteristic set of landforms.

Inherited landscapes should not be equated with persistent landscapes as elaborated by Brunsden (1993b). The latter are landscapes which maintain their principal characteristics over time, and they do this for two main reasons. The first is that they have undergone very little change since their period of origin. The second is that persistent characteristics are maintained either because they are renewed or because they are uniformly changed. Thus while they maintain their original morphology, the currently observed landforms are not necessarily old. Therefore 'steady-state' landscapes are persistent. Zones of high relief may be maintained, for example, by erosion being compensated by isostatic adjustment. Hence, while every inherited landscape is persistent, persistent landscapes are not themselves necessarily inherited from the distant past.

Evidence for inheritance of land surfaces comes from three main sources. The most convincing one is offered by the occurrence of sediments of known age within denudational landscapes; these parts of landscapes are then at least contemporaneous with the sediments. Remnants of weathering mantles, including DURICRUSTS, are essentially of the same significance as the sediments, although the age of weathering residuals is much more difficult to establish. The third line of evidence is provided by the landforms themselves. In many cases there exists a sharp boundary between landscape units which helps to differentiate between older (inherited) and younger landforms. Examples include glacial U-shaped valleys breaching former watershed ridges, antecedent valleys cut into older surfaces and sea cliffs truncating subaerial landforms of long geomorphic history.

The survival of ancient landscapes means little geomorphological change, and these low rates of long-term denudation may be causally linked to one or more factors (Twidale 1999; Migoń and Goudie 2001). Subsiding areas retain their inherited geomorphological features for longer than areas subjected to surface uplift because erosional dissection and water divide lowering within the former is less effective. Moreover, as dissection proceeds and more rock mass is eroded, isostatic recovery (see ISOSTASY) becomes increasingly important, promoting further erosion and destruction of any remnants of old surfaces. However, rapid surface uplift with limited dissection seems capable of elevating a palaeosurface without transforming it into an all-slope topography. An inherited landscape will then be present at a high elevation.

Further reasons for inheritance may include climatic conditions unfavourable to rapid progress of denudation, high bedrock resistance, distance from base level or other lines of active erosion, and the protective role of durable duricrust blankets. In formerly glaciated countries the protective role of cold-based ice or the location within ice divides may play an additional part. A separate category is temporary burial and later re-exposure of an ancient landscape.

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SEE ALSO: climato-genetic geomorphology

PIOTR MIGOŃ

INITIATION OF MOTION

The initiation of sediment transport (also known as incipient motion or the critical condition) describes when one or more particles are moved from a stationary bed. In river or air flows, the force balance for a particle resting on the surface can be resolved into three components: a drag force acting parallel to the bed, known as the shear stress (τ), an upward lift force produced by turbulence and the downward frictional force caused by the particle's weight. In 1936, Shields showed that the onset of particle motion is defined as $\tau_c^* = \tau_c/(\rho_s - \rho)gD$ where τ_c^* is the dimensionless shear stress at the critical condition, τ_c is the tangential fluid shear stress at the boundary as grain movement begins and ρ_s, ρ, g and D are the sediment density, fluid density, acceleration due to gravity and grain diameter at the bed surface, respectively. Using a series of laboratory flume experiments with sediment of uniform grain size, Shields showed that the value of τ_c^* lies within a narrow range for hydraulically rough and turbulent flow (mean $\tau_c^* \approx 0.046$ when corrected for sidewall effects and form drag). Consequently, the Shields equation stated above simplifies to $\tau_c \alpha D$ which implies that a stronger fluid force will move coarser particles. Some authors refer to this finding as the principle of 'selective entrainment'.

Although intuitively attractive, it has now been shown that there is a frequency distribution of τ_c^* values for any particular grain size. This is because the three-dimensional forces of turbulent shear stress that impact on the bed vary through space and time and the potential for movement of a particular grain is strongly dependent on the physical arrangement of the bed surface (e.g. grain pivoting angle, degree of packing and sorting, mixture of grain shapes). The variation in τ_c^* is most marked over coarse-grained and hydraulically rough boundaries where different particle sizes protrude or are 'hidden' behind and below other particles on the bed surface. Researchers have shown that τ_c^* decreases with an increase in relative particle size (i.e. the ratio between the size of the surface particle and the ambient grains). So long as there is an overrepresentation of coarse sediment on the bed surface, sediment entrainment may achieve what is termed 'equal mobility' whereby all particles in the surface may be moved regardless of their absolute size or weight and the BEDLOAD grain size distribution matches that of the bed subsurface.

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SEE ALSO: downstream fining; fluvial armour; mobile bed

PHILIP J. ASHWORTH

INLAND DELTA

The term was coined by cartographers in the late nineteenth century to describe a morphologically unusual reach of the Niger River in the interior of west Africa (Mali), which they termed the Niger Inland Delta. Here, in the vicinity of the confluence (see CONFLUENCE, CHANNEL AND RIVER JUNCTION) of the Niger and Bani Rivers, the two rivers divide to form a network of anastomosed (see ANABRANCHING AND ANASTOMOSING RIVER) channels spread over a FLOODPLAIN approximately 100 km wide and 200 km long. The channels terminate in Lakes Debo and Korientze, which are ponded against aeolian dunes on the southern edge of the Sahel. A single channel forms beyond the dunefield. The adjective 'inland' was added to distinguish this feature from the delta (see RIVER DELTA) at the mouth of the Niger. A form of the term was also applied to the terminus of the Okavango River in the endoreic Kalahari Basin of central southern Africa (Botswana). After meandering through a 10 to 15 km wide, 100 km long corridor known as the Panhandle, the Okavango River abruptly divides into a number of radial, distributary channels extending over a floodplain some 50,000 km^2 in area, known as the Okavango Delta. The adjective 'inland' was omitted in this case because the Okavango River, unlike the Niger, does not reach the ocean, so there is no risk of confusion. A geomorphologically similar feature is also developed on the White Nile River (Bahr el Jebel) in Sudan, where the river divides into several distributary channels forming a triangle 520 km long and 150 km along its base. The term 'delta' has not been applied in this case, however, and the feature is known by its Arabic name, the Sudd.

The Niger Inland Delta, Okavango Delta and the Sudd have several features in common. They occur in actively subsiding half-grabens, possibly related to incipient rifting. The multiple channels of these inland deltas form in response to the loss of confinement of flow as the rivers enter the grabens. They have relatively low gradients: 3 cm km^{-1} for the Niger, 10 cm km^{-1} for the Sudd and 28 cm km^{-1} for the Okavango. River discharge is seasonal and the flood wave moves slowly down the floodplain, taking three months to cross the Niger Inland Delta, four months to cross the Sudd and five months to cross the Okavango Delta. There is considerable evapotranspirational loss of water – 98 per cent in the case of the Okavango.

Whilst the channel geometry of these inland deltas is superficially similar to those on deltas at river mouths, the sedimentation processes are fundamentally different. In river deltas, sedimentation occurs primarily on the delta front as the distributary channels lose their ability to transport sediment on entering standing-water bodies. River deltas therefore build outwards into water bodies. In contrast, on inland deltas sediment deposition occurs primarily as a result of flood water spilling from channels and spreading laterally as sheet flow. Inland deltas therefore aggrade vertically. Local AGGRADATION around channels results in instability, causing AVULSION which gives rise to new channels. In this way, sediment is spread uniformly across the delta surface. This sedimentary style is more akin to that occurring on ALLUVIAL FANs, and these inland deltas constitute a variety of low gradient alluvial fan.

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SEE ALSO: alluvial fan; megafan; river delta

T.S. McCARTHY

INSELBERG

Inselberg is a descriptive term, derived from German (*inselberg* literally means 'island hill') and adopted in English, used to describe a hill which stands in isolation and rises sharply above the level of the surrounding plain. It was first coined by W. Bornhardt, a German naturalist travelling in East Africa at the turn of the nineteenth century, to emphasize visual similarities between steep-sided hills dotting an otherwise flat savanna and islands rising from the sea. Inselbergs often tend to occur in groups, to form inselberg landscapes.

It is generally accepted that the term applies to a hill produced by lowering of the surface around it. Therefore volcanoes and small tectonic horsts are normally not called inselbergs (Kesel 1973). It was also noted that the term is rather infrequently used to describe isolated hills built of flat-lying sedimentary rocks such as MESA or BUTTE, hence the impression arises that inselberg landscapes are restricted to basement rock areas. There have been attempts to make the definition more strict by using certain quantitative constraints for inselbergs. These included minimal height of 15 m, length to width ratio not exceeding 4 : 1, or minimal distance to the nearest neighbour of 0.8 km (Faniran 1974). None of these proposals has been universally accepted and there is quite a degree of freedom in deciding which hill is to be called an inselberg, which makes comparative analyses of inselberg landscapes difficult.

Visual differences provide the basis for classification of inselbergs, most readily applicable to granite hills. Three main types are distinguished (Thomas 1978; Twidale 1981). Domed inselbergs have slopes convex-upward and are built of massive, poorly jointed rock, with little REGOLITH at the surface. They are called BORNHARDTs, but it is important to note that the two terms are not equivalents. Hills having all the characteristics of bornhardts listed above can also occur within hilly and mountainous terrains, and hence are not by any means inselbergs. Examples include 'sugarloaf' hills in Rio de Janeiro, or granite domes in the Yosemite National Park in the USA. Castellated inselbergs are built of well-jointed but bedrock-rooted rock and their detailed morphology consists of towers, pillars and walls, separated by joint-aligned avenues. Boulder inselbergs are composed of loose boulders, chaotically lying one upon another. They typically form through advanced degradation of a domed or castellated inselberg. In addition, some inselbergs, especially in metamorphic rocks, may be conical, and there are also hills built entirely of SAPROLITE.

Inselbergs tend to be built of hard, resistant igneous and metamorphic rocks such as granite or gneiss. They are particularly common in coarse-grained and poorly jointed granites, rich in potassium feldspar (Pye *et al.* 1986). They are also known from gabbro, syenite, rlyolite and migmatite. However, there are relatively few instances where the base of the hillslope coincides with a lithological boundary, nor are differences in mineralogical composition between the hill and the plain significant (Brook 1978). The majority of inselbergs show joint- rather than lithological control. Bedrock to build an inselberg is typically massive, with few open fractures and the predominance of tight SHEETING joints. The latter are often arranged concentrically to form a structural dome, the outline of which is followed by the topography. It is generally assumed, although usually not proved, that joint density around a hill is higher than within the hill, and the existence of an inselberg reflects primary variations in the degree of rock mass fracturing. Another manifestation of joint control is seen in plan. Outlines of inselbergs frequently follow master joints (Twidale 1982; Selby 1982).

Notwithstanding the impression that inselbergs are only supported by igneous rocks, spectacular inselberg assemblages can also be built of sedimentary rocks, especially of massive sandstone and conglomerate. Ayers Rock (Uluru) in central Australia, one of the most famous inselbergs

Plate 65 The granite inselberg of Sptizkoppe in the Namib Desert is built of extremely massive – and therefore most resistant – granite, which accounts for its impressive height of more than 600 m

worldwide, is made of steeply dipping arkosic sandstone, whereas the adjacent Olgas are composed of massive conglomerate. Further examples of isolated sandstone towers have been reported from certain arid areas, such as the Sahara Desert in Niger and Mali, or the deserts of south-west Jordan. Interestingly, sandstone inselbergs in these areas are not associated with any actively retreating escarpments but rather have originated due to advanced dissection of a former plateau.

The origin of inselbergs had been a matter of hot debate, but it is now accepted that they may form in various ways, and contrasting evolutionary pathways may produce landforms looking superficially very similar. There are at least three theories present in the literature. Perhaps the most universally accepted one holds that inselbergs are products of two-stage development involving differential deep weathering in the first phase and stripping of the weathering mantle in the second one, which leaves an unweathered rock compartment exposed at the surface (Thomas 1965). Reasons for survival of such compartments include significantly wider spacing of joints, the presence of primary poorly fractured mass, enrichment in quartz and/or potassium feldspar, or petrological differences. Validity of the two-stage hypothesis has been confirmed in deep excavations and quarries in equatorial Africa, where massive hill-like features up to 50 m high have been found within a thick mantle of decomposed bedrock. In reality, the average thickness of weathering mantles seems insufficient to account for the height of many inselbergs, which may exceed 200–300 m. Therefore their exposure is more likely to have been accomplished in many stages of weathering and stripping (Twidale and Bourne 1975). Minor landforms on hillslopes such as flared slopes and platforms are considered as the evidence of multiphase exposure.

Another hypothesis invokes scarp retreat across unweathered, but possibly differentially jointed, bedrock and links the origin of inselbergs with the cyclic development of relief and retreat of big escarpments separating denudational landscapes of various ages (King 1949). Clustering of inselbergs in front of major escarpments might validate this theory, but other people argue for an important role of deep weathering in the development of scarps and inselbergs too (Bremer 1993).

The massive nature of many prominent inselbergs found in the semi-arid and arid zone, and the occurrence of more jointed compartments around their flanks, suggest that exposure and growth in height may not necessarily be associated with deep weathering. Long-term lowering of differentially jointed bedrock probably accounts for the origin of the spectacular inselbergs of the Namib Desert (Selby 1982) and may be applicable elsewhere. In the same way, some minor granite intrusions may have been exposed as inselbergs as the surrounding less resistant schist has been completely eroded away.

Inselbergs, once exposed or isolated, undergo further development and are the scene of competing processes of continuous growth and destruction. Many authors emphasize that inselbergs are very durable, long-lived landforms because their surfaces shed rainwater and remain dry, being therefore immune to chemical weathering. At the same time, because they are built of poorly jointed bedrock, they are resistant against physical weathering too. Fast runoff from the slopes of an inselberg provides additional moisture to its footslopes, so the rate and intensity of weathering are enhanced. Episodic removal of regolith cover from around the inselberg may then result in the increase of its height, providing the lowering of its summit proceeds at a slower rate. In Australia, it is argued that inselbergs have been rising for millions of years and their top surfaces may date back to the Mesozoic (Twidale and Bourne 1975; Twidale 1978).

How inselbergs are reduced in height and extent depends on their jointing patterns. Massive domed inselbergs are subject to mega-exfoliation due to pressure release and opening of sheeting joints. Individual slabs are separated from the underlying rock mass and fall or slide off the slopes, forming big debris cones mantling lower slopes. Ongoing exfoliation will gradually reduce the surface area of an inselberg. Rock fall is also typical for slope development of sandstone inselbergs in deserts. In orthogonal patterns, vertical joints open too, topples occur, and the summit part assumes ruiniform relief. Minor weathering features play an important part in the development of inselbergs (Watson and Pye 1985). Selective weathering along fractures and growth of TAFONI reduces rock mass strength and facilitates mass movement, whereas horizontal surfaces are destroyed by enlargement of WEATHERING PITS. Caves and massive overhangs are frequently reported from granite inselbergs and develop either through mechanical widening of fractures, preferential weathering along sheeting joints, or chaotic accumulation of big boulders on lower slopes.

The discussion about the origin of inselbergs has a direct bearing on the issue of their significance in geomorphology, especially in CLIMATIC GEOMORPHOLOGY and CLIMATO-GENETIC GEOMORPHOLOGY. Two positions emerge from the literature. One holds that inselbergs occur all around the world and cannot be considered as indicators of environments, present or past (Kesel 1973; King 1975). King argued that the process of scarp retreat is climate-independent, hence inselbergs are not dependent on climatic conditions either. By contrast, a German school of geomorphology has maintained that inselbergs are specific products of landscape development in the seasonally humid tropics and develop through deep weathering and stripping, which are the processes acting at their highest efficacy in this zone. Consequently, inselbergs present in middle or high latitudes, or in arid areas, would be relict landforms, inherited from the geological past.

Although it is probably true that humid tropical environments with ubiquitous deep weathering favour the development of inselbergs, a claim that inselbergs are by definition 'tropical' landforms is most likely wrong. Inselbergs are present in many desert areas of the world, including the long-lived ones such as the Namib, and in many evidence for inheritance from previously humid conditions is clearly lacking. Nevertheless, the origin of those present in central and northern Europe, as well as in North America, is usually traced back to the Early Tertiary, when climate was warmer and wetter, and deep weathering was widespread. It is worth noting that even if these inselbergs are indeed ancient Tertiary features, it does not mean that tropical climate was essential for their origin.

Inselberg landscapes have been reported from all around the world, including parts of Antarctica, but they are probably most widespread in Africa, within extensive tracts of crystalline rock terrain. Classic examples are known from Nigerian savannas, East African plains in Kenya and Tanzania, Zimbabwe, South Africa, Namib Desert and Angola. Namibian inselbergs, such as the almost 700-m high Spitzkoppe, belong to the highest in the world. There are also examples from the Sahara and its southern margin, from Sudan, Niger and Libya. Inselbergs are also common in Australia, especially in the central and western part of the continent. Further examples are known from the Indian Peninsula and basement areas of South America. Numerous, purportedly inherited, inselberg landscapes have been described in Europe, including Germany, the Czech Republic, Poland, Hungary and Scandinavia.

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PIOTR MIGOŃ

INSOLATION WEATHERING

Insolation weathering (thermoclasty or thermal stress fatigue) is the rupturing of rocks and minerals primarily as a result of large daily temperature changes in dry environments which lead to temperature gradients within the rock mass. Fires can operate in a similar way, though the temperature extremes are greater (see FIRE). Areas that are heated expand relative to the cooler portions of the rock and stresses are thereby set up.

In igneous rocks, which contain many different types of mineral (polymineralcy) with different coefficients and directions of expansion, such stresses are enhanced. Moreover, the varying colours of minerals exposed at the surface (polychromacy) will cause differential heating and cooling.

Daily temperature cycles under desert conditions may exceed 50 °C, and during the heat of the day rock surfaces may occasionally exceed a temperature of 80 °C. However, rapid cooling takes place at night, creating, it has been thought, high tensile stresses in the rock. Desert travellers, like David Livingstone, have claimed to hear rocks splitting with sounds like pistol shots in the cool evening air – certainly, split rocks are evident on many desert surfaces. Insolation was a mechanism that found considerable favour with pioneer desert geomorphologists (e.g. Hume 1925; Walther 1997).

At first sight the process of insolation weathering seems a compelling and attractive mechanism of rock disintegration. However, doubt has been cast upon its effectiveness upon a variety of grounds (Schattner 1961). The most persuasive basis for doubting its power was provided by early experimental work in the laboratory by geomorphologists like Blackwelder, Griggs and Tarr. They all found that simulated insolation produced no discernable disintegration of dry rock, but that when water was used in the cooling phase of a weathering cycle disintegration was evident. This highlighted the importance of the presence of water. Likewise, studies of ancient buildings and monuments in dry parts of North Africa and Arabia showed very little sign of decay except in areas, for example close to the Nile, where moisture was present (Barton 1916). Indeed, there are many situations where there is moisture in deserts (e.g. where there is fog, dew and groundwater seepage). When water combines chemically with the more susceptible minerals in a rock they may swell, producing a sufficient increase in volume to cause the outer layers of rock to be lifted off as concentric shells, a process called EXFOLIATION. Thus, some of the weathering that used to be attributed to insolation may now be attributed to chemical changes produced by moisture, including HYDRATION.

However, the importance of insolation cannot be dismissed entirely. The early experimental work had grave limitations: the blocks used were very small and unconfined, the temperature cycles were unrealistic and only a limited range of rock types was used (Rice 1976). Moreover, engineering and ceramic studies have shown that a threshold value for thermal shock approximates to a rate of temperature change of $2\,°\mathrm{C\,min^{-1}}$. Datalogger studies show that such rates can occur, not least in polar regions (Hall and André 2001). In addition, consideration of fracture patterns observed on rock in cold, dry environments appears to show very similar forms to those produced in thermal shock experiments in the laboratory (Hall 1999). Finally, because of the temperature response of calcite crystals, marble seems to be especially prone to thermal degradation (Royer-Carfagni 1999)

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A.S. GOUDIE

INTEGRATED COASTAL MANAGEMENT

Coastal geomorphology and associated coastal resources are subject to increasing pressure from human impact. In a global context, this is significant for two reasons. First, the majority of the world's population currently lives near the coast and the proportion of coastal dwellers is projected to increase in the future. The world's coastal population has been variously estimated as over half living within 60 km of the coast (UNCED 1992: 17.3); 1.2 billion within 100 km of the coast (Nicholls and Small 2002, based on 1990 data); or 3.2 billion within 200 km and two-thirds within 400 km (Hinrichsen 1998). Second, humans have a high dependence on coastal resources. According to key researchers in resource economics (Costanza *et al.* 1997) the coastal biome currently contributes over 40 per cent of the total global flow value of ecosystem services.

These global population pressures and human dependence on the coast require appropriate management strategies which recognize the dynamic nature of the coast at various geomorphic timescales and across different spatial dimensions. While longer timescales may be less problematic from a management perspective, the identification of rapid geomorphic change may be difficult to separate from human-induced change. For example, coastal wetlands may be subject to human impacts of reclamation, development or subsidence through groundwater withdrawal acting simultaneously with natural processes of wetland loss such as local relative sea-level change or sediment compaction. It is also important to recognize that local coastal impacts may have broader spatial linkages to marine or land-based processes. For example, coastal pollution, erosion or accretion near a river mouth may be attributed to poor catchment management practices rather than localized coastal processes.

For these reasons, there has been a recognition that the coast, along with other types of environment, needs to be managed in a holistic rather than a sectoral manner. This has given rise to an increasing acceptance of integrated resource management in general, and more specifically to integrated coastal management (ICM). Although the concept of coastal management has been around for more than thirty years, particularly with the introduction of coastal legislation to the United States in 1972, it was the United Nations Conference on Environment and Development (UNCED) (also known as the 'Earth Summit') held in Rio de Janeiro in 1992 that created an international push for the adoption of 'integrated' coastal management with an agreement that coastal states should 'commit themselves to integrated management and sustainable development of coastal areas and the marine environment under their national jurisdiction' (UNCED 1992: Agenda 21: 17.5).

Similar objectives are contained within the United Nations Framework Convention on Climate Change (1992) which outlines the need to develop integrated plans for coastal management. In the following year (1993), the Council of the Organisation for Economic Cooperation and Development and the first World Coastal Conference adopted and produced guidelines for the integrated management of coastal resources. These *inter alia* required that nations developing their ICM should take into account the traditional, cultural and historical perspectives and conflicting interests and uses (IPCC 1994). In order to define ICM it is useful to draw on earlier definitions from the IPCC (1994: 40), Cicin-Sain and Knecht (1998: 39) and also from the Joint Group of Experts on the Scientific Aspects of Marine Environmental Protection (GESAMP 1996: 2).

Integrated Coastal Management is a continuous and dynamic process incorporating feedback loops which aims to manage human use of coastal resources in a sustainable manner by adopting a holistic and integrative approach between terrestrial and marine environments; levels and sectors of government; government and community; science and management; and sectors of the economy.

It is important to realize that although there are a number of definitions of ICM, there is still some confusion over the use of other related coastal management terms and definitions. Cicin-Sain and Knecht (1998) and Burbridge (1999) note that there has been a major change in emphasis away from coastal 'zone' or 'area' management towards 'integrated' coastal management. Cicin-Sain and Knecht argue that the terms integrated coastal zone management (ICZM), integrated marine and coastal area management and integrated coastal management (ICM) all refer to the same concept and they adopt ICM for reasons of consistency and simplicity. Others still use the term ICZM (see Salomons *et al.* 1999) although some authors in the same volume (Burbridge 1999; Harvey 1999) prefer the term ICM. In a major review of the coastal management literature Sorensen (1997) distinguishes between ICM as a concept or field of study and ICZM as a programme which has the task of defining the boundaries of the coastal 'zone'. On balance it appears that although the use of the term 'zone' was originally intended to be flexible, it can also be interpreted as prescriptive if the identification of boundary conditions mitigate against the need to integrate across them. For this reason, the use of ICM is becoming more acceptable and common in the literature.

There is now a global trend toward a more integrated approach for coastal management which incorporates linkages between activities in coastal lands and waters. A decade ago, the Earth Summit (UNCED 1992) recognized the need for a new approach to marine and coastal area management developed at the national, subregional, regional and global levels. It also commented that any new approach to coastal management should be integrated in content, and precautionary and anticipatory in its scope. Subsequently, there have been various international attempts to develop guidelines for ICM, stressing the importance of strengthening and harmonizing cross-sectoral management. While there are different approaches for achieving ICM, most agree that horizontal and vertical integration and co-ordination must be part of any attempt to achieve ICM.

Cicin-Sain and Knecht (1998) suggest that the rationale for an integrated approach is first to examine the effects of ocean and coastal use, as well as activities further inland, on ocean and coastal environments; and second to examine the effects that ocean and coastal users can have on one another. The World Coast Conference puts it simply as the need for co-operation between all responsible actors involved in coastal management (IPCC 1994: 25). The key elements of integration in coastal management can be defined as follows:

- Intergovernmental integration (vertical integration) between different levels of government such as national, provincial or state and local governments;
- Intersectoral integration (horizontal integration) between different government sectors: such as industry, conservation, recreation, tourism, beach protection and integration of policies between different sectors of the economy;
- Community integration with government producing effective community participation and involvement in coastal management;
- Spatial integration between management of the land, ocean and coast;
- Integration between science and management particularly between different disciplines; scientists and managers; including economic, technical and legal approaches to coastal management;
- International integration between nations on trans-boundary coastal management issues.

There are a number of requirements for the success of ICM. There is a need for a long-term strategy with clearly defined national objectives and guiding principles for coastal managers. It is also necessary to have a defined authority with responsibility for the strategy along with monitoring of key performance indicators and most importantly there is a need for the political will for implementation of ICM. Various conceptual models (e.g. Cicin-Sain and Knecht 1998: 58) of the different stages in the ICM process all emphasize the cyclical nature of the process and the need for continual re-evaluation. The implication of this is a need to proceed through all stages at least once before the success of ICM can be properly evaluated. This is likely to take a number of years.

Since the Earth Summit and the World Coast Conference, ICM has been adopted by many nations with coastal management programmes and associated legislation. Cicin-Sain and Knecht (1998) conclude from their cross-national ICM survey in 1996 that there were approximately 150 national ICM efforts globally, including the following countries: United States, United Kingdom, Belize, Brazil, Costa Rica, Ecuador, Sri Lanka, Turkey, Australia, Canada, Italy, China, Mexico, Nigeria, Venezuela and Pohnpei State (Federated States of Micronesia). Subsequently, Sorensen has created a global data base (www.uhi.umb.edu) which, in 2000, contained a total of 385 'ICM efforts' comprising 250 in 87 countries, plus 100 in the United States and 35 internationally. However, care is needed in interpreting Sorensen's data on ICM efforts which contain a mixture of programmes, policy statements and feasibility studies.

Notwithstanding the increasing global number of ICM efforts, there is a paucity of data on their success; for example, Cicin-Sain and Knecht's comment that their 1996 survey results provided scanty evidence on the extent of ICM implementation and effectiveness. They concluded that it was very difficult to produce a model of successful ICM because there was a lack of objective evaluation data for any of the ICM examples they described (Cicin-Sain and Knecht 1998: 294). Sorensen (1997) comments that there is uncertainty in our knowledge about the important implementation phase of ICM programmes which comes after the adoption phase of various plans and policies. Burbridge (1997) makes a distinction between developing ICM initiatives and assessing their success in meeting stated goals and he suggests that there are not many good examples of fully developed ICM strategies, plans or practices that extend beyond a local or problem-specific level (Burbridge 1997: 181). It is also important to note that there are significant socio-economic and political differences between coastal nations which need to be considered in assessing ICM achievements.

Thus the concept of ICM is internationally accepted as an appropriate method for sustainable management of coastal resources. The concept has evolved from a realization that a sectoral approach to coastal management is fundamentally flawed. There is now a global proliferation of what has been loosely termed ICM efforts although there is scope for more rigorous survey of these. Finally, it is clear that the international acceptance of ICM as an approach is not matched by definable criteria or models to judge its success or best practice.

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NICK HARVEY

INTERDUNE

Interdune areas are the depressions between dunes. Interdune areas occur in a variety of shapes, which reflect the morphology of the dunes with which they are associated. Fields of crescentic dunes typically have ellipsoidal interdune areas that are elongate with the dune crestline. Extremely long interdune corridors occur with linear dunes, whereas more irregular, interconnected interdune areas accompany complicated dune shapes such as star dunes. There is a complete gradation from interdune flats that cover a greater portion of the field than the dunes, to interdune depressions between the lee and the stoss (upwind) slopes of adjacent dunes.

At least for transverse dunes, the formation of an interdune area can be visualized by watching a fixed point at the base of the stoss slope of a migrating dune. The interdune area begins at this erosional point and extends downwind as an interdune BOUNDING SURFACE as the stoss slope migrates. The surface continues to extend downwind until it is buried by the lee deposits of the next dune upwind, which has now migrated to the fixed point.

Although easy to visualize, the explanation for interdune areas remains controversial, and is intimately linked with the explanation of how dunes come to be regularly spaced. In one hypothesis, dune spacing is a function of the fluid dynamics established with airflow over dunes. With transverse dunes, there is flow separation at the dune brink, creating a separation cell in the immediate lee, and flow reattachment at a distance of a few dune heights from the brink (Walker and Nickling 2002). At the reattachment point, a new BOUNDARY LAYER is created. Wind speed and shear stress increase downwind within this new boundary layer, owing to the downward flux of momentum from the overlying, relatively high-speed flow. As long as the surface winds are accelerating, there is at least potential deflation of the interdune area (Kocurek *et al.* 1991). The boundary layer, however, is also expanding vertically, and at some point this expansion overwhelms the effects of the momentum flux, and wind speed and shear stress within the boundary should decrease. At that point, deposition (and initiation of a new dune) can begin. In this way, the spacing of the next dune downwind and the length of the interdune area is defined by the fluid dynamics established by the upwind dune. Field data, however, show that flow speed and shear stress reach a steady-state condition that favours sediment bypass, and not deceleration and deposition (Frank and Kocurek 1996). Adequate data do not yet exist to evaluate the role of turbulence, which may play a more significant role in the spacing of subaqueous dunes (Nelson *et al.* 1995).

In an alternative view, fields of dunes are viewed as self-organizing and the interdune areas exist by default. Computer simulations have modelled the formation of all major dune types as a function of the direction and duration of sand-transporting winds, and independent of the creation of an internal boundary layer (Werner 1995). In this hypothesis, dunes begin as small, randomly spaced collections of sand that merge and grow larger as they migrate, progressing to a steady-state condition where little change is possible because all the dunes are about the same size and, therefore, migrating at about the same speed. Initiation of sand dunes and their development to a steady-state field on Padre Island, Texas, strongly resemble the model (Kocurek *et al.* 1991). The interdune areas in this model are then simply the areas where the dunes are not (Werner and Kocurek 1999).

Regardless of their origin, interdune areas contain some of the most diverse features seen in dunefields. The first division is between those in which accumulation occurs, and those that are deflationary, in which older dune accumulations, reg or bedrock can be exposed. For deflationary interdune areas, or ones in which sediment bypass only occurs, the interdune flats essentially form a continuous surface over which the field of dunes migrate. Characteristic of dunefields where the water table controls the level of deflation (wet aeolian systems), corrugated surfaces develop on the interdune floor that reflect the stratification types of underlying dune accumulations.

Where sediment accumulation occurs on the interdune surface, these sediments can be classified as dry-, damp-, and wet-surface deposits (Kocurek 1981). Dry-surface, sandy accumulations are typically those from wind ripples, grainfall from the upwind dune and satellite dunes. Because the surface is dry and subject to deflation, these interdune accumulations within dry aeolian systems are usually restricted to interdune depressions that exist within the low-wind-speed separation cell. Extensive dry interdune flats, however, are common and these may contain concentrations of less-easily transported coarse grains, typically organized into granule ripples and zibar dunes. Sandy deposits are usually concentrated around sediment-trapping vegetation.

Where the interdune surface is wet or damp, typically because the water table is near the surface, a major division occurs between those that contain evaporate minerals and those that do not. Interdune areas in temperate climates (e.g. coastal fields) typically lack evaporites and the deposits are dominated by ADHESION structures, wrinkle marks, fluid-escape structures and subaqueous ripples, mud drapes, algal mats, channels and other features formed during ponding. In arid climates, wet- or damp-surface interdune areas are nearly always characterized by evaporites. These may occur as interdune SABKHAs, which show a characteristic crinkly sediment texture that results from deposition by salt ridges. Because interdune surfaces in wet aeolian systems have a high capillary water content or cementing evaporites, accumulation can occur on these flats in spite of flow acceleration within the interdune boundary layer, and commonly these interdune areas contain the finest sediment within the field.

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SEE ALSO: adhesion; boundary layer; bounding surface; dune, aeolian; sabkha

GARY KOCUREK

INTERFLUVE

Interfluves are areas of relatively high ground that lie between adjacent valleys in a drainage basin. In its most literal sense of land 'between rivers', the term interfluves refers to undissected ridges that lie between streams. However, it is often used in a more general sense to describe generally high terrain (including lower order streams and/or VALLEYs) lying between major river systems. It has also been applied to land between glacial valleys and subglacial streams. In addition, the concept of interfluves also has application in studies of ancient sedimentary sequences, where it is used to distinguish paleouplands from incised valleys.

In many parts of the world the spacing between ridges and valleys is remarkably regular. For example, the coastal hills of northern California, USA, contain many examples of ridge–valley topography in which first-order DRAINAGE BASINs flanking a relatively linear main valley are quite evenly spaced (e.g. see figure 1B of Dietrich *et al.* 2003). This is also true, for example, of hollows along quartzite ridges in the ridge-and-valley province of Pennsylvania, USA. On a larger scale, many linear mountain belts show a strikingly regular space of main valleys (Hovius 1996). Neither the origins of this regularity nor the controls on valley spacing are well understood, but some clues emerge from analyses of process dynamics. Many have argued, for example, that the transition from hillslope to valley (in the absence of obvious lithologic or structural controls) reflects a switch in process dominance (Gilbert 1909; Smith and Bretherton 1972; Tarboton *et al.* 1992). Valleys, according to this argument, represent landscape zones dominated by processes that are fed by upslope surface or groundwater discharge, and whose effectiveness therefore grows with increasing drainage area (sometimes abruptly, as with Horton's (1945) concept of an OVERLAND FLOW erosion threshold). Such 'concentrative' processes can include overland flow erosion, groundwater (see GROUNDWATER) sapping, glacial scour and some forms of landsliding (see LANDSLIDE). Closer to drainage divides, processes that do not have this upstream dependence (e.g. mass movement processes such as SOIL CREEP) hold sway.

Theories based on this process-dominance concept (or the related process-threshold concept; see Kirkby 1994) successfully explain observed relationships between gradient and drainage area around valley and channel heads (Kirkby 1987; Montgomery and Dietrich 1989). They do not, however, account explicitly for the spacing between ridges and interfluves, nor do they explain the instabilities that lead to spontaneous formation of channels and valleys in the first place. The problem of spontaneous formation of channel networks was first analysed mathematically by Smith and Bretherton (1972). Using linear stability analysis, they demonstrated the conditions under which spontaneous formation of incipient channels could occur under steady, uniform sheet flow on an undissected slope. One useful result of this analysis was the notion that there will be a tendency toward CHANNELIZATION when sediment transport capacity increases more than linearly with discharge per unit contour width, so that two units of flow together carry more than twice the sediment of one. The simplifications in the Smith–Bretherton analysis, however, did not allow for prediction of the incipient spacing between channels. This shortcoming was overcome in later work (Loewenherz 1991; Izumi and Parker 2000). For example, Izumi and Parker (2000) used linear stability analysis to predict incipient channel spacing in the case of sub-critical sheet flow (see SHEET EROSION, SHEET FLOW, SHEET WASH) on a convex-upward surface of cohesive sediment. Their analysis predicted that incipient channel spacing should depend on flow depth and roughness. Using reasonable values for these parameters they were able to account for incipient channel spacing on the order of tens of metres, a value not too dissimilar from observed channel spacings of the order of 100 metres (Montgomery and Dietrich 1989; Dietrich and Dunne 1993). Numerical models of drainage basin development have also been able to reproduce hillslope–valley topography and have revealed some important controls on hillslope scale and valley density (see, for example, Willgoose *et al*. 1991; Howard 1997; Tucker and Bras 1998).

As noted above, the concept of interfluves often refers not just to individual ridges, but also to much larger areas of land lying between incised valleys. An example is the Colorado Plateau region, USA, where expanses of high-elevation, low-relief terrain are cut by abrupt, steep-sided canyons (the largest being the Grand Canyon in northern Arizona). These less-dissected upland surfaces, though still subject to active fluvial erosion, can be considered interfluves by virtue of their lesser degree of incision.

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SEE ALSO: drainage density; hillslope-channel coupling; hillslope, form; hillslope, process; valley

GREG TUCKER

INTERMONTANE BASIN

Topographic basins of various shapes and sizes are common within many uplands and low to medium-height mountains, not affected by very recent differential tectonics. They are surrounded by higher terrain on all sides and drained by rivers which typically leave the basin floor through a narrow valley. Bounding slopes are often steep and there may exist a sharp junction between the floor and the marginal slope. In the majority of cases active tectonics does not contribute to the development of the basins, hence they are features due to differential denudation. Little accumulation normally takes place within intermontane basins, except for thin alluvium along river courses and localized slope-derived deposits.

The location of intermontane basins is often influenced by lithology and structure. Basin outlines tend to follow structural lines in the bedrock, e.g. ancient faults or regional fractures, or lithological boundaries, or master joints cross each other within the basin floor. The origin of the basins can then be ascribed to selective weathering and erosion, which exploit weaker rock compartments more effectively. For example, Thorp (1967) demonstrated the existence of structure-controlled basins in granite massifs of Nigeria.

However, Bremer (1975) maintains that many basins in tropical areas cannot be explained by unequal resistance and applies the concept of divergent weathering, which is topographically rather than rock-controlled. An initial depression receives and holds more water than its drier surroundings, hence weathering mantle develops, increasing in thickness and maturity. Occasional stripping of the SAPROLITE from the adjacent slope exposes the bedrock surface, which remains dry and sheds more water into the depression. This enhances local contrast in the intensity of weathering and leads to progressive deepening of the basin and steepening of its bounding slopes, as weathering products are episodically eroded away. According to Bremer, lateral enlargement of the basins plays a minor part. Basin floors may be flat and retain a weathering mantle, or they may have some relief moulded at the WEATHERING FRONT. Low hills, groups of boulders and occasional tors may occur. Thus, basins develop through localized deep weathering and are examples of long-term, two-stage landform development.

Basins developing in the way outlined above are not to be confused with basins forming basin-and-range terrains. The latter are typical tectonic features and originate due to downfaulting of crustal blocks in areas subjected to extensional regime. Moreover, they carry a thick fill of sediment washed down from surrounding uplifted terrain.

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PIOTR MIGOŃ

INVERTED RELIEF

A type of surface morphology within which former valley floors constitute the most elevated parts of the landscape whereas pre-existing valley sides and divides have been lowered to the extent that they now form topographic lows. Relief inversion occurs when materials on the valley floors are, or become, more resistant to erosion than those underlying the adjacent slopes. This may be the case with lava flows filling concave landforms, and with DURICRUSTS developed through the induration of alluvial or lake sediments. Silcretes, calcretes and ferricretes particularly often occur in inverted position.

Typical landforms due to relief inversion are sinuous flat-topped ridges indicating former valley courses. As erosion proceeds they are reduced to isolated elongated plateaux or MESAS capped by a remnant of the more resistant material. Recognition of inverted relief may have important economic implications as former placer deposits would now occur in elevated position.

Large-scale relief inversion is often hypothesized in folded terrains (see FOLD). It is often observed that drainage lines follow axes of anticlines whereas synclines underlie ridges on both sides of a valley which seems to be in contrast to original relief. The explanation holds that once erosion reaches the softer core of an anticline, it accelerates and outpaces any lowering of adjacent synclinal structures, leading to relief inversion. Erosion hollows of the Negev Desert in Israel (*makhteshim*) are examples of this kind of inversion. A more complicated scenario invokes planation of an original mountain range in the first phase and subsequent development of inverted relief at the expense of the planation surface.

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PIOTR MIGOŃ

ISLAND ARC

Around the western edge of the Pacific, many islands are arranged as festoons of arcs, some distance from the continent, in either single or double rows of islands. Other island arcs are the Indonesian, Caribbean, Middle America and Scotia Arcs. The arc radius ranges from about 4,000 km for Java–Sumatra to 1,500 km for Japan, and only 250 km for the hairpin Banda Arc.

The arcs face different directions: the Aleutian Arc is convex to the south, Sumatra–Java to the west, Mariana to the east, and Papua New Guinea to the north. Some ‘straight arcs’ such as the Solomon Islands and the Tonga–Kermadec Trench have many of the attributes of arcs without the curvature.

Some arcs are separated from the continent by backarc basins (Japan). The Scotia Arc has no backing continent, only the Pacific seafloor. The Caribbean Arc is backed by the straight Middle America Arc. The Mariana Arc is separated from the Philippine Arcs by a complex of backarc basins.

The standard arrangement of landforms in a simple arc is: continent; backarc basin; arc – forearc basin; accretionary prism; trench; ocean (Figure 94). The simplest arcs are lines of active volcanoes (e.g. Kurile Arc), which may be subdivided on the basis of underlying rocks into volcanic or continental based. The commonest type of double arcs comprises two rows of volcanoes, one older than the other. The older, usually outer line may be wholly or partially covered by limestone. The Sumatra–Java Arc is volcanic, with an outer arc of sedimentary rocks.

Topographically, the front of an island arc is generally rising while the back of the arc is sinking. In New Britain the front of the arc is rising, with flights of uplifted coral terraces, while the other side has drowned coasts, indicating sinking. Many arcs have large, mountainous islands built of sedimentary, metamorphic and granitic rocks (Japan, Papua New Guinea). They were eroded to a planation surface before the modern mountains were uplifted. The Virgin Islands display an erosion surface at about 300 m, and many terraces. Uplift occurred within the past 2 million years.

On the ocean side of arcs lie deep trenches. The greatest known depth in the ocean is the Mariana Deep (11,035 m). Traverses of trenches reveal numerous normal faults and graben. Trenches contain variable amounts of sediment: part of the Chile Trench is virtually empty; the Aleutian Trench contains up to 4 km of undeformed horizontal layers of sediment. An area of relatively shallow water separates the islands of the arc from the trench bottom. This is called the arc–trench gap. It is usually over 100 km wide, and is 570 km in the eastern Aleutian Arc. It is underlain by thick, generally horizontal sediments of the forearc basin.

At the outer edge of the forearc basin there is a marked break in slope, and the steep slope bounding the trench may constitute an accretionary wedge. The accretionary wedge (or prism) is a package of highly deformed sediment, and perhaps oceanic basalt, presumed to be scraped off the downgoing slab and accumulated at the edge of the overriding slab. Alternatively, the deformation structures can be interpreted as gravity-tectonic structures with décollement (also called detachment faulting) and thrusting. The whole area from the volcanic arc to the ocean, including the trench, is called the forearc. An alternative plate tectonic scenario is that the sediment, instead of being scraped off, is subducted beneath the arc.

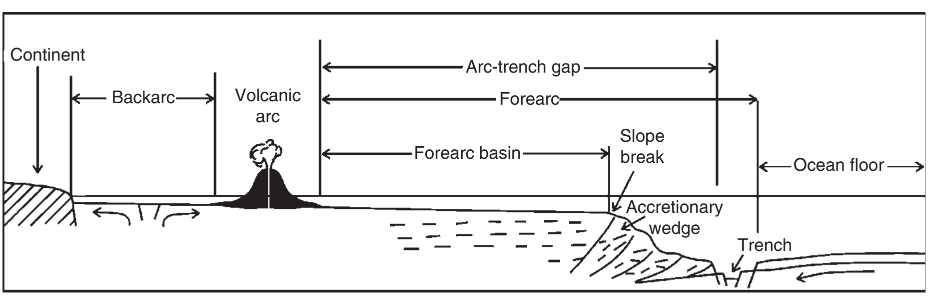

Figure 94 Nomenclature of features in a typical island arc

Behind island arcs are shallow seas known as backarc basins. Backarc basins are spreading sites (Taylor 1995). The Japan backarc basin has up to 2 km of sediment. The curved shape of the arc is not easily related to the angular pattern of multiple spreading sites behind the arc (e.g. Scotia Arc). Nor is the curve related to the approaching Pacific Plate: part of the Aleutian Arc is parallel to the movement, and part is almost perpendicular.

Active volcanoes on arcs frequently erupt with great violence, as at Krakatau (1883), Mont Pelée (1902) and Mount Pinatubo (1991). Stratovolcanoes are the commonest type, with numerous calderas. Andesitic volcanoes predominate in arcs, but some notable volcanoes are basaltic, including Mt. Fuji, Japan. Basalt makes up 70 per cent of the South Sandwich Islands. Arcs with volcanoes were formed in the past, as they are today, sometimes making double arcs. In Papua New Guinea the New Britain Arc has Quaternary volcanoes on the inside and the Palaeogene arc on the outside, but the New Ireland Arc has inner Palaeogene volcanoes and an outer arc of Quaternary volcanoes.

In some arcs (New Hebrides, Middle America) normal faults break the Earth's surface into fault blocks. Normal faults are also common in the trenches and backarc basins. Strike-slip faulting is important in some arcs, with lateral displacements of many kilometres along faults roughly parallel to the arc. These faults are a major feature in the geomorphology of Sumatra and Java. The Middle America Arc is offset in Nicaragua by a strike-slip fault oblique to the arc, which is a continuation of the Clipperton Fracture of the Pacific floor.

The American arcs are somewhat different from the rest. The Caribbean Arc lies east of the Caribbean Plate. This is bounded to the north by what looks like streaked-out bits of North America, consisting of continental rocks and making the islands of the Greater Antilles as far as Puerto Rico. The southern side is similarly made of rocks like those of South America. The true arc is the north–south part of the Leeward Islands, a typical double arc with the Limestone Caribees to the east and Volcanic Caribees to the west. The western side of the Caribbean Plate is bounded by the Middle America Arc, with the Middle America Trench on the Pacific side. The trench has no accretionary prism, and sediments are horizontal with no evidence of compression. The arc has some old continental rocks, and on the western side old volcanic rocks. Topographically the arc is a plateau tilted up to the west, but there is a down-faulted strip on the Pacific side, and this is where the current volcanic arc lies. The Isthmus of Panama is not part of the volcanic arcs, but consists of block-faulted continental rocks.

The Scotia Arc lies east of the Scotia Plate, which has continental rocks from South America streaked out on the northern side, and Antarctic rocks on the southern side. The South Sandwich islands make a true arc, with a spreading site behind that separates the Scotia Plate from a much smaller Sandwich Plate which is less than 8 million years old. The islands are volcanic, rugged and glaciated. The Scotia Plate is separated from the Pacific by a transform fault.

A negative gravity anomaly lies over trenches, and a positive anomaly on the continental side about 115 km distant. Arcs are characterized by earthquakes. The foci sometimes appear to fall in a zone (called the Benioff zone), about 50 km thick and reaching depths of several hundred kilometres. This dips towards the concave side at varied angles, and is vertical under the Mariana Arc. It is presumed to mark a slab being subducted at the trench. In some arcs (e.g. Aleutians) the Benioff zone emerges not at the trench, but on the arc–trench gap. The relation between arc, trench, gravity anomalies, earthquakes and volcanoes is in fact quite variable.

Some arcs run aground onto continents. The Sumatra Arc goes via the Andaman Islands to Burma (Myanmar), and continues as the land-bound arc of the Himalayas of about the same size. The Aleutian Arc continues in Alaska; and the southern arm of the Caribbean Arc in Colombia. This is significant in theories that relate island arcs to mountain building, a major concept in PLATE TECTONICS. Since backarc spreading increases the space between continent and arc it might seem difficult to make the two collide, but Hamilton (1988) wrote that island arcs are 'conveyor-belted' towards subduction zones, so that island arcs collide with one another and with continents. Island arc concepts are often used to interpret mainland features including ancient structures and modern mountains such as the Apennines.

The ruling theory of plate tectonics explains island arcs as places where oceanic crust is subducted at trenches. This produces the paradox of having a compressive mechanism, when the normal faults in trenches, islands and backarc basins all indicate tension. Other suggested mechanisms of arc formation include geotumours, mantle diapirism, subsided blocks at the trench, and surge tectonics.

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SEE ALSO: plate tectonics

CLIFF OLLIER

ISOSTASY

A term introduced in 1882 and derived from the Greek words *iso* and *stasis* that means equal standing. It is used to describe a condition to which the Earth's crust and mantle tend, in the absence of disturbing forces – a condition of rest and quiet in which the lighter crust floats on the dense underlying mantle (Watts 2001). However, isostasy is disturbed by such processes as sedimentation, erosion, volcanism and the waxing and waning of ice sheets (see GLACIAL ISOSTASY). In its simplest form the isostasy concept envisages that rigid blocks of crust, buoyantly supported by the underlying fluid medium of the mantle, are free to move vertically until their weight is balanced by their buoyancy ('isostastic equlibrium'). Differences in the density and thickness of the crust are to a considerable extent responsible for variations in the isostatic adjustment of the lithosphere. If it is in isostatic equilibrium, one portion of the lithosphere will stand higher than another, either because of a lower density (the so-called Pratt model) or because it is of the same density but thicker (the Airy model), or from a combination of both.

Isostatic equilibrium is disturbed by various geomorphological processes. Erosion makes crustal blocks thinner and lighter in weight so that an eroding mountain will tend to rise to maintain equilibrium. Conversely, deposition of sediments, as for example in a delta, represents an added load, so that sinking tends to occur. Extension of the crust by rifting thins it whereas compression in mountain-building thickens it. Loading and unloading by waxing and waning glaciers also affects isostatic equilibrium as do changes in the volume of water in the ocean basins or in lakes – hydro-isostasy (Bloom 1967). Similarly, the extrusion of large amounts of volcanic material can weigh down the crust and cause subsidence to occur.

Isostasy is central to understanding a range of geomorphological phenomena. For example, areas that were once loaded by great ice sheets (e.g. Fennoscandia) are now areas of uplift as they recover from the load of the ice. Conversely areas peripheral to the ice sheets, which bulged up during glacials, are now areas of subsidence (e.g. the southern North Sea area). Likewise, the development of GUYOTs, seamounts and atolls in the Pacific may be related to subsidence caused by volcanic eruptions loading the crust (McNutt and Menard 1978). Load-induced vertical movements have a profound effect on deltas. In particular, subsidence increases the overall water depth and so increases the accommodation space that is available for the prograding sediments. The relief characteristics of rift valleys, features caused by crustal extension, also show the effects of isostasy through their association with broad topographic swells. The rift flanks of passive margins show evidence of uplifted flanks and erosional unloading (as in Namibia) and the presence of rift flank uplifts may explain the deflection of drainage systems towards the continental interiors (e.g. the Kalahari) (Gilchrist and Summerfield 1990). The shorelines of great palaeo-lakes, like Lake Bonneville in the southwest USA, have deformed as the crust has adjusted to the removal of the weight of the lake as it desiccated (Crittenden 1967).

The continental shelves were depressed as the weight of water from rising postglacial sea levels (the Flandrian Transgression) was applied to them. Finally, erosion that deepens and widens river valleys but does not erode the peaks to the same degree, will reduce the mass of an area and so drive isostatic uplift. As a consequence, the altitude of the peaks could increase at the same time that the mean height of a region is decreasing (Molnar and England 1990).

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A.S. GOUDIE

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J

JOINTING

Joints are cracks in rocks formed by stress that results from tectonic events, cooling, or isostatic rebound. They range in length from millimetres to kilometres. In outcrop, joints can be small hairline cracks only millimetres in length or long open fissures a metre or more across. They may be open or filled; common fillings are soil and clays. Joints can also be sealed as a result of hydrothermal activity. They are distinguished from faults (see FAULT AND FAULT SCARP) by the lack of movement between the two sides of the joint. Most joints are tension fractures.

Joints occur in sets, groups of nearly parallel fractures that formed under the same stress regime. At least three joint sets commonly occur in most outcrops: in sedimentary and schistose/foliated metamorphic rocks. There is typically one set parallel to layering and two mutually perpendicular sets normal to layering. In massive rocks like granite, one set is usually horizontal and the others are vertical or steeply dipping. Columnar joints in basalts (e.g. the Giants Causeway, Northern Ireland) are a special example of cooling joints. These joints bound areas where cooling fronts meet so that a hexagonal crack pattern forms where the boundaries converge.

Joints exert significant control on landform shape. Most granite landforms, such as domes (e.g. Stone Mountain, Georgia, USA and Ayers Rock, Australia) and TORs (e.g. Haytor Rocks, Dartmoor, UK), are joint controlled. Other joint-controlled landforms not dependent upon lithology include geos and gulls. A geo is a deep, narrow cleft or ravine (see GORGE AND RAVINE) along a rocky sea coast that is flooded by the sea. Valley orientation in general often parallels major joints. A gull is a joint that opens on escarpments because of tension produced by cambering (see CAMBERING AND VALLEY BULGING). Grikes in LIMESTONE PAVEMENTs (e.g. the Burren, Co. Clare, Eire) are joints enlarged by solution that separate clints, the raised portions of these pavements.

Joints are also very important with respect to both MECHANICAL WEATHERING and CHEMICAL WEATHERING, and slope stability. Joints are the most important zones of weakness in any given rock mass, and provide primary access for moisture to enter rock. A dense pattern of closely spaced joints will thus hasten chemical weathering, leaving upstanding areas where joints are more widely spaced. The WEATHERING FRONT often occurs along a horizontal joint. In addition, most limestone CAVEs originate as joints, and sink holes commonly occur at joint intersections. Corestones are remnants of joint blocks, often occurring in a matrix of weathered debris or as boulder fields after the weathered matrix has been stripped. Furthermore, the stability of a given slope can often depend on the orientation of the joints. Sliding is likely to occur if the joints dip toward the slope face, for instance, and toppling occurs along vertical joints oriented parallel to the slope. Joints also commonly form the slip surface in larger rotational landslides. DRAINAGE PATTERNs are also often controlled by the joint pattern of the underlying rocks, e.g. rectangular drainage patterns in limestone, and drainage density may reflect joint density.

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SEE ALSO: bedrock channel; bornhardt; granite geomorphology; inselberg

JUDY EHLEN

K

KAME

The term kame is derived from the Scottish word *kaim* and was introduced by Jamieson (1874) to mean a short, steep-sided mound or ridge of water laid sands and gravels deposited from melting ice. As a single term it cannot be applied to a specific landform, depositional process or sediment type. A variety of landforms resulting primarily from glacifluvial deposition in association with melting and buried ice are now recognized as kame forms. They include kames, kame terraces, kame complexes, and kame and kettle topography. These landforms occur in subglacial and ice-marginal environments (Holmes 1947; Price 1973; Bennett and Glasser 1996; Benn and Evans 1998).

Kames form subglacially or ice-marginally as small hills or ridges that may occur as isolated features or in a group of similar forms. Some form through deposition of sand and gravel at the base of a glacier by meltwater descending from the ice surface in a moulin, or from an englacial water body. Stream deposition within a channel bounded by ice walls may produce a short kame ridge. Other kames form as small delta-kame surfaces at the ice margin or terminus, and frequently contain beds of sand and gravel, debris flow diamicton and laminated silts and clays. Melting of adjacent and underlying ice results in normal faulting of the sediments on mound margins and sagging and folding of beds. Clay and silt beds often have slump, flow and load structures.

Kame terraces are usually formed either at the margin of the glacier between the debris-mantled decaying ice and the valley slope, at the terminus of the glacier along the ice front or between the decaying ice and an obstructing moraine. A marginal kame terrace surface may slope slightly down-glacier or towards the valley slope. The primary terrace may be fragmented into several sections due to collapse of underlying ice. Several kame terraces can be developed on a slope as the glacier margin recedes. The depression formed between the ice margin and valley slope diverts the meltwater laterally along the ice margin. The depression may be temporarily filled with water to form a narrow marginal lake. Glacifluvial sediment transported along the ice margin and from the ice edge is deposited either onto the buried ice on the floor of the depression or into the lake, or both.

The sediments in the kame terrace vary rapidly in texture and structure laterally and vertically. They consist predominantly of horizontal to gently inclined alternating beds of sand and small gravel due to glacifluvial deposition, but may exhibit laminated silts and deltaic foreset beds due to deposition in relatively deep water. On the ice proximal side of the terrace interbedded masses of debris flow diamicton (flow till) from the ice margin may occur while on the valley side paraglacial deposition of alluvial fans and debris flows can occur due to high discharges and snowmelt in adjacent valleys and on slopes. Melting of buried ice during and subsequent to sediment deposition results in formation of kettle holes that are more frequent near the former glacier margin and sometimes contain small lakes. Removal of support from the sediment by melting of marginal and buried ice results in collapse of the terrace edges to give a crenulated form, due to debris flows and landslides in the ice contact zone. Normal fault, fold, slump and flow structures may be formed within the sediments.

Kame terraces of the terminal zone usually form steep-sided ridges or steeply bounded low plateau surfaces. They may form where ribbon lakes are developed along the ice edge and are filled primarily with glacifluvial sands and gravels. They can also form where a small lake is impounded between the decaying ice and a bounding moraine. A small stream may flow from the ice bringing sand and gravel but much of the sediment consists of laminated silts interbedded with flow diamictons from the ice surface. Melting of underlying and removal of supporting marginal ice usually results in the sediments exhibiting fault, fold, slump, flow and load structures.

Kame complexes occur as a number of steep-sided mounds or ridges usually within a limited area. They result from the deposition of sands and gravels by meltwaters descending from within the ice to the glacier bed where the deposits accumulate in numerous cavities. Since deposition of kame complex sediments mainly takes place subglacially, the glacifluvial cores of the ridges are often discontinuously draped by diamicton resulting from meltout and flowage from the overlying ice during its final decay. The sediments may show faulting, folding, sag, slump and flowage structures.

Kame and kettle topography is recognized in a landscape by the occurrence of many steep-sided hills or short ridges closely juxtaposed and separated by relatively deep circular, oval or elongate depressions. Several depressions may contain ponds. Kame and kettle topography differs in origin from kame complexes by the glacifluvial sedimentation having occurred over abundant buried ice at and beyond the glacier margin. Ten to over twenty metres of glacifluvial sands and gravels are often deposited and the dips and directions of the beds are highly variable being related to deposition into depressions in locally stagnant ice. Melting of the buried ice causes normal faulting and folding of the sediments and a characteristic inversion of the topography with the kames occupying the former sites of deposition and the kettle depressions and lakes the sites of the thickest buried ice masses.

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KAOLINIZATION

Kaolin ($Si_2Al_2O_5(OH)_4$) is a clay mineral which can be produced by prolonged weathering during which chemical elements are progressively lost. The more complex clay montmorillonite ($Al_{1.7}Mg_{0.3}Si_{3.9}Al_{0.1}O_{10}(OH)_2$) is comprised of not only the basic crystal lattice structure of aluminium and silicon atoms but it also contains magnesium. Kaolin is simply comprised of the more resistant silicon and aluminium, the more soluble elements such as magnesium having been lost through weathering. Thus kaolinization represents a process of the loss of the more soluble chemical elements producing a much simpler crystalline clay over time and can be produced by the prolonged weathering of feldspars, micas and other primary aluminosilicate minerals which may originally have contained magnesium, calcium, sodium, iron and potassium. Kaolinization thus refers to the changes from primary rock aluminosilicate minerals through weathering and their transformation to kaolin as a residual clay mineral.

Kaolin is formed under conditions of slight acidity and free drainage which remove most of these other cations. Free drainage and high rainfall are important factors in the removal of the soluble cations and in areas with annual rainfall of below around 500 mm montmorillonite clays dominate, between 500 and 1,500 mm kaolin clays dominate, with iron oxides also occurring at rainfalls above 1,500 mm (Thomas 1974). The SOLUBILITY of kaolinite is least at pH 6 but increases both as pH rises and falls from this value; so as pH departs from 6 and under strongly oxidizing conditions in the humid tropics, intense weathering tends to remove most of the silica, leaving the oxides of iron or aluminium as residual minerals rather than kaolin.

The distribution of kaolinized mineral material through a vertical weathering profile characteristically shows an increase towards the surface with a concomitant decrease in primary minerals. In a study of weathering profiles in Ghana (Ruddock

1967), some 60 per cent of particles finer than 40 μm were kaolinite at a depth of some 5–6 m and this dropped to around 30 per cent at around 30 m depth while feldspar constituted 25 per cent of the particles over 40 μm in size at 30 m and was absent at 5–6 m. The zone of greatest kaolinization may be around the zone of the water table, with kaolin often accumulating at the base of the vadose zone. The presence of clays at this depth may reduce vertical permeability to water and thus facilitate lateral water movement.

It has been suggested that the presence of kaolin is indicative of past tropical weathering conditions. What is more the case is that kaolin is indicative of prolonged weathering with slightly acid conditions, high rainfall and good drainage to facilitate cation loss. While it is thus true that kaolin is present in currently tropical areas, the presence of kaolin is only indicative of prolonged weathering under appropriate and stable conditions. It should be remembered that kaolin is also produced by metamorphosis in proximity to molten magmas.

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KARREN

Karren are surface solutional weathering features of varied size and shape, found on karstifiable (see KARST) rocks, usually limestones, dolomites and dolomitic limestones, but including gypsum, rock salt and silicates. The French word is *lapié* but the German *karren* generates terms now widely accepted. Wherever karstifiable rocks exist there will generally be karren features. Early research, in the nineteenth century, focused on observations in Alpine Europe. Later knowledge of karren extended to many climatic zones. Considerable experimental and field research to examine form and conditions of formation has been carried out.

Karren extend from the nanoscale of individual features to landform complexes measured in metres or tens of metres, the latter often having smaller karren on their surfaces. This rock sculpturing by DISSOLUTION, involves complex, and, sometimes, complete surface patterning. Removal (negative) forms are fundamental, e.g. pits, rills, or runnels, created directly by process (Ford and Lundberg 1987). Other karren are remnant positive forms, e.g. flachkarren (clints) and spitzkarren. Microscale forms, e.g. pits and rillenkarren, are on a centimetre scale or less. Large solution runnels, carved pinnacles, karrenfelder, giant grikelands, LIMESTONE PAVEMENTs and other complex polygenetic assemblages lie at the other extreme. Features vary in basic shape, e.g. some are linear, others circular. The largest linear solution forms can be tens of metres long although the commonest are tens of centimetres long. Width of features is also usually in tens of centimetres. Restrictions on dimensional development are often due to downslope available rock distance. Depths likewise vary considerably due partly to topographic opportunity. Small-scale surface roughness varies, with some karren characteristically smooth, others spikey or rough (Plate 66). Karren-like forms are also found underground.

Karren development is principally affected by: the type and intensity of chemical processes attacking the rock; rock characteristics including intrinsic lithology, disposition (structure) and relation to surrounding topography; the nature of any cover material; and the time available, and the changes in conditions during that time, for processes to act.

Karren classification has been attempted. Some schemes examine processes: Bögli's (1960, trans. in Sparks 1971) involves the effects of the main carbonate solution processes, depending on whether the limestone is exposed to the air, partly

Plate 66 Spitzkarren in the Triglav area of the Julian Alps, Slovenia

covered, or entirely covered by soil. This cover factor fundamentally influences solutional processes on limestone, affecting the amount of CO_2 available for solution and the time period and speed of solution. Many karren are examples of BIOKARST as they are fundamentally the result of biological corrosion.

However, many karren are not simple genetically, for they reflect past as well as present processes, and many forms are polygenetic. Thus Ford and Williams's 1989 classification is subdivided to allow for genetic factors, rather than a purely genetic system. They retain descriptive karren terminology, finding form intuitively useful. They distinguish circular karren forms, those which are linear and controlled by structure, linear forms controlled hydrodynamically, and polygenetic forms. However, these distinctions may be blurred; for example, after karren initiate hydrodynamically, field evidence suggests that small-scale lithological structures affect development. Weaknesses affect fluid movement, and may interrupt karren, causing capture or change. The importance of lithology at the microscopic scale has been stressed by researchers such as Goudie *et al.* (1989). Slope significantly influences karren, especially the pattern and complexity of branching networks.

It is useful to describe the main forms. *Rillenkarren* are tiny gravitomorphic packed channels, starting from a crest and extinguishing downslope. Width is about 1 to 3 cm, and length a few tens of cm. They are separated by sharp ridges.

Rinnenkarren and *rundkarren* are Hortonian channels with an unrunnelled catchment surface above their commencement points. They enlarge downslope depending on available rock. Width varies but rundkarren often stabilize at 20 to 30 cm and rinnenkarren are narrower. High in the long section, rundkarren develop parabolic cross sections, deepening downstream before stabilizing at the bottom end. Occasionally overdeepening develops where the runnel flows into a grike. Rundkarren are smooth features with well-rounded crests, whereas rinnenkarren are distinguished by sharper crests. The former result from covered conditions and soil removal makes them visible. Rinnenkarren generate much debate, but are considered to be formed in free or, possibly, half-free conditions. Flow forms may meander (*meanderkarren*) depending partly on slope. More complex branching of rundkarren is found on gentler slopes.

Trittkarren are step or heel-print-like features, about 10 to 30 cm in scale, essentially features of very gentle slopes. Their headwall is arc-shaped, they are flat floored and they open downslope (Vincent 1983). Trittkarren appear related to the ripples forming from sheet flow across a gentle sloping rock surface (White 1988).

Spitzkarren are peak-shaped features remaining from surface solution widespread over horizontal or gently sloping surfaces. Their sides are carved by rillenkarren. Their diameter is typically 50 cm and their height about 10 cm. They tend to merge into other positive features, pinnacles in particular, and are essentially polygenetic (Plate 66).

Kluftkarren are clefts, fissures or grikes. These are the major splits into limestone surfaces formed by widening, deepening and eventually merging of small solution features developing along linear weakness in the rock. Mature examples may run considerable lengths, e.g. several metres. Depth varies with bed thickness. By definition grikes should have broken through at least the top bed of an outcrop. This distinguishes them from runnels, which may develop into grikes (Plate 67, and see LIMESTONE PAVEMENT, Plates 73 and 74).

Kamenitzas are solution basins or pans (see WEATHERING PIT). Size varies enormously, but small ones have flat floors and scalloped edges about 3 cm deep, with diameters of several cm or more. Many are tens of cm in diameter with the largest examples measurable in metres. Development is, however, limited, as when the pan becomes drained either at the surface over

Plate 67 Sloping limestone pavement (12° to 15°) on Farleton Fell, Cumbria, UK: showing strong karren formation on a wide range of clints across several limestone beds

a rim, or from beneath via solutionally opened cracks in the floor. The feature may become more complex; perhaps part of a staircased runnel, or of a complex hole into which several runnels drain. Kamenitzas occur in both free and covered conditions. True kamenitzas enlarge outwards at their rims where solution conditions remain ideal. Kamenitzas may merge into kluftkarren, or coalesce, leaving peaky sharp Spitzkarren.

Kluftkarren also merge into larger weakness-oriented features, e.g. bogaz, lapiés wells, etc. Several joints crossing focuses processes, resulting in mill-like features possibly several metres across and complex in form. This is demonstrated in limestone pavement areas in the UK, and in high mountain areas such as north Norway and the Alps, where large amounts of glacial meltwater may have enhanced them.

Terminology cannot encompass the full variety of natural sculptured forms. Features grade into each other, cut across, develop idiosyncratically according to local conditions and deteriorate when ideal formation conditions change. Destruction of rundkarren, formed under soil, illustrates this: on soil removal their smooth surface undergoes etching in subaerial conditions into sharper rougher features, e.g. kamenitzas or rillenkarren. Destruction may include mechanical weathering effects involving freeze and thaw, for example, and ends with broken rubble fields over underlying intact rock layers. However, another change in conditions could find water flow, or even a return to soil cover: both would alter the karren again. In karst areas such changes of local conditions can happen especially easily and quickly because of capture of active-process locations.

Rock type is fundamental to karren development. Strong, pure limestones or dolomites produce the best development. On rock salt, karren form easily due to high solubility but only persist in relative aridity. Gypsum karren are intermediate in persistence as gypsum's solubility is such that karren develop in humid conditions, but these conditions also favour their destruction. Silicate rocks are only slightly soluble; only in prolonged warm and wet conditions will karren, like other karst features, form. Sandstones in temperate areas may display weak karren but development is only significant in the wet tropics. Research has placed a general timescale for development of limestone karren in thousands of years, gypsum features in hundreds of years and rock salt karren in tens of years (Mottershead and Lucas 2001).

Other lithological factors influence particular karren forms. Research has considered karren initiation at tiny weaknesses or variations in the rock (Moses and Viles, in Fornos and Gines 1996). Chance unevenness allows rainwater to pond and start surface solution, simple plants may then develop. If soil develops so do higher plants and accelerated biological corrosion becomes very important, especially in warm, wet locations. The forms produced can be very striking. On bare rock, flow processes down any slope, however smooth, result eventually in flow-concentration into channels. A slight slope gives flow rather than ponding, a slight dip can give ponding before flow.

Sense can be made of karren by considering their form, mode of origin and development conditions. However, field situations demonstrate that, although 'perfect' examples of types are found, there will always be a wide spectrum, and merging of features is both possible and common even without changes in external factors such as climate.

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SEE ALSO: biokarst; dissolution; karst; limestone pavement.

HELEN S. GOLDIE

KARST

Karst is terrain with distinctive hydrology and landforms arising from a combination of high rock solubility and well-developed secondary porosity (Ford and Williams 1989). It is commonly associated with carbonate rocks such as limestone, marble and dolomite and is well known for features such as caves, enclosed depressions, fluted rock outcrops, underground rivers and large springs. It takes its name from a limestone region in the northern Adriatic inland of Trieste on the Slovene–Italian border where such features are particularly well developed. The region is called *kras* (Slovenian) or *carso* (Italian), but this was Germanicized to *Karst* in the period of the Austro-Hungarian Empire, when the first scientific studies were made of the region's geomorphology and hydrology. By extension, other areas with similar features are also referred to as karst, and this includes places where it is developed on other soluble rocks such as gypsum and rock salt (see GYPSUM KARST; SALT (EVAPORITE) KARST). This discussion focuses on karst in carbonate rocks. These outcrop over about 12 per cent of the ice-free continental areas (Figure 95), with well-developed karst covering about 7–10 per cent of the continental area. Karst also develops beneath the surface when karst rocks are interbedded with other lithologies; this is known as *interstratal karst*. Limestone (as opposed to karst) geomorphology refers to landscapes developed on carbonate rocks, and includes landforms that are not necessarily produced by karst processes (e.g. coral atolls, glacial troughs).

Cvijić (1960) defined different morphological types of karst, including *holokarst* where karst reaches its fullest development in thick carbonate rocks that extend below sea level, for example in the extensive limestones of the Dinaric region; *merokarst* where karst development is evident but rather poorly expressed, by virtue of not very suitable lithologies or rather thin limestones, such as in the chalk of northern France or in some areas of Jurassic limestones such as the Swabian

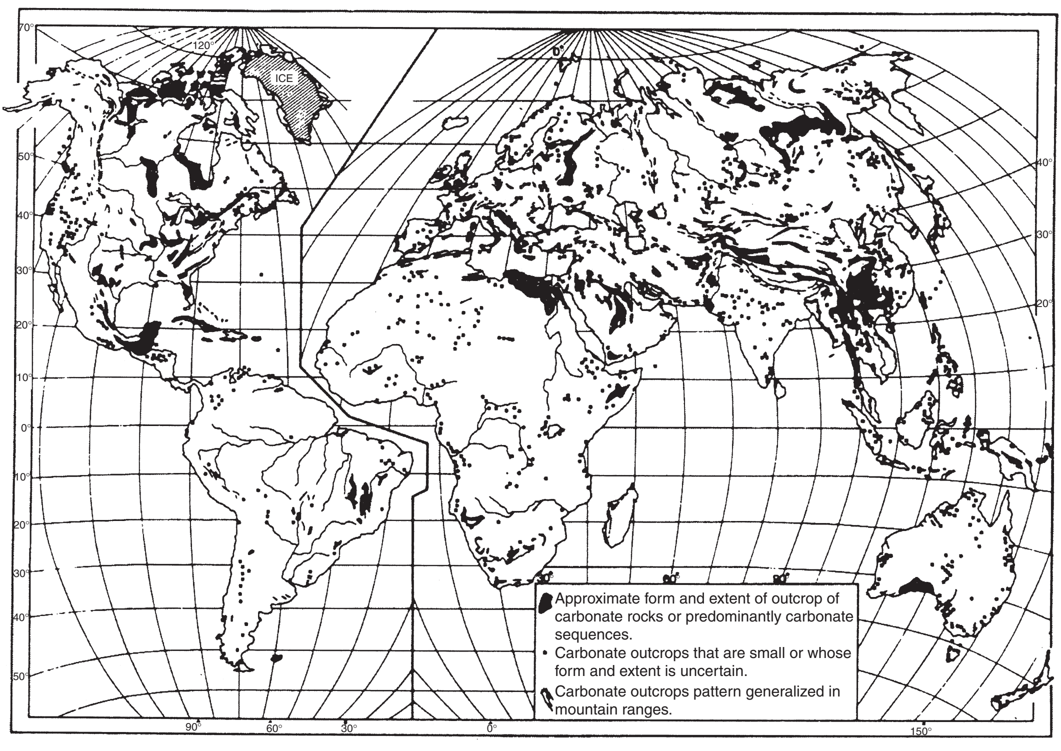

Figure 95 Global distribution of carbonate rock outcrops. Karst occurs over most of this area and also in subcrop beneath various coverbed lithologies (from Ford and Williams 1989.)

Alb of Bavaria or the Cotswold Hills of Britain; and transitional types where the carbonate rocks are quite thick and well karstified, but underlain by or interbedded with non-carbonate formations, as in the Causses of France. However, these terms are seldom used. When the imprint of now dry river valleys is evident in the landscape, it is sometimes referred to as *fluviokar*st.

The occurrence of pure carbonate rocks with high solubility is insufficient to produce karst, because the structure, density and thickness of the rock are also important. The carbonate rocks in which karst reaches its best expression are thick, dense, pure ($>$90 per cent calcium carbonate) and massive with low primary porosity, but with well-developed secondary porosity along fissures such as joints, faults and bedding planes. Even pure soluble rocks such as coral and chalk have relatively poorly developed karst, because their very high primary porosity (which can be 30–50 per cent) leads to diffuse groundwater flow, which is not conducive to extensive cave and closed depression development, though they have some. Impure carbonate rocks such as argillaceous limestones hardly support karst, partly because they tend to have high primary porosity but especially because insoluble residues inhibit the growth of secondary porosity by clogging groundwater pathways as they form. Nevertheless, there is a spectrum of rock types and degrees of purity, with a corresponding spectrum of karst development.

Chemical processes

The dominant process that produces karst features is the solution of the rock by rainwater. The chemical process is DISSOLUTION (or corrosion); in a carbonate karst context the process can be summarized as:

$$CaCO_3 + H_2O + CO_2 = Ca(HCO_3)_2$$

Calcium carbonate, $CaCO_3$, dissolves in the presence of water and carbon dioxide ($H_2O + CO_2 = H_2CO_3$ or carbonic acid) to yield the more soluble calcium bicarbonate, $Ca(HCO_3)_2$, which is readily transported away in solution.

The process of carbonate rock solution can be conceptualized as operating in a situation in which two hydrological and geochemical subsystems interact. The hydrological cycle provides the main source of natural energy that powers the system and drives the evolution of karst, because water is the solvent that dissolves karst rock and then carries it away in solution. Geochemical processes control the rate of dissolution (the speed with which solid rock is converted into ions in solution), which in a carbonate karst depends very strongly on strength of acidification by dissolved carbon dioxide during its passage through the atmosphere and soil layer before making contact with the limestone. The concentration of CO_2 in the open atmosphere is about 0.03 per cent by volume, whereas it is commonly 2 per cent in the soil and can even reach 10 per cent. A factor of 100 in the concentration of CO_2 results in ~5 times increase in the solutional denudation rate (White 1984). Although this is important, the amount of rainfall is even more significant, the wettest places in the world having the fastest rate of limestone solution. For example, limestone denudation by solution processes has been estimated as high as $760\,m^3\,a^{-1}\,km^{-2}$ (cubic metres per year per square kilometre of limestone outcrop) in very wet places such as parts of Papua New Guinea where rainfall can reach 12,000 mm per year, but as low as $5\,m^3\,a^{-1}\,km^{-2}$ in some arid zones like the Nullarbor Plain in southern Australia with rainfall of less than 350 mm per year. The amount of solutional attack on the limestone rock therefore depends on the concentration of the solute (determined by biogeochemical processes) and the volume of solvent (determined by the rainfall). The solute load of a karst spring is the product of its discharge and the concentration of limestone salts in solution.

Biochemical and physical processes associated with various organisms also assist in weathering limestones, especially in the intertidal zone, and produce a suite of landforms known as BIOKARST.

Landscape development

A conceptual model of the karst system is presented in Figure 96. Karst evolution is explained by White (1988), Ford and Williams (1989) and Gabrovšek (2002), and is summarized in Williams (2003). In order for major karst landforms such as enclosed depressions and caves to develop, the rock removed in solution must be carried right through the body of karst rock and be discharged at springs. Thus the development of an underground plumbing system is a necessary precursor to surface landform evolution. When a continuous conduit of 5–15 mm extends right through the rock, the drainage can become turbulent and

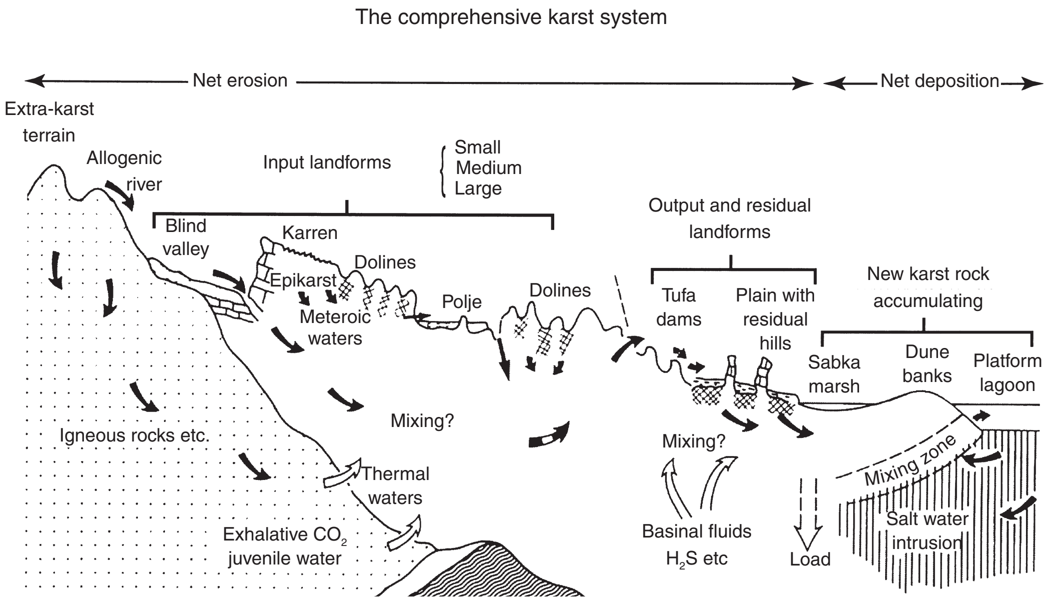

Figure 96 A conceptual model of the comprehensive karst system (from Ford and Williams 1989.)

then can begin to discharge fine insoluble particles. It takes the order of 5,000 years for a conduit of 1 km length to develop up to this size (White 1988). But once the hydraulic threshold of laminar to turbulent flow has been passed, cave enlargement proceeds more rapidly. Cave passages to 3 m diameter can develop in about 10^4–10^5 years.

The development of subterranean karst hydrology also depends on the manner in which water enters the karst. Rainwater that falls directly onto the limestone outcrop is known as *autogenic* recharge. It infiltrates diffusely into the rock via countless fissures. By contrast, rain that falls onto impervious non-karstic rocks but later flows onto the karst is known as *allogenic* recharge. It runs off as an organized stream, which sinks underground at the end of a BLIND VALLEY soon after encountering the limestone. Places where streams disappear underground are called *swallow holes*, *swallets*, *stream-sinks* or *ponors*. Sites where water cascades steeply into open pits are sometimes called pot-holes or abîmes (French).

Autogenic waters are mainly acidified by carbonic acid, but allogenic waters may have flowed from peat bogs (and hence contain organic acids) or have encountered sulphide minerals when draining from shales and hence contain sulphurous acid. They also tend to have a greater mechanical load, which can abrade limestone and help to incise cave floors. Streams sinking along the allogenic input boundary converge underground and emerge at springs at the output boundary of the system, thereby establishing dendritic subterranean drainage networks. The flow in these conduits is turbulent, as opposed to the much slower flow in tiny interconnecting pores and fissures which is laminar. Processes of CAVE network development (*speleogenesis*) are discussed by White (1988), Ford and Williams (1989), Gillieson (1996) and Klimchouk *et al.* (2000).

Some karsts have very extensive cave systems. The longest in the world is the Mammoth–Flint Ridge–Roppel–Procter System in Kentucky, USA. It comprises an interconnected system of essentially horizontal dendritic passages at different levels, totalling over 530 km in length, but developed in only about 100 m vertical stratigraphic thickness of limestone. The world's deepest known caves are Voronja Cave, Arabika massif, West Caucasus, at over 1,750 m deep, Reseau Jean Bernard (1,602 m) in France and Lamprechtsofen–Vogelschacht (1,535 m) in Austria.

Caves can be some of the world's oldest landforms, because they are located deep underground and so are protected for a long time from surface denudation. Thus sediments in the Mammoth Cave system have been dated by radioactive decay of cosmogenic isotopes to 3.5 Ma (Granger *et al.* 2001), and fossil hominid and animal remains in cave sediments in South Africa have been dated by palaeomagnetism to a similar age.

Once the input–output connections are established diffuse autogenic recharge infiltrates the bedrock. Most of the dissolutional attack takes place on bedrock just beneath the soil, close to where CO_2 is generated and consequently where the percolating water attains its greatest aggressivity. Thus up to about 90 per cent of the corrosion is accomplished in the top 10 m or so of the limestone outcrop. Since water penetrates underground mainly by means of joints and faults, these fissures become more widened by corrosion near the surface than they are at greater depth. The surface of karst is therefore very permeable, but permeability (the capacity to transmit water) decreases with depth. This highly corroded superfical zone is termed the EPIKARST (or alternatively the *subcutaneous zone*).

Before rainwater drains underground it flows across rocky outcrops on the surface. The resulting corrosion of these outcrops yields a small-scale solution sculpture of vertically fluted rock and widely opened joints, collectively known as *lapiés* (French) or *Karren* (German). Rocky spires produced in this way can sometimes be tens of metres high, although individual KARREN are normally smaller. These forms are particularly common above the tree-line where soil and vegetation are thin or absent, but karren also develop beneath a soil and vegetation cover. When the bedrock has been scoured by glaciation and stripped of loose debris, postglacial weathering produces bedding-plane surfaces with joints opened by dissolution. In northern England the whole surface is known as a LIMESTONE PAVEMENT, with the widened joints called *grikes* and the intervening blocks called *clints*. Large solutionally widened joint corridors are known as *bogaz* and complex networks of such features are sometimes known as labyrinth karst. Areally extensive expanses of any or all of these features, especially above the tree-line, constitute a *karrenfeld*.

Sometimes part of the carbonate dissolved near the surface is precipitated further down in pores and fissures in the bedrock. This is particularly common in porous limestones such as coral, and results in several metres of the bedrock near the surface being hardened by being made less permeable, a process known as CASE HARDENING. This is encouraged in hot climates in particular by evaporation near the surface. Carbonate crusts produced by secondary precipitation of carbonate are known as CALCRETE or *caliche*. Sometimes induration of the rock by case hardening proceeds in calcareous dune limestones (AEOLIANITE) at the same time as karst development is occurring. This produces a style of landscape known as SYNGENETIC KARST. After further percolation the seepage waters may emerge into cave passages. Since cave air usually has CO_2 concentrations similar to the open atmosphere, the emergence of supersaturated percolation waters results in CO_2 degassing and in the precipitation of calcite in the form of stalactites, stalagmites and flowstones (collectively termed SPELEOTHEMS).

Whereas the most characteristic subterranean features of karst are caves, the most typical surface landforms are closed depressions, especially DOLINEs, which are enclosed bowl or saucer-shaped hollows, usually of a few hundred metres in diameter and some tens of metres deep. When dolines occupy all the available space, the surface has a relief like an egg-tray and is known as *polygonal karst*, but this does not always develop and often dolines are dispersed or in clusters across an undulating surface. Polygonal karsts can have doline densities ranging from 4–55 per square kilometre. The particularly large and correspondingly deep solution dolines of some tropical and subtropical karsts are also called *cockpits*, a Jamaican term.

Solution dolines (Plate 68) develop in the subcutaneous zone and drain water centripetally to enlarged fissures that discharge it vertically to the deep groundwater system (Figure 97). Small

Plate 68 The hallmark landform of karst: solution dolines near Waitomo, New Zealand

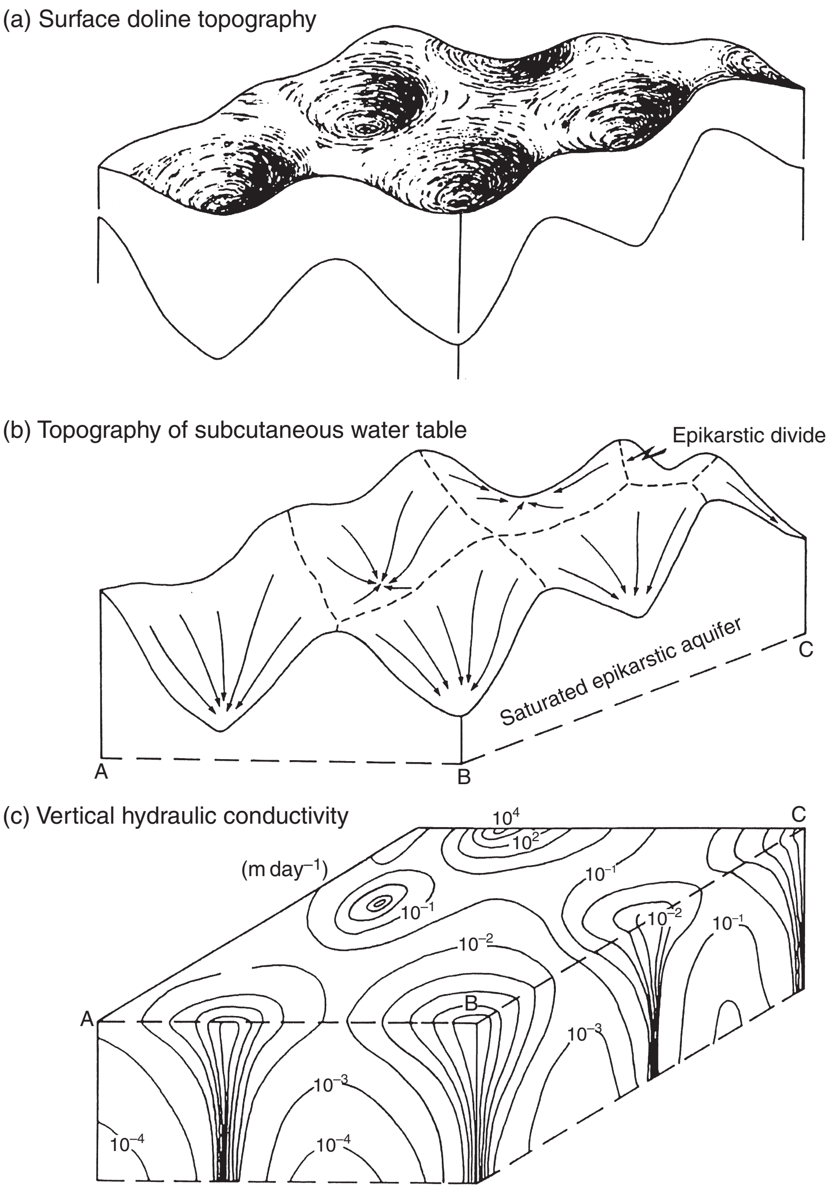

Figure 97 A model showing the relationship of solution doline development to flow paths in the epikarst (subcutaneous zone) and vertical hydraulic conductivity (from Williams 1985.)

allogenic streams also form enclosed basins, where they disappear underground. Large allogenic streams penetrate further into the karst in well-defined valleys before they sink, and produce landforms known as BLIND VALLEYS, because their valleys usually terminate abruptly in a cliff or steep slope. The sinking streams give rise to caves, and if a cave roof is close to the surface it sometimes collapses, producing a cylindrical or crater-like depression termed a *collapse doline*. A collapse that exposes an underground river is sometimes called a *karst window*. Some caves can be completely unroofed by progressive collapse, producing a gorge of cavern collapse (though not all gorges in karst are produced in this way, many being produced by antecedent drainage). Where doline collapse intersects the water table, the steep-sided enclosed depression holds a lake, such features being called CENOTES in the Yucatan Peninsula.

As dolines evolve, they often enlarge laterally and coalesce, producing compound closed depressions known as *uvalas*. Where the rate of vertical incision of dolines is significantly greater than the rate of solutional denudation of the intervening land, the inter-doline areas develop into hills. This is particularly common in humid tropical and subtropical karsts, where residual hills can be so well developed that they visually dominate the landscape, giving rise to a style of landscape called *cone karst*. In China, *fengcong* is the term used to describe such karsts (Plate 69).

Many karst areas have developed on rocks that have been folded and faulted. These tectonic influences considerably complicate karst evolution and are of major significance in guiding groundwater flow and denudation of the surface. Faulted terrains often provide the conditions in which the largest enclosed karst depressions – known as POLJEs – are developed, some exceeding 100 km^2 in area. The term 'polje' is of Slav origin and means 'field', probably because it was the largest area of flat tillable land in the karst. Many examples of these features are found in the Dinaric karst, where poljes are often located in faulted basins. Ford and Williams (1989) define three types: border polje, structural polje and baselevel polje – according to the dominant influences in their evolution (Figure 98).

Sometimes a particularly large blind valley encloses a basin of a square kilometre or more with a well-developed flat, floodplain floor, and it may receive more than one allogenic stream sinking at different points. Such large enclosed depressions at the edge of karsts are known as *border poljes*. Because of their relationship to sinking streams the floors of poljes often flood, particularly when the discharge of the inflowing river is greater than can be absorbed by the stream-sink(s). There is no clear demarcation between blind valleys and border poljes. They are transitional forms, the larger ones with particularly flood-prone flattish floors being called poljes.

Poljes may also be found in the interior of karsts, where structural dislocations have produced tectonic depressions with inliers of relatively impervious rocks. In these cases, the inlier acts as a dam on regional groundwater movement, forcing it to emerge as springs on the upstream side of the barrier. Water then flows across the impermeable inlier, to sink in ponors on the downstream side, the intervening region being developed into an alluviated plain. These features are known as *structural poljes*.

Genetically distinct from the above is the *base-level polje*, which is a very large enclosed depression entirely in karst rock that has been incised by solution down to the level of the *epiphreatic* zone (the zone of fluctuation of the water table). Such poljes are typically located close to the outflow boundary of a karst. They have swampy floors and can be envisaged as windows on the water table. Hence they inundate when the regional water table (or piezometric surface) rises in the wet season.

When vertical denudation eventually reduces the bottom of closed depressions to the level of the regional water table, they can incise no further, so instead they widen their floors. As a result residual hills between dolines become isolated, such

(a)

(b)

Plate 69 Two examples of humid subtropical karst in Guangxi province, China. The skylines are dominated by the conical forms of the hills, but enclosed depressions occur between them. When depression floors reach the water table they widen at their base, isolate the hills and extend the floodplain surface

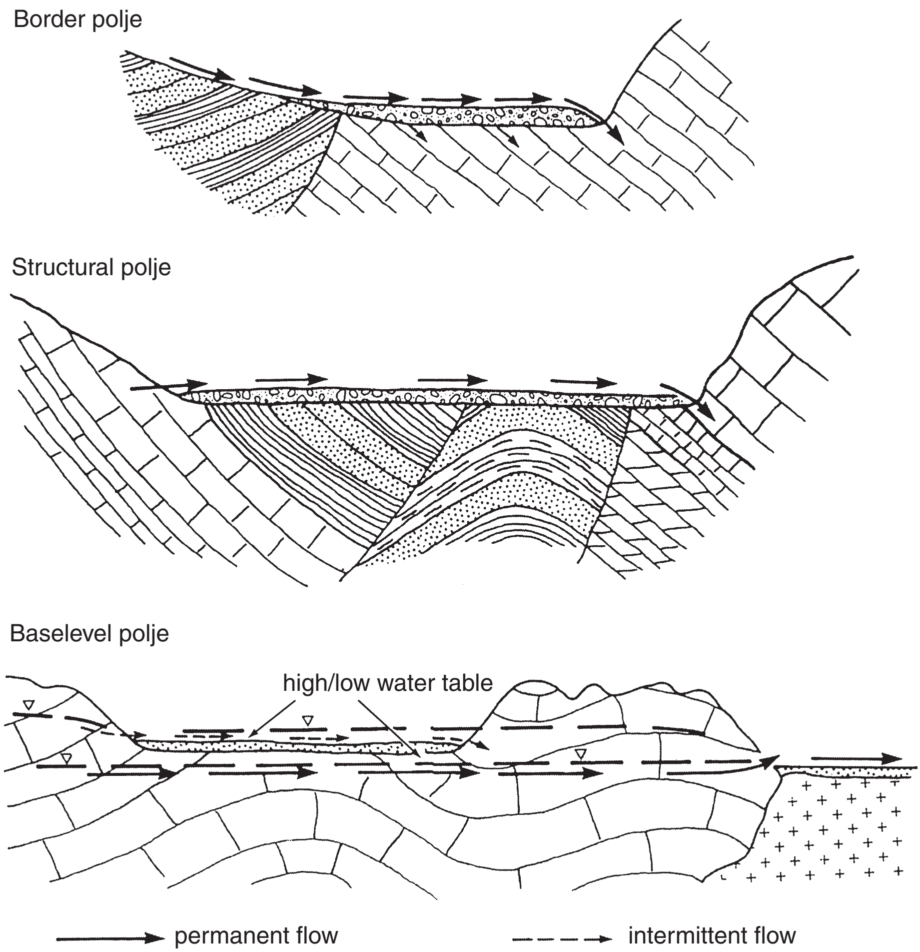

Figure 98 Three main types of polje. These are the largest enclosed depressions found in karst landscapes. Their flat floors are prone to flood and may cover many square kilometres in area (from Ford and Williams 1989.)

landforms being called *hums* in Europe. The lower slopes of such hills, which are usually of a rounded conical shape, can be over-steepened by undercutting and may collapse at their base, a process brought about by the corrosional attack of swamp waters – made particularly vigorous if allogenic rivers periodically flood the intervening plains. This is especially common in tropical humid karsts, where landscapes of steep isolated hills are referred to as *tower karst* (*Kegelkarst*, German), superb examples being known in southern China (where it is called *fenglin*). In the Caribbean isolated karst hills produced in this way are called *mogotes*.

During the end stages of karst denudation caves are drained and dismembered and their remnant passages are left at various elevations within residual karst hills, until eventually even the residual hills are removed by solution and only a corrosion plain is left. A superb example of a corrosion plain is the Gort lowland of counties Clare and Galway in western Ireland, where Pleistocene glaciations have stripped away the mantle of residual soil, alluvium and loose rock to reveal the karstified bedrock beneath.

Uplift can rejuvenate karst systems. But whereas in the first cycle of karstification the rock was unweathered and had only primary porosity, in the second cycle there is an inheritance of landforms on the surface and secondary porosity underground. Thus a new phase of karst evolution would exploit the inherited features and develop them further.

Modelling karst evolution

Various attempts have been made to model karst landscape development. Early conceptual models of karst evolution were presented by Grund (1914) and Cvijić (1918) (see translations into English in Sweeting 1981), but it was not until the late twentieth century that models became quantitative.

Ford and Ewers (1978) used a physical laboratory model to elucidate the development of protocaves and successive flow paths. White (1984) developed a theoretical expression showing the relationship between chemical and environmental factors in the solutional denudation of limestones and he also developed a model of the development of cave passages (White 1988).

Ahnert and Williams (1997) developed a 3-dimensional model of surface karst landform development that started with a terrain in which proto-conduit connections were already established and then showed how the relief might develop given different assumptions about starting conditions, such as randomly disposed sites of greater permeability or random variations in initial relief (Figure 99). This model illustrates sequential steps in the development of doline and polygonal karst and reveals the importance of divergent and convergent flow paths in explaining the development of residual cones between incising depressions.

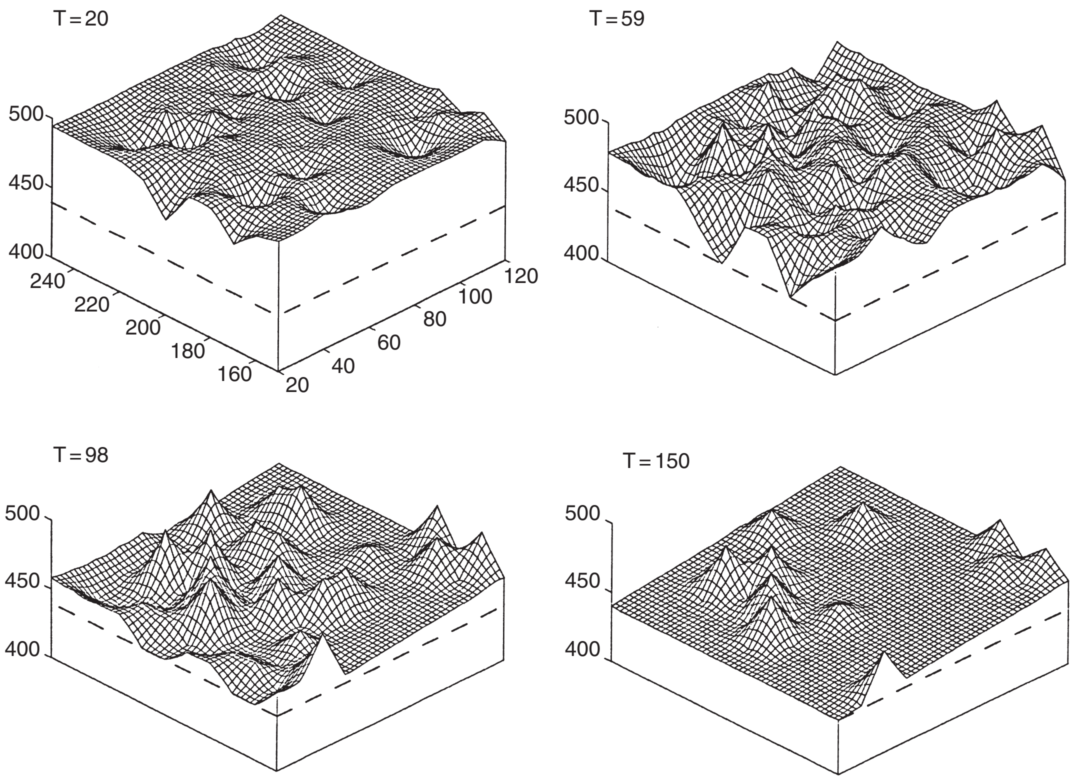

Figure 99 An illustration from one run of a three-dimensional theoretical model of karst landform development. The model shows an undulating surface with dolines at time 20 (T = 20), the development of polygonal karst by T = 59 (when some doline bottoms attain base level (shown by dashed line), the commencement of isolation of residual hills by T = 98, and the development of a corrosion plain with isolated hills by T = 150. The corrosion plain has a gradient towards the left because of the slope (hydraulic gradient) of the water-table (from Ahnert and Williams, 1997.)

Other more recent models associated with dissolution and the processes governing the evolution of karst are presented in Klimchouk *et al.* (2000) and Gabrovšek (2002).

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PAUL W. WILLIAMS

KETTLE AND KETTLE HOLE

Kettle holes are depressions formed by the melt of discrete blocks of glacier ice that have been partially or completely buried by GLACIFLUVIAL sediments. Kettle holes have been reported from many present-day proglacial environments. Ice blocks originate in three ways: (1) detachment from the glacier snout due to ablation (Rich 1943), (2) transport on to the outwash plain or 'sandur' by glacial streams or rivers (Maizels 1977) and (3) release and transport on to the sandur surface during jökulhlaups (glacier OUTBURST FLOODS) (Fay 2002). Kettled or pitted sandur is glacial outwash in which numerous kettle holes have formed.

Kettle holes may be inverse-conical or steep-walled in shape. Inverse-conical kettles form due to the melt of partially or totally buried ice blocks and develop by sediment slumping and avalanching down the kettle walls. Inverse-conical kettles formed by the melt of partially buried blocks may possess raised diamict (see DIAMICTITE) rims (rimmed kettles) and/or diamict mounds in the base of the depression. Steep-walled kettles form by collapse of overlying sediment into voids created by the *in situ* melt of completely buried ice blocks. Ice blocks transported on to the sandur by water are often progressively buried by glacifluvial sediments. However, during a jökulhlaup, sediment deposition can be so rapid that small ice blocks are incorporated within the flow and deposited simultaneously with flood sediments.

The physical properties of the sediment in which kettle holes form control their collapse sequence (Fay 2002). Shallow kettles with vertical or inward-dipping walls, whose base is a coherent block of sediment, form in coarse, clast-supported sediments. In coarse sediments dominated by matrix-support or in entirely fine-grained sediments, deeper kettles with steeply dipping or overhanging walls form, often through sudden roof collapse. Steep-walled kettle holes may form over small, buried blocks, or over larger, buried ice blocks which melt irregularly. Steep-walled kettles can develop into inverse-conical kettle holes by slide or avalanche of sediment into the kettle hollow.

Since the development of kettle holes is similar over both stagnant glacier ice and flood-related ice blocks, it may be difficult to differentiate between flood-related kettles and kettles produced by non-fluvially driven processes. However, on a palaeoflood surface, a flood origin for kettle holes is indicated by a distinct radial pattern of kettle holes on prominent outwash fans reflecting flow expansion, and/or a decrease in kettle size down sandur relating to a progressive decrease in stream power.

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HELEN FAY

KNICKPOINT

Knickpoint (nickpoint) refers to a substantially steepened section of a stream long profile. In cyclic interpretations of landscape (Davis 1899) knickpoints carried a new 'cycle' of erosion inland replacing an older, more elevated low relief surface and producing a gorge (see GORGE AND RAVINE). For knickpoint erosion in alluvial material, soil and rill systems, and in the laboratory, the term 'headcut' is used (Bennett 1999), or for extended steep reaches, knickzone (Downs and Simon 2001). The term WATERFALL is synonymous with knickpoint. The steepened valley walls below the knickpoint subsequently collapse by mechanical failure following stress release (Philbrick 1970), enhanced groundwater drainage, and by stream erosion undercutting the valley wall. Despite morphological similarities between knickpoints in cohesive clay materials and those in bedrock, they may not retreat by similar processes. Both cases depend upon the headwall being continually re-steepened with the bulk of any detrital material present at the base of the knickpoint being removed. In weak material incapable of maintaining a cap to the headwall the knickpoint face may rotate as it retreats: longitudinal slope diminishing until the knickpoint is removed.

Niagara Falls has a mean flow rate (before abstraction for power generation) of $5{,}730\,m^3\,s^{-1}$. It retreated at a rate $1.5\,m\,yr^{-1}$ in the last decades of the nineteenth century (Philbrick 1970); a figure consistent with the estimated rate for the entire postglacial period (14,000 calendar years). When flow was only 10 per cent of the present value (Tinkler *et al.* 1994) for 5,000 years, the recession rate was between 0.10 and $0.15\,m\,yr^{-1}$, and with a reduced elevation for the headwall. Although $1.5\,m\,yr^{-1}$ is high, the recession rate of the 12 m St Anthony Falls on the Missouri may have been comparable in the nineteenth century (Winchell 1878; Sardeson 1908). The inference for both waterfalls is that the headwall retreated in an essentially parallel fashion during periods of thousands of years.

Recession rates for other major knickpoints are known only for a few systems at present. In south-east eastern Australia (van der Beek *et al.* 2001) rates of the order of 1,000 m per million years have been estimated, and half the rate that has been reported in adjacent areas of Australia (Nott *et al.* 1996; Seidl *et al.* 1997). In southern Africa Derricourt (1976) has suggested rates for the recession of the Batoka Gorge, below the Victoria Falls, have varied between 0.09 and 0.15 $m\,yr^{-1}$ over the last million years or so. Stranded and stationary surface knickpoints have been reported in karst terrains (Fabel *et al.* 1996; Youping and Fusheng 1997).

The process of knickpoint recession is far from clear. There is no published evidence that for large waterfalls headwall undercutting takes place in the plunge pool. In non-cohesive sediments with low slope the entire headcut may be submerged but this has only rarely been described for large bedrock forms (Rashleigh 1935). More probably headwall recession above the plunge pool level proceeds by subaerial weathering in a very damp environment; by sapping, from water fed to the vertical face from the upper river bed (Krajewski and Liberty 1981); and by stress release close to rock face weakening joints and bedding plains prior to block release. In the upper caprock zone (if present) accelerating water approaching the waterfall edge eventually exerts enough force to tear blocks out of the undermined capping beds (Philbrick 1970). At full flow, on waterfall faces less than vertical, erosional wear by water and entrained sediment may effect rock removal (Bishop and Goldrick 1992; and examples in Rashleigh 1935). Knickpoints in alluvial sediments can progress much faster (Simon and Thomas 2002) cite rates of 0.7 to $12\,m\,yr^{-1}$.

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KEITH J. TINKLER

L

LAGOON, COASTAL

There are three main meanings to the term lagoon. The most usual is to describe a stretch of salt water separated from the sea by a low sandbank or coral reef. Another meaning is that for a small freshwater lake near a larger lake or river. The third usage is for an artificial pool used for the treatment of effluent or to accommodate an overspill from surface drains. This entry is soley concerned with the first of these meanings.

Coastal lagoons are mostly to some degree estuarine, are usually shallow, and have generally been partly or wholly sealed off from the sea by the deposition of spits or barriers, by localized tectonic subsidence, or by the growth of coral reefs. They range in size from over 10,000 km^2 to less than 1 ha. They occur on about 12 per cent of the length of the world's coastline (Bird 2000). They are best formed on transgressive coasts, particularly where the continental margin has a low gradient and sea-level rise is slow. They are ephemeral features and their depths and areas become reduced by sedimentation from inflowing rivers, as well as by accumulation of sediment washed in from the sea, wind-blown material, and chemical and organic deposits. Indeed, they can be classified on the basis of whether they are infilling or increasing in size (Nichols 1989). The infill of some lagoons, particularly those that are parallel to the shore, may involve the development of cuspate divisions that divide the lagoon into a series of segments. These divisions have been attributed *inter alia* to winds blowing along the length of the lagoon producing waves which build spits that isolate the lagoon into separate basins.

The entrances between lagoons and the sea vary in origin and form. Some are residual gaps that persisted between spits or barrier islands where the lagoon was never completely sealed off from the sea. Others are caused by breaching either by storm waves or by floods from on land. Their configuration is the outcome of a contest between the inflow and outflow of currents, which keeps entrances open, and the effects of onshore and longshore drift of sediment, which tends to seal them off. Lagoon entrances tend to be larger, more numerous and more persistent on barrier coastlines where relatively large tidal ranges generate strong currents.

A good review of the diversity of coastal lagoon morphologies and evolution is provided by Cooper (1994).

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A.S. GOUDIE

LAHAR

An Indonesian word originally used by Escher (1922) for a hot volcanic mudflow that had been generated by an eruption through a crater lake. The term specifically applied to the 1919 volcanic mudflows originating from the crater lake of Mt Kelut, on Java, which inundated over 130 km^2 of the surrounding lowlands, with the loss of over 5,000 lives. The word quickly gained acceptance as a general term for a mudflow on the flanks of a volcano, irrespective of its triggering mechanism.

With increased understanding of water-saturated flow processes, the term has evolved to include both mudflows and debris flows, and later hyperconcentrated flows, on the flanks of a volcano. This has led to the more generally accepted definition: a rapidly flowing mixture of rock debris and water (other than normal stream flow) from a volcano (Smith and Fritz 1989). This definition describes the flow but not the resultant deposit.

Lahars are denser than normal stream flow because of their high sediment loads, causing them to move faster due to greater gravity forces and damped internal energy losses. The highly concentrated slurry usually exhibits a yield strength, behaving like wet cement but also exhibiting features of cohesionless grain flow. At lowering velocities, high concentration lahars may quickly halt, 'freezing' the coarser fraction within its finer matrix as a single massive sedimentary unit. The matrix strength, buoyancy and grain-dispersive pressures help support the coarser particles (Iverson 1997). Thus, resultant deposits are poorly sorted and may show reverse or no grading because little or no time was available for settling to occur. Thin, fine-grained sole-layers are attributed to shear and cataclasis within the basal flow. Lahars may travel as successive flow surges, which in high sediment concentration flows tend to continually shunt and overtop previous deposits. Clay-poor lahars may exhibit a frontal cliff or snout and show coarse bouldery levees at their margins.

A feature of lahars is 'bulking', when on the steep slopes of a volcano the flow typically erodes loose sediment over which it is flowing to increase its volume several times. On lower gradient slopes the reverse may happen leading to progressive dilution of the flow and formation of a HYPERCONCENTRATED FLOW or a normal flood (Pierson 1998). A water wave ahead of the lahar has been observed in some instances, which may represent a solitary wave or soliton (Cronin *et al*. 2000) or may be caused by under-ramping of the dense flow behind (Manville *et al*. 2000).

Lahars, armoured with their coarse bouldery loads, are a devastating volcanic hazard, capable of killing large numbers of people and removing all structures in their path. In recorded history at least 64,000 persons have been killed by lahars (Neall 1996). Lahars may vary in size from small volume events (<0.1 million m^3) to large-scale collapses of volcanic edifices (>3,500 million m^3). Measured peak discharges of historical lahars range from 100 to >100,000 m^3 s^{-1}. Lahars are capable of travelling up to 150 km/h and may travel for hundreds of kilometres down valleys. This mobility is attributed to positive pore-fluid pressures, which greatly decrease internal friction within the flow (Hampton 1979). Hence the most hazardous lahar zones are adjoining river courses draining volcanoes, particularly those draining crater lakes.

Lahars may be a more common hazard at stratovolcanoes than the geological record attests. Of twenty lahars observed in the Whangaehu catchment at Mt Ruapehu, New Zealand, during 1995, only one resultant deposit is preserved. This highlights the preservation potential of only larger events.

Lahars may be generated by both eruptive and non-eruptive mechanisms. Lahar triggering eruptive mechanisms include phreatic or phreatomagmatic explosions (often from hydrothermally altered and structurally weakened edifices), displacement of waters by eruptions through a crater lake, pyroclastic flows admixing with water in rivers or lakes, and volcanic melting of snow and ice. Non-eruptive triggering mechanisms include collapse of the walls of crater lakes, and heavy rains falling on recently erupted materials.

In the most recent large-scale volcanic disaster of modern times more than 23,000 persons were killed by lahars from Nevado del Ruiz, Colombia, in 1985. Relatively small eruptions at the summit produced pyroclastic surges, which quickly gouged and melted 0.06 km^3 of snow and ice to form lahars peaking at 48,000 m^3 s^{-1} and totalling 40–60 million m^3.

Large and extensive prehistoric lahar deposits are reported from many stratovolcanoes. One of the first to be recognized was the Osceola Mudflow from Mt Rainier, Washington State, USA, which filled proximal valleys 85–200 m deep before spreading over 350 km^2 of the Puget Sound lowlands about 5,600 years ago.

Mitigative measures to reduce damage from lahars include adequate real-time warning systems (such as acoustic flow monitors), reducing the level of water in crater lakes by engineering methods (such as tunnels at Mt Kelut, Indonesia), dams to reduce gradient and encourage sediment deposition, reduction of lake levels in hydro dams to accommodate sudden inflows, construction of embankments to divert flow away from assets at most risk, or as at Mt Pinatubo, Philippines, continually elevating houses above each successive lahar deposit.

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VINCENT E. NEALL

LAKE

Lakes are defined as bodies of slow moving water surrounded by land. They represent approximately 2 per cent of the Earth's surface but contain only about 0.01 per cent of the world's water (Wetzel 2001). While they are temporary features of the landscape on a geological timescale, they can exist for very long periods and therefore strongly influence the human development of a region. They can also provide a record of the region's environmental history in their sedimentary record. The study of lakes is called limnology and limnologists characterize lakes in a number of ways, including their geologic origin, mixing behaviour and nutrient status. While these classifications appear to be distinctly disciplinary, the geology, physics and chemistry of lakes interrelate significantly and all act to regulate the biological dynamics in lakes.

The depressions, or basins in the Earth's surface, that collect water to become lakes can be formed by several geologic and geomorphic processes. Catastrophic geologic origins include tectonic activity and volcanism. The deepest of the Earth's lakes are caused by tectonic faulting, whereas the clearest of lakes are found in the craters of old volcanoes. The majority of the Earth's lakes (72 per cent) have been created by glaciation (Kalff 2002). High alpine cirque (see CIRQUE, GLACIAL) lakes, lowland KETTLES AND KETTLE HOLES and glacial ice-scour lakes are abundant in the regions once covered by ice. Other lakes are generated by the modification of drainage systems, or the impoundment of flowing waters by natural disasters such as landslides or human activity such as damming. Riverine or fluvial lakes, such as those developed on floodplains, deltas and blocked valleys, comprise 10 per cent of the world's lakes and are the dominant lake type in low latitudes (Kalff 2002). Chemical dissolution of rocks also generates basins that collect water. Wind and ocean shoreline processes create barriers which act to block fresh water while animals and meteorites cause terrestrial depressions creating specific lake types. Some lake basins are excavated by wind (see PAN). The fact that lakes are most often created by processes that have dominated that landscape results in lakes of similar origin being regionally grouped. This is why there seems to be a preponderance of Lake Districts around the world, where bodies of water that appear and behave similarly due to their shared origin are identified as a lake grouping (e.g. Cumbria Lakes, England; Finger Lakes, New York; Great Lakes, North America). Such groupings provide opportunities for regional-scale research and have facilitated the study of lake types and processes.

The shape of the lake basin, or its morphometry, is a function of the lake origin but over time it does change as sedimentation and shoreline infilling occurs as part of the natural succession process. Zonation in lakes is also a function of morphometry. The nearshore region of lakes is called the littoral zone while the open water deeper portion is called the pelagic zone. Littoral zones support rooted plants, some of which are visible above the water surface and are called emergent plants. The littoral extends to the depth of water able to support rooted plants and it represents that region to which light can penetrate, allowing photosynthesis of these primary producers. The littoral zone may be extensive in a shallow dish-shaped lake or it may represent only a small nearshore area if the slope is very steep (see Figure 100). Primary productivity (plant growth) in the littoral is usually greater than in the pelagic zone as it is more closely linked to the catchment from which it receives nutrients for plant growth.

Lakes are classified physically by their mixing behaviour which is a function of their thermal structure. Solar energy enters the lake at the water surface and is attenuated as it moves down through the water column. Heat is therefore transferred predominantly to the surface waters. Warm water lies above the colder, denser water creating a separation or stratification of water layers. The warmer surface waters are known as the epilimnion while the cooler bottom waters are called the hypolimnion. The point where the temperature transition is most extreme between these two layers is called the thermocline and that portion of the lake is termed the metalimnion (see Figure 100).

Water is an unusual liquid in that it is most dense at 4 °C, meaning that water both warmer and cooler than 4 °C will be more buoyant and rise to a layer above. Water in its solid form, ice, is less dense than water and thus floats at the lake surface. The unique density characteristics of water ensure that lakes do not freeze from the bottom up and that most temperate lakes have bottom waters with moderate under-ice temperatures (4 °C) to support life.

Depending on the temperature (and therefore density) differences between the epilimnion and hypolimnion the two parcels of water may be very resistant to mixing. During periods of strong stratification the two compartments of water do not interact, or mix, and therefore their exchange of materials is restricted. The movement of both soluble and low density particulate material entering a lake can be confined to surface waters if the density differences at the metalimnion are extreme. When denser particulate matter settles through the metalimnion and enters the hypolimnion it will be stored and/or decomposed, but until the stratification is reduced all these materials (e.g. nutrients, pollutants, particles) will remain in the bottom waters. Chemical changes associated with the decomposition of the particles, such as reduced oxygen levels, are now confined to the bottom waters as there is little exchange of water and chemicals across the thermocline. In well-stratified, productive lakes oxygen depletion or anoxia often occurs in the bottom waters due to the oxygen demand of the organic-rich sediments (see Figure 100). Anoxic conditions at the sediment–water interface also generate chemical alterations in the sediment, resulting in increased exchanges between sediments and overlying water.

When stratification breaks down, due for example to the cooling of surface waters with changing seasons, and the water column becomes isothermal (one temperature) there is little resistance to mixing. Wind energy at the water surface can mix the full water column which is called ‘turnover’.

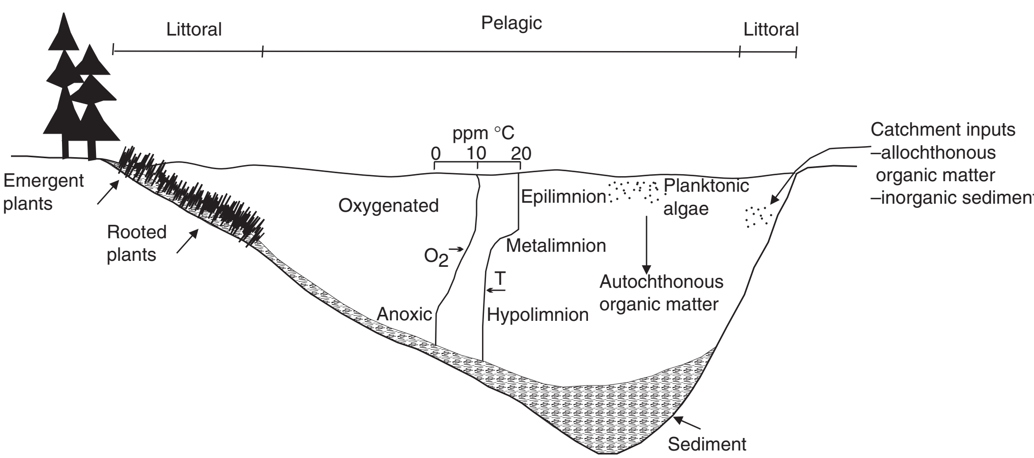

Figure 100 Limnological terms for the zonation, thermal structure and particulate sources to lakes. The general distribution of lake primary production and resultant oxygen profiles are also presented

This full lake mixing brings nutrients and chemicals from the hypolimnion to mix with the surface waters. Lakes in different climatic regions exhibit different annual patterns of stratification and so can be classified in this way. Lakes which mix twice a year are common in the temperate regions where cold winters give way to springtime heating generating isothermal conditions as the lake warms to 4 °C. Again in the autumn the lakes cool, causing destratification and an isothermal situation which allows autumnal winds to mix the bottom with the surface waters. This mixing, twice annually, is termed dimictic. Monomictic lakes mix only once per year and are found in high elevation and high latitude areas, while polymictic lakes, commonly located in equatorial regions, mix many times a year. Note that mixing is important in the transfer of both nutrients and pollutants in lakes. A less common but environmentally significant situation is a meromictic lake which has a bottom water layer that is chemically different from the rest of the water body. This generates a large enough density difference that even when the system is isothermal a chemical density gradient restricts this bottommost layer from mixing. This is often caused by salinity differences or groundwater inputs of differing chemistry. In this case the exchanges between the bottom sediments underlying the meromictic layer are restricted from delivering to the full water column.

Nutrient status is one variable used to chemically classify lakes. The terms oligotrophic, mesotrophic, eutrophic and hypereutrophic represent the spectrum of conditions from nutrient-poor through to nutrient-rich systems. In general, primary productivity increases across the spectrum with hypereutrophic lakes representing the stage where organic nutrient inputs (carbon, nitrogen, phosphorus) are high. Organic inputs to lakes can be derived from within the lake, termed autochthonous, or they can be delivered from the catchment or atmosphere, when they are called allochthonous. In-lake primary production of free-floating photosynthetic plants, or planktonic algae, and littoral rooted plants are the main sources of autochthonous organic matter. Catchment inputs from river inflows represent the main allochthonous contribution of organic matter. If the organic matter in lakes is not either eaten by an organism or decomposed during settling it will accumulate in the bottom sediments. Organic sedimentation rates increase over the trophic range as productivity increases.

The bottom sediments are also comprised of inorganics which have been eroded from the landscape and delivered via river inflows or shoreline erosion. As well, some autochthonous inorganics can be generated by chemical precipitation and biogenic processes. Sedimentation rates vary as a function of the location, size and activities in the catchment as well as the in-lake productivity, but in most natural temperate lakes ranges between 0.1 and 2 mm per year (Kalff 2002).

Globally much research has focused on the process and remediation of cultural eutrophication (Cooke *et al*. 1993). This accelerated change in tropic status occurs when unnaturally high phosphorus loads are received by the lake over relatively short time periods (decades). It is termed cultural eutrophication because the sources of the growth nutrient, phosphorus, are associated with human activities in the catchment, such as agriculture and sewage treatment. Increased primary productivity in both the littoral and pelagic zone, increased sedimentation rates, hypolimnetic oxygen depletion and the coarsening of fish species are often associated with anthropogenically induced alterations of tropic status. Remediation and management approaches for these problems are presented in Cooke *et al.* (1993).

As lakes are bodies of water that receive and store material from the surrounding catchments and the atmosphere they are of interest to geomorphologists as the accumulated sediment can reflect regional changes over time. Catchment erosion rates associated with changing land use, sediment source tracing, climatic variations, historic pollutant loads, flood records and vegetation patterns can be detected by evaluating different characteristics of the accumulated lake sediments. Sediment, collected by coring down through the accumulated material, can be horizontally sliced to differentiate sediments from specific time periods. Paleolimnology, or the use of lake sediments for reconstructing past events, requires some means of dating the accumulated material (see DATING METHODS) and a variety of methods exist (e.g. ^{210}Pb, ^{14}C, ^{137}Cs, thermoluminescence) but the precision and accuracy of each is restricted to specific time intervals. Given this, and the fact that lake sediments are temporally and spatially variable, it is very important to design the collection of cores and the analytical methods to suit the questions being addressed. Dearing and Foster (1993) provide a useful discussion of the problems, errors and implications of using sediment cores in geomorphic research. An earlier text by Hakanson and Jansson (1983) introduces the topic of lake sedimentology and provides information on physical, chemical and biological aspects of sediment.

Since 1970 the focus of limnological research has moved away from viewing the lake as a closed system upon which to do ecological research and more effort has been placed on linking catchment processes to lake conditions (Kalff 2002). Lakes are intimately connected to their catchments and therefore the role of lakes in geomorphological research and the role of geomorphologists in the interdisciplinary study of lakes are significant.

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ELLEN L. PETTICREW

LAND SYSTEM

Land systems are an integral part of the land evaluation process. The first land systems studies were devised and carried out by Australia's Commonwealth Scientific and Industrial Research Organisation (CSIRO) as part of their *Land Research Series* (Christian and Stewart 1968; Stewart 1968). A hierarchical classification of land was used. The smallest areal unit is the *land facet*, which is a relatively homogeneous area of land in terms of slope and soil; such as the midslope unit of a CATENA. Repetitive sequences of land facets form *land units*, which are the fundamental mapping unit of the land systems approach. Land units comprise a distinctive pattern of land facets (slopes, soils and vegetation), are essentially catenary in nature and classically were illustrated by block diagrams with explanatory annotations. *Land systems*, then, are larger areas comprising groups of land units representing recurring patterns of soils, topography and vegetation.

Work commenced in Australia in the late 1940s and continued to the 1970s, later refined by government agencies in some States. Similar surveys were carried out by the British Ministry of Overseas Development, particularly in Africa (Land Resources Development Centre 1966 onwards), and by the French scientific organization ORSTOM and the Netherlands International Institute for Aerial Survey and Earth Science (ITC) in many countries (Nossin 1977).

The objective was to provide rapid, cost-effective and objective assessments of land resources, initially for agriculture and forestry but more recently as frameworks for regional or national planning, in remote areas for which there was little if any base information. There were no maps depicting contours, soils or vegetation and little geological or climatic information for these areas. The primary data source was small-scale (1:80,000), monochrome aerial photography, flown primarily for military purposes from the 1940s. Thus the method of land systems surveys was essentially an air photo interpretation exercise, supported by ground truthing. Patterns of tone and texture were delimited on the aerial photographs. Offroad ground traverses were then designed to visit examples of each pattern on the ground in order to record the vegetation, soils, topography and geology. Since the tonal and textural patterns primarily reflect vegetation, plus areas of bare ground, rock outcrops and water bodies, vegetation associations were particularly important in characterizing land units and land systems.

However, several members of the early land systems teams were British-trained geomorphologists (including Mabbutt, Ollier, Twidale and Young). Not surprisingly, therefore, geomorphology played a large part in the written interpretations of the land systems. That geomorphology was heavily influenced by the then prevailing Davisian cycle of erosion which, as pointed out by Chorley (1965: 35–36) in a different context, has little if any bearing on the prediction of land suitability for agricultural or other land uses. This is also largely true of genetic interpretations of soils.

The first land system study in Australia was of the Katherine–Darwin region (Christian and Stewart 1953). As a result of the survey a relatively small area was identified as being possibly suitable for agriculture, and a research station was established to conduct detailed field investigations and experiments. A similar outcome was obtained from the East Kimberley survey. Both research stations are still operational, and in the East Kimberley the eventual outcome was the development of the Ord irrigation scheme. While the entire area of Papua New Guinea was eventually covered by land systems surveys, the spatial coverage of over twenty reports in the Land Research Series in Australia was more scattered. Debatably, the developments outlined above were the only non-educational, positive outcomes of the series of Land Research reports.

Weaknesses of the technique included: its reconnaissance-type of approach (which ironically was also its strength); its static coverage (processes were not included); its geomorphic base with a strong, denudation chronology flavour; its exclusive focus on the biophysical environment; and the crudity of the data. Nevertheless, land systems may provide a useful basis for later and more dynamically oriented work, especially when there is little background information on the environment or on ecological conditions. In Victoria, for example, geomorphic elements have been refined as part of a land systems review to provide a suitable framework for describing the spatial attributes of land (Rowan 1990).

Only in the Hunter Valley of New South Wales was a CSIRO land systems survey conducted in an area already well-developed for agriculture. The Hunter Valley survey covered a smaller area than previous surveys and that, plus the larger amount of information which was available for the region compared with the more remote regions, made it possible to provide greater detail (Story *et al*. 1963). But that survey also exposed the limitations of focusing on the biophysical environment. In most areas, planning for future land uses does not take place in unoccupied country but in places which are already occupied and the land already subject to land-use practices. Existing uses, therefore, must be incorporated in any survey designed to assist planning for future land uses.

The South Coast study

Realization of this self-evident truth led to the eventual abandonment by CSIRO of its limited land systems surveys (although the basic method is still used widely in various contexts), and the development of two separate but related lines of further work. One was a focus on databases, which eventually merged with the currently thriving area of research into and applications of Geographical Information Systems in land use planning; and the other was the *South Coast study* (also in New South Wales) (Basinski 1978). This was a very ‘geographical’ survey, in the sense that its approach would be familiar to any regional geographer. Not surprisingly, geographers were prominent amongst the team members – as physical geographers had been in the earlier land systems survey teams.

The South Coast study, a large, multidisciplinary project, conducted the familiar, integrated, land systems survey of the biophysical environment of the 6,000 km^2 region. Importantly, however, present land use, population demographics and socio-economic aspects such as settlements, transport networks, and the economy and social structure of the region were also included. The study explicitly set out to provide a ‘rational basis for planning decisions on a wide variety of land uses’, as well as investigating methods of providing and analysing biophysical and socio-economic data for the region (Basinski 1978). This survey was a precursor of ‘land suitability’ surveys and regional planning and, more particularly, of CSIRO’s Siroplan approach to land suitability evaluation for integrated, regional land use planning (Cocks *et al*. 1983).

Recent approaches to land evaluation

The term ‘land systems’ continues to be used and modified in some jurisdictions, particularly by agricultural agencies (at least in Australia). But it is the mapping unit of the Land Research Series, the *land unit*, which provides the basis of several approaches; notably FAO’s land suitability evaluation (FAO 1976) and land use planning frameworks (FAO 1993), as well as regional planning, at least in Western Australia, in the form of the basic planning and land management unit (example: WAPC 1996).

Improved technology and data availability have made possible considerable improvements in land evaluation methods and the quality of the output, and Davidson (2002) has provided a useful review of recent developments in the assessment of land resources. The availability of high resolution, satellite-borne remote sensing imagery, the rapid development of geographic information systems (or science) (GIS), and geostatistics (spatial data analysis, modelling and fuzzy set algebra) have meant that current methods of land evaluation would be unrecognizable to the early practitioners. Nevertheless, there is still a need to map areas of land ('land units') and to obtain field-derived data. The quality of existing, mapped soil data (and other land attributes) in many developing countries necessitates field-based surveys. Unfortunately there is a widespread tendency amongst GIS practitioners to consider that computer manipulation and interpolation of existing data is all that is required. There is a renewed need to educate natural land resource assessors to obtain data which are relevant to the problem in hand and not to 'make do' with – often inappropriate or inaccurate but readily available – existing data.

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ARTHUR CONACHER

LANDSCAPE SENSITIVITY

The sensitivity of a landscape to change is the likelihood that a given change in the controls of a system will produce a sensible, recognizable and sustained response. The sensitivity is a function of the propensity for change which is measured by the size of the impulse required to initiate change.

The necessary impulse is determined by the number, type and magnitude of the barriers to change. The barriers to change are complex and include the resistance of the rocks, their structure and their resilience (strength resistance); the slope, relative relief and elevation that determine the potential and mobilized energy (morphological resistance); the distance to BASE LEVEL, sources or energy or other barriers (location resistance); the ability of the system to transmit energy, waste and water, the density of pathways, stream density and joint frequency (transmission resistance); and the strength of the linkages between components of the system and the degree of hillslope-channel coupling (structural resistance).

A system also has a varying capacity to absorb the impulses of change. There are shock absorbers, filters, void spaces and energy drains. For example a BEACH disposes of wave energy by moving particles and doing work (displacement, friction, heat) and by drainage.

Sensitivity is also determined by the inheritance or degree of influence of previous system states, process systems and landforms. Of particular interest are systems that have experienced a very efficient sediment flux regime but which after environmental change or re-specification of the controls finds itself too flat or exhausted to permit further vigorous change. Large events may have a dramatic effect on a hillslope and remove all available (weathered) material. The barrier to further change is then the time (efficiency of preparation processes) needed to 'ripen' the system.

There are two limiting types of system. Mobile, fast-responding landforms have a high sensitivity, react quickly and relax to new states with facility. If they have high 'permeability' they may act as an energy filter so that change is 'skin deep' and fluctuates about an average form (e.g. a beach profile). They may also store small impulses and accumulate them into a large impulse that may cross an internal threshold value. Such systems are often morphologically complex and the correlative deposits are fragmentarily preserved. They are usually capable of rapid restoration.

Slow-responding systems may be insensitive because they are too flat, too far from boundary (energy) changes, propagate events slowly, have large storage or low concentrations of sediment transport or progressive weakening axes. Change does not take place easily so that, once their form is established, they may require large or effective events to achieve adjustment. However, if change does take place the results may be dramatic. For example, gully incision into a plain, or landsliding following progressive softening. Generally there is a persistence of relief and pattern, stagnancy of development, a palimpsest of forms and 'traditional' development in which the inherited landforms continue to develop as before even though the controlling environment may have changed.

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DENYS BRUNSDEN

LANDSLIDE

Landslides belong to a group of geomorphological processes referred to as MASS MOVEMENT. Mass movement involves the outward or downward movement of a mass of slope forming material, under the influence of gravity. Although water and ice may influence this process, these substances do not act as primary transportational agents. Landslides are discrete mass movement features and are distinguishable from other forms of mass movement by the presence of distinct boundaries and rates of movement perceptibly higher than any movement experienced on the adjoining slopes. Thus this group of processes includes falls, topples, slides, lateral spreads, flow and complex movements as classified by Cruden and Varnes (1996). Widespread diffuse forms of mass movement such as creep, subsidence, rebound and sagging are generally not treated as landslides.

The criteria used to distinguish different types of landslide generally include: movement mechanism (e.g. slide, flow), nature of the slope material involved (rock, debris, earth), form of the surface of rupture (curved or planar), degree of disruption of the displaced mass, and rate of movement (see MASS MOVEMENT).

Form and behaviour

There are several morphological features that can be recognized to a greater or lesser extent in most landslides (Figure 101). The uppermost part is the depletion or concave zone (erosional, generating or failure zone) where slope material has failed and become displaced downslope. In some cases the displacement may be only a few metres while in others the failure zone will be completely evacuated to expose the surface of rupture and to leave a distinctive scar on the hillslope (Plate 70). The displaced mass may remain close to the failure zone or it may continue to travel downslope leaving a transport track ending in a colluvial accumulation zone or acting as a supply to some other geomorphic agent (e.g. river, sea or glacier). The distance that landslide material travels (runout) is a characteristic of the type of landslide. For example, controlled by the height of fall and volume, rock avalanches can travel at high velocity for several kilometres. Runout distance and velocity for other types of landslide are controlled by factors such as volume, slope angle and morphology, clay content, water content, and surface frictional characteristics of the runout pathway.

Landslide movement may be instantaneous, with failure, transport and deposition taking place in a matter of seconds or minutes. Other landslides are known to have been intermittently active over tens of thousands of years, undergoing successive periods of reactivation. Eight states of

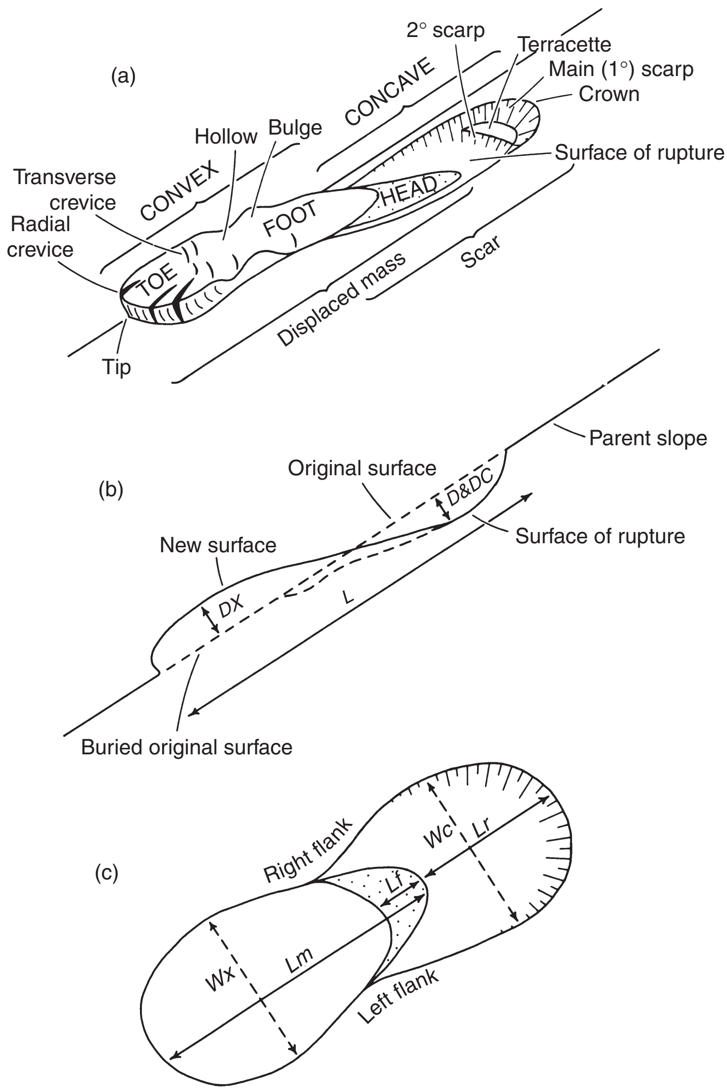

Figure 101 Morphological components of a landslide, indicating depth (*D*), length (*L*), and width (*W*) measurements

Plate 70 Hope rock slide (British Columbia), triggered by an earthquake in 1965 showing the depletion zone almost entirely evacuated of sediment

activity are recognized (Cruden and Varnes 1996) including such categories as ‘active’ (currently moving) ‘dormant’ (no movement in the last twelve months but capable of being reactivated) and ‘relict’ (unlikely to be reactivated under present climatic or geomorphologial conditions).

Rates of movement for different types of landslide are highly variable. Some landslides record only a few centimetres of movement a year, sustaining this rate for decades. Certain debris flows have recorded velocities of 100 $\mathrm{km h^{-1}}$ while large rock avalanches are capable of reaching velocities of 350 $\mathrm{km h^{-1}}$. Landslide deposits range in texture from dislodged blocks of intact source material to highly comminuted sediments forming a poorly sorted, unstratified diamictite.

Significance

The potential impact of landslides depends not only on velocity but also on volume. One of the largest landslides described is Green Lake Landslide, Fiordland, New Zealand which is estimated to be 13 $\mathrm{km^3}$ in volume. Submarine landslides of a similar magnitude are evident on edges of the continental shelf while recorded events such as the 1989 Ok Tedi mine landslide, Papua New Guinea and the 1970 Huascaran rock avalanche, were both estimated to exceed 50 million $\mathrm{m^3}$ in volume. Large volume, high velocity movements can also create substantial LANDSLIDE DAMs impacting fluvial systems and often posing a significant dam burst hazard.

Landslides are a manifestation of slope instability (see SLOPE STABILITY) and occur when changing conditions on a part of the slope allow shear stress to exceed shear strength. This can be brought about by a decrease in the effectiveness of factors that promote strength (e.g. a reduction in friction caused by increased pore-water pressures) or by an increase in shear stress (e.g. slope steepening). However, when a landslide takes place it changes the relative stability conditions from an unstable to a stable state, by reducing slope angle, height or weight, or by removing susceptible material. If boundary conditions are stable over a long period of geomorphic time, continued landsliding in a particular region may so alter slope conditions that other slower acting processes such as soil creep become dominant.

As most slopes are stable most of the time, landslides when they occur can be seen as a rapid and effective geomorphic response to the appearance of destabilizing changes in boundary conditions, enabling rapid adjustment and an eventual return (or tendency) toward more persistent landscape forms.

Destabilizing conditions (preparatory factors) in natural systems may be instigated by disturbances such as tectonic uplift, oversteepening of slopes by erosional activity, climatic change, deforestation or slope disturbance by human activity. The effectiveness of preparatory factors in reducing stability depends on preconditions such as material properties and slope geometry. The degree of stability afforded by preconditions and preparatory factors defines ‘landslide susceptibility’. The occurrence of landslides in space can be related to susceptibility thresholds (e.g. minimum critical slope angle) while occurrence in time can be related to exceeding magnitude thresholds for triggering agents (e.g. rainfall intensity, or seismic shaking).

Landslides represent a significant hazard to life, livelihood, property, infrastructure and resources in many parts of the world. Some individual catastrophic failures have been associated with high death tolls; the 1970 Huascaran rock avalanche in Peru killed 18,000 and deaths in the 1920

Kansu landslide in China are estimated at between 100,000 and 200,000. However, much landslide damage is less obvious, seriously depleting soil resources, reducing primary productivity and destroying property by slow chronic movement.

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LANDSLIDE DAM

Landslide dams are naturally occurring stream blockages caused by hillslope-derived MASS MOVEMENT. They represent end members in the spectrum of landforms of geomorphic HILLSLOPE-CHANNEL COUPLING, and arise from a temporary or permanent transport-limitation of the fluvial system due to excess lateral sediment delivery. The impounded natural reservoir is referred to as a *landslide-dammed lake*, whereas the term *landslide pond* relates to small water bodies perched on top of LANDSLIDE deposits, which need not necessarily be associated with stream blockage.

Landslide dams commonly form in steep terrain of many upland areas throughout the world, and constitute some of the highest natural dams on Earth. Yet the majority of occurrences are short-lived: 85 per cent of 185 worldwide examples have failed within less than one year, and nearly half within ten days (Costa and Schuster 1988). Typical failure mechanisms comprise overtopping by either lake-level rise or landslide-induced displacement waves, breaching, piping, gravity collapse of the dam face, or artificial spillway control. Rainfall, snowmelt and earthquakes are amongst the main triggers of landslides causing temporary or permanent blockage of rivers. Other mechanisms include volcanic eruptions, fluvial undercutting or, in some instances, anthropogenic activity. Stream blockages formed by LAHARS or pyroclastic flows (see PYROCLASTIC FLOW DEPOSIT) may be regarded as phenomena on a continuum between landslide and volcanic dams.

Landslide dams constitute a variety of GEOMORPHOLOGICAL HAZARDS, which may extend for considerable distances up- and downstream of the initial point of blockage. Long-term impoundment of river channels may cause extensive backwater flooding and associated SEDIMENTATION. Substantial physical impact may be created by sudden dam failures leading to catastrophic OUTBURST FLOODS of landslide-dammed lakes, which can turn into DEBRIS FLOWS, given sufficient amount of valley floor deposits to allow for alluvial bulking. Downstream reaches usually experience massive AGGRADATION, lateral channel instability and AVULSIONS, in the wake of landslide-dam failures. Rapid drawdown of the draining lake reservoir may cause further secondary loss of SLOPE STABILITY.

Landslide dams often create profound geomorphic legacies in the context of long-term landscape evolution of VALLEY floors. These include partly BURIED VALLEYS, drainage reversal, large intramontane alluvial flats resulting from the infill of landslide-dammed lakes, LAKE terraces, spillway gorges (see GORGE AND RAVINE) and RAPIDS, as well as conspicuous step-wise disruption of river long profiles (see LONG PROFILE, RIVER).

The 60-km long Lake Sarez, Tajikistan, is recognized to have been impounded by the world's highest existing landslide dam (~700 m), which has been formed by an earthquake-triggered rock avalanche of ~2×10^9 m^3 in volume near the former village of Usoi, in 1911 (Alford *et al*. 2000).

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SEE ALSO: dam

OLIVER KORUP

LARGE WOODY DEBRIS

Large Woody Debris (LWD) refers to fallen trees and detached wood within streams or along stream banks. The term 'large' applies to wood that is big enough to affect stream hydraulics, sediment transport, bank erosion and the resultant stream morphology, or is big enough to provide cover and habitat for aquatic organisms. In practice, LWD often refers to wood that is at least 1–2 m in length and a minimum of 10–30 cm in diameter, although the definition varies among geomorphologists and with stream size. Living trees along stream banks and vegetation outside stream channels are not referred to as LWD, although these types of wood also affect geomorphic processes (see EROSION; FOREST GEOMORPHOLOGY; RIPARIAN GEOMORPHOLOGY).

LWD is widely recognized as a critical component of aquatic ecosystems (Maser *et al*. 1988), although as recently as the 1970s management agencies removed wood from streams to reduce flooding and promote fish passage. LWD provides organic material, traps sediments, creates pools and, in general, increases the habitat diversity of streams. Resource managers now regularly emplace LWD in rivers to promote natural stream processes and recovery.

LWD enters the stream channel as logs, branches and root balls derived from bank failures, windthrow, landslides, debris flows, natural death and breakage, and human disturbances such as logging. LWD in small headwater streams often bridges the channel, only affecting morphology where breakage and decay create local log steps that promote sediment accumulation and channel widening (Nakamura and Swanson 1993). As stream size increases, the LWD falls into the stream rather than spanning it. These intermediate size streams cannot easily transport the wood, which is trapped and accumulates as single pieces or in piles known as 'jams'. The trapped wood forms log steps and pools and increases sediment storage, although channel widening can occur where logs deflect flow against a bank. In larger order streams, channel dimensions significantly exceed the size of LWD, and the wood can be transported by the flow. Larger streams sweep up the LWD as single pieces or jams along banks, on bars, and in backwater eddies (Marcus *et al*. 2002). When deposited along banks, wood in these larger systems can reduce bank erosion and promote pool development within the channel. When deposited in mid-channel, however, jams in larger streams can deflect the current and promote development of secondary channels that widen the overall stream width and increase bank erosion.

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SEE ALSO: debris flow; hillslope-channel coupling; landslide

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LAVA LANDFORM

Volcanic eruptions produce two main kinds of material – lava and pyroclasts (see VOLCANO). While most eruptive episodes will yield quantities of each, they can, in general, be described as predominantly effusive (producing lavas) or explosive (pyroclastic). The focus here is on the effusive variety and on extrusive rather than intrusive landforms. Lava is always partially molten on eruption but can contain a considerable volume fraction of both crystalline phases (minerals) and gas bubbles (vesicles). First-order controls on the eventual geomorphology of lava flows are provided by eruption rate and total amount of lava emitted, viscosity, the topography over which the lava flows, and the external environment (i.e. atmosphere, water or ice). Viscosity, the ratio of shear stress to strain rate, is a measure of the internal resistance to flow when a force is applied to a fluid. Magma viscosities are highly variable because there are many controlling factors, including temperature and composition of the melt. The presence of crystals increases viscosity. Bubbles, on the other hand, may increase or decrease viscosity depending on their size, properties and flow rate of the magma. Dissolved water also plays a role because it interrupts silicon bonding in the melt, hence reducing viscosity. In short, magma viscosities can vary by many orders of magnitude due to cooling, crystallization, vesiculation or loss of gas, and thereby strongly influence the nature of lava landforms.

Lava lakes and flows

Low silica, mafic lavas such as those erupted at Kīlauea (Hawai'i) have typical viscosities on eruption of only 10^2–10^3 Pa s, not much more than that of mayonnaise. When low viscosity lava erupts within a crater, it will tend to be confined within the vent region and crater, forming a lava pond or lake. Active lava lakes are connected to a deeper reservoir of magma, while passive lava lakes are not. Long-lived lava lakes are a comparatively rare phenomenon on Earth. They have been reported at several volcanoes including Kīlauea, Nyiragongo (Congo), Erebus (Antarctica) and Erta 'Ale (Ethiopia) but very few individual examples have reportedly persisted for more than a few decades. These include Halema'uma'u (Hawai'i, 1823–1924) and Erta 'Ale (probably active for at least the past century; Plate 71). Lava lakes have also been observed at oceanic ridges (Fouquet *et al.* 1995), and on the Jovian moon, Io (McEwen *et al.* 2000).

While all the manifestations discussed here are flows of lava and hence lava flows, the term 'lava flow' usually refers to erupting lava that has the opportunity to run down the flanks of a volcano or cross open ground. The expression is used

Plate 71 Erta 'Ale lava lake in the Danakil Depression (Ethiopia), is the longest-lived active lava lake on Earth. Its dimensions are approximately 80 × 90 m, and its power output is around 100–200 MW (Oppenheimer and Yirgu 2002)

both for active flows during emplacement, and for the resulting landform. In the course of a long-lived eruption, a lava flow field may develop by the superposition of many individual flow units. The current eruption of Kīlauea has yielded over 2 km^3 of lava and built up a flow field around 100 km^2 in area since it began in 1983 (a mean eruption intensity of around 3 m^3 s^{-1}, Plate 72). Areas of higher relief within a flow field that are above the 'high-tide' mark of the erupting lavas are called kīpuka. These 'islands' may preserve mature vegetation destroyed elsewhere by the active flows. Where lava enters the sea, as at Hawai'i, a lava delta or bench may build seawards, though these are often unstable features. As active lava interacts with seawater on the shore in front of the lava bench, steam explosions can construct ephemeral littoral cones.

Gas-rich, low-viscosity magma can erupt in quite spectacular fashion, with lava fountains ('fire fountains') playing to heights of up to several hundred metres. These can occur at individual vents or along fissures, which may become delineated by spatter cones or ramparts. Intense lava fountains can feed clastogenic lava flows when the spatter expelled loses little heat during its transit through the air.

The largest lava eruption of the past millennium was that of Laki (Iceland) in 1783/4.

Plate 72 The current eruption of Kīlauea (Hawai'i) has been in progress since 1983 and has formed a lava flow field some 100 km^2 in area. Most of the lava seen here is of the pāhoehoe variety, though the dark patch on the slope in the background consists of 'a'ā

It emplaced an estimated 14.7 km^3 of mafic lava, 40 per cent of which was erupted in the first twelve days at peak rates exceeding 5,000 m^3 s^{-1} (Thordarson and Self 1993). Bigger still, however, are the flood basalt eruptions or 'large igneous provinces' that punctuate the geological record. These have occurred both on land and in the oceans, and can contain 10^5–10^7 km^3 or more of lava, covering areas of 10^6 km^2, and erupted over timescales of perhaps 1 Myr, e.g. the 65-Myr-old Deccan Traps of India. The resulting landforms are known as lava plateaux. Where they are dissected by erosion, a staircase topography results from differential weathering of the rubbly boundaries and massive inner core of each superposed flow unit. This is the origin of the term 'trap', after an old Swedish word for staircase. Cooling and contraction of the cores of lava bodies can result in columnar jointing, seen spectacularly at the Giant's Causeway (Northern Ireland) and Devil's Postpile (USA).

Active lava flows radiate prodigious quantities of heat near the vent such that they rapidly form a surface crust. This can thicken sufficiently to insulate the core of the flow from thermal losses, lowering the rate of viscosity increase, and thereby promoting longer travel distance. Mafic flows quite often crust over completely, with lava continuing to flow in tunnels, which can grow in cross section by thermal erosion of the walls. When the supply of lava at the vent ceases, the last slug of lava may drain downslope leaving an empty conduit or lava tube. On Kīlauea, much of the lava flow between the Pu'u 'O'o vent, and the coastline where lava pours into the sea (a distance of over 10 km), takes place in a tunnel network, with only sporadic breakouts at the surface.

As silica content and crystallinity of lava increases, and eruption temperature drops, viscosities climb many orders of magnitude. As a result, much thicker accumulations of lava are required to overcome the resistance to flow.

Lava domes and coulées

The most silicic lavas attain viscosities of up to 10^8–10^{10} Pa s. On eruption, viscous flow is strongly resisted and slip may become concentrated along shear planes. Mounds of rock that accumulate around the vent, are called lava domes, and if they show some flow away from the vent, they are termed coulées. Sometimes, lava domes are intruded just beneath the surface causing bulges known as cryptodomes. In the weeks and days before its major 1980 eruption, a cryptodome of around 100 m in height developed on the upper flanks of Mount St Helens (USA). The eruption of Soufrière Hills Volcano (Montserrat), which began in 1995, produced an intermediate composition dome exceeding 10^8 m^3 in volume (Druitt and Kokelaar 2002). Such domes are really composite features, the construct of many individual lobes extruded at the surface of the dome (exogenous growth) and also intruded within it (endogenous growth). The frequent gravitational collapses and explosions that are typical of many dome-building eruptions can represent a severe hazard at volcanoes in populated areas. Frequent block-and-ash flows generated as the result of failure of hot sections of the dome of Merapi volcano (Indonesia) have claimed many lives.

Lava flow surface textures

The spectrum of surface textures of lava flows is impressive, and identical lava compositions can display very different textures due to subtle variations in effusion rate or cooling history. 'A'ā lava is characterized by loose clinker-like rubble, often overlying a more massive core. 'A'ā flows often build longitudinal levees by accretion along their margins. These are sometimes breached by new slugs of lava moving down the channel, and some civil protection efforts have attempted to divert flows by purposefully excavating flow levées. A curious feature of ʻaʻā flows are accretionary lava balls. They form in snowball fashion as a solid core rolls along the surface of a lava flow, picking up sticky molten rock. They can reach several metres in diameter, and gain sufficient momentum to roll some distance ahead of the lava flow front.

In contrast to ʻaʻā, pāhoehoe lava has a comparatively smooth surface composed of interlocking lobes, often adorned with a millimetre–centimetre-scale texture of interwoven threads. It can appear like coils of cord in ropy lava, like intestines in entrail lava, platy in slab lava, or blistered in shelly lava. Toothpaste lava forms by squeezing through cracks in the solid crust of a flow. Pāhoehoe can transform into ʻaʻā lava as it moves downslope due to changes in viscosity or strain rate but the reverse transformation is never observed. Investigations of the shapes of lava flow margins have revealed further insights into pāhoehoe and ʻaʻā flows. The former can be described by a scale-independent, fractal dimension, suggesting that pāhoehoe spreading is controlled by the ratio of the finely balanced forces driving advance and of those imposed by formation of the crust (Bruno *et al.* 1992). In contrast, the spread of ʻaʻā flows is dominated by the driving forces, and, as a result, the shape of their margins is not fractal. These observations have potential application in interpretation of volcanic terrains on other planetary bodies.

More viscous flows often develop a blocky surface texture, consisting of fractured chunks of lava, up to several metres across, with angular facets. Giant pressure ridges called ogives sometimes wrinkle the surface of very viscous flows. Lava domes often develop spines when they are active but these are usually rather ephemeral. After the devastating 1902 eruption of Mont Pelée (Martinique), an exceptional 300-m high spine was extruded.

Smaller features surrounding openings on mafic lava flows include hornitos (literally ‘little ovens’), which are chimneys or pinnacles of lava spatter and dribbles squeezed up through openings in the roof of lava tubes. Pāhoehoe flows on low slopes often display elliptical, domed structures known as tumuli. These form when the magma pressure within an active lava tunnel ruptures the overlying surface of the flow field. Fractures usually extend along the length of a tumulus, and often lava squeezes out through these and other cracks on the sides to build a larger feature.

Lavas erupted subaqueously resemble toothpaste squeezed out of a tube. The bulbous lobes that form are known as pillow lavas. These are commonplace along oceanic ridges but can also form in shallow water found in coastal, river and lake environments.

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SEE ALSO: volcano

CLIVE OPPENHEIMER

LAWS, GEOMORPHOLOGICAL

The whole concept of a scientific ‘law’ is far from straightforward and may be seen to become more and more problematic the further we get from the basic physical principles which can be translated into rational equations. Harvey (1969) concluded that all laws are statements of some universality, embedded within a theoretical structure. Haines-Young and Petch (1986: 13) consider that laws ‘describe some characteristic or behaviour of all the members of a class of things’ but that that something ‘would not usually be employed in recognizing it’. They go on (pp. 14–17) to give two, contrasting examples of scientific laws: that described by Coriolis force, governing the deflection of moving bodies to right or left in the northern and southern hemispheres; and Playfair’s ‘Law of accordant junctions’, which states that river valleys meet ‘on neither too high or too low a level’ (1802: 102). Whereas the first statement is a purely physical construct capable of mathematical formulation, the second is more problematic in that it is not evident how far it was intended to apply to the confluences of river channels or to valley floors (Haines-Young and Petch 1986: 16–17). Nor did Playfair formulate his observation as a law.

Since geomorphology is best seen as a historical science, it has provided many examples of both the ‘qualitative’ laws such as Playfair’s and the ‘physical’ laws, such as that of Coriolis force. Although A.N. Strahler in his very influential 1952 paper called for geomorphologists to strive for ‘the deduction of general mathematical MODELS to serve as quantitative natural laws’ (p. 937), there has been rather limited success in that direction, since not all the phenomena of interest to geomorphology have proved equally amenable to mathematical formulation. This is recognized by Rhoads and Thorn (1996: 123) who conclude that:

> [T]he study of complex phenomena poses a problem both for inductively establishing the existence of underlying causal laws by combining mathematical formulations of these laws in predictive models. This situation may account for the fact that geomorphology has not been very successful at developing its own laws, or in using simple models that combine a few basic physical laws to predict the form and dynamics of specific landforms.

They are clearly following Strahler’s exhortation in their evaluation of success or failure. Yet this is to ignore some of the most fundamental principles (or laws) which have governed the historical science of the study of landforms since the eighteenth century. We can make a crude distinction between the qualitative expressions of universality made by – most notably – James Hutton, John Playfair, G.K. Gilbert, W.M. Davis and W. Penck; and the quantitative formulations of R.E. Horton, S.A. Schumm, M.A. Melton, J.T. Hack and R.L. Shreve; and J.W. Glen.

By far the most basic proposition accepted by geomorphologists is that first recognized by Hutton in 1785, namely, that there is no need for recourse to extraordinary processes to account for the fashioning of the Earth’s surface, providing we accept an almost unbelievably long lapse of time. Hutton’s ‘Principle of Uniformitarianism’ has been much debated and travestied and – as Shea (1982) has forcefully reminded us – it remains a principle to be followed by the geomorphologist, not a ‘truth’ about the workings of the planet and its processes. Nevertheless, most of us would date the origin of geomorphology to Hutton and, possibly more convincingly, to John Playfair’s (1802) extension and elaboration of Huttonian views. It is in the course of that development of an effectively modern geomorphological argument that Playfair described the system of valleys and ‘the nice adjustment of their declivities’ that came to be known as ‘Playfair’s Law’. In so far as fluvial valleys do, indeed, tend to show that mutual adjustment, whereas glacially eroded troughs do not, Playfair’s Law is, indeed, both universal and causal.

The next significant and widespread development of principles and laws came from the United States in the latter nineteenth century: J.W. Powell’s principle of BASE LEVEL; W.M. Davis’s CYCLE OF EROSION and the ‘laws’ of the explanatory significance of ‘structure, process and stage’. But there is also the whole set of propositions, explicitly termed laws, which G.K. Gilbert propounded from his great study of the badlands on the rim of the Henry Mountains, Utah, in 1877. Chorley *et al*. (1964: 550–567) discuss these contributions in some detail, which include: the law of uniform slopes; the law of declivities; the law of structures; and the law of divides. All these are seen as universal ideals, towards which a fluvial landscape will tend, but which it will never attain. One may consider the models of SLOPE EVOLUTION of W. Penck to belong in a similar tradition.

By the middle of the twentieth century, the qualitative laws discussed above were felt to be insufficiently precise. R.E. Horton (1945) set out what became known as HORTON'S LAWS of drainage basin composition (which were expanded by S.A. Schumm and M.A. Melton) covering the regularities in stream number, length, area, frequency and slope as basin order changed. To these was later added Hack's Law (Hack 1957), which relates the length of the longest stream in a drainage basin to the area of the basin (see Rodriguez-Iturbe and Rinaldo 1997). Still later, R.L. Shreve (1966) demonstrated that drainage networks could be considered to follow the statistical law of topological randomness. The most fundamental of these laws are, in some sense, related to the universal properties of networks and it is not altogether evident how far they may be related to basin geomorphology. The law which describes the deformation of ice over time – discovered by J.W. Glen in 1955 – may be more closely linked to geomorphological processes as it successfully predicts the extreme sensitivity of secondary creep in ice to changes in shear stress (Paterson 1981).

There would seem some prospect that the development of approaches such as those examined by Rodriguez-Iturbe and Rinaldo (1997) that are based upon advanced computer modelling and the concept of FRACTALS may, in time, be able to provide quantitative expressions of some of the most basic qualitative geomorphological laws (cf. Rodriguez-Iturbe and Rinaldo 1997, Chapter 6). Until that time, it must be accepted that the key laws and principles of the subject remain verbal rather than mathematical. Geomorphology is still a 'consumer' rather than a creator of quantitative physical laws.

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SEE ALSO: base level; complexity in geomorphology; confluence, channel and river junction; Cycle of Erosion; fractal; glacier; Horton's Laws; slope, evolution; unequal slopes, law of

BARBARA A. KENNEDY

LEACHING

Leaching is the removal in solution of constituents by water or other percolating solutions. It is usually applied to chemical DISSOLUTION and removal as a liquid moves through a porous solid such as soil or a rock mass. Leaching is also used in a broader context to apply to the removal of soluble compounds from solid waste, or the extraction of metals or salts from ores. The most important factor influencing ion mobility is the amount of available water, and that in turn depends on a number of other factors. Leaching directly affects the Eh–pH conditions surrounding minerals and thus determines which elements will stay in solution. Other factors affecting the degree of leaching or weathering include texture, permeability, initial carbonate content, subaerial climate, depth of the wetting front, precipitation of secondary carbonates, soil moisture and temperature, plant transpiration, root mat extraction, porosity of the material, and depth to the saturated zone of ground water.

Leaching affects the alteration of rocks and sediments. As rocks and sediments are weathered they are altered in 3-dimensions beneath the land surface. Layers or zones of altered material may form subparallel to the land surface. Leaching is the mechanism by which dissolved ions and clay are transferred from zone to zone. These layers or WEATHERING zones may be defined in various ways. They may differ physically, chemically and mineralogically from adjacent layers and have lateral extent. The vertical section through a stack of these zones is often referred to as a weathering profile. A soil profile is one type of weathering profile in which layers or horizons are designated. A descriptive terminology for geologic weathering zones has long been in use (e.g. Kay 1916; Kay and Apfel 1929; Leighton and MacClintock 1930; Frye *et al.* 1968) and provides a shorthand for field description of material below the solum (the A and B horizons of a soil). Hallberg *et al.* (1978) discuss the importance of standardizing weathering zone descriptive terminology and subdivide the terminology by material. The terminology describes Quaternary sediments in terms of their colour and the presence or absence of soluble carbonate minerals. In the mid-continent USA, a typical weathering profile in unconsolidated Quaternary sediments such as till consists of an oxidized and leached zone, an oxidized and unleached zone, and a reduced (deoxidized in loess) and unleached zone (Ruhe 1975). In weathered rock, zones may be determined based on the ratio of core stones to weathered matrix (Ruxton and Berry 1957). Examples of these types of weathering profiles are illustrated in Figure 102.

A few caveats are necessary when considering the use of weathering zone terminology. Colour is somewhat problematic as a descriptor because of the concepts associated with weathering zone terms such as deoxidized, reduced or unoxidized. They are not synonymous with reduced chemical states. Second, although abrupt contacts between weathering zones may coincide with stratigraphic breaks, the contacts do not indicate that stratigraphic boundaries are present. For example, a single till unit may have a prominent yellowish-brown oxidized weathering zone overlying a light-grey deoxidized zone. The zones give the unit an appearance of two tills but in actuality there is only one. Weathering zones have been related to various hydrologic conditions. In the absence of continued leaching, weathering zones function as closed systems in which chemical weathering ceases.

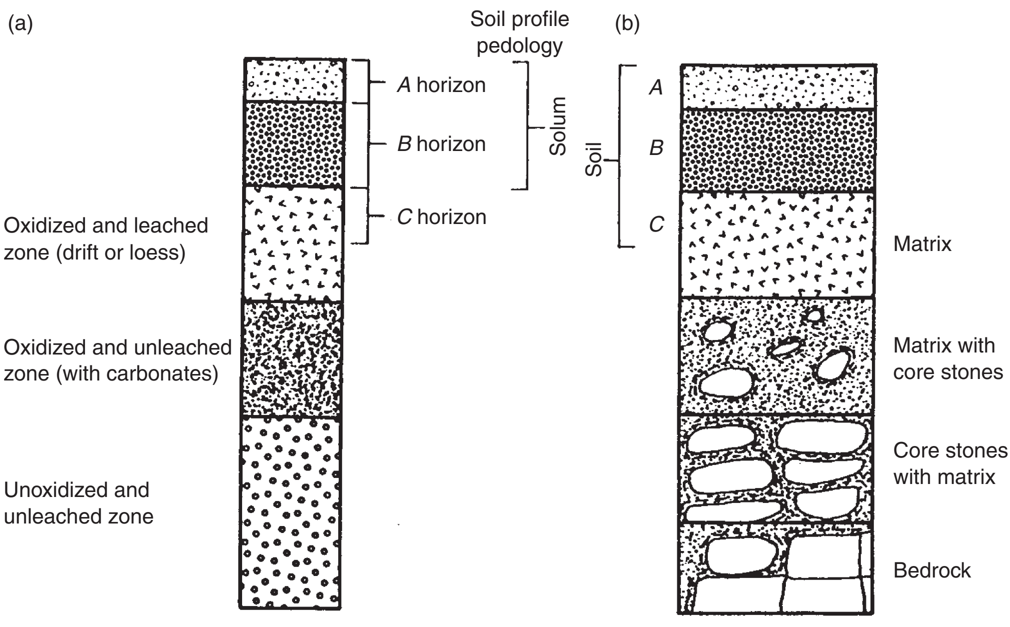

Figure 102 Weathering profiles in sediment (a) and in rock (b) with soil profiles developed in the surface materials for comparison. Modified from Ruhe (1975)

Leaching rates are important in assessing rates of groundwater contamination and plant growth. Organic chemicals are prone to be lost to leaching as their solubility in water increases. The greatest risks for hazardous leaching, rapid and unaltered dispersal away from sources, occurs in highly permeable materials such as sands that have little or no organic material. The leaching capability of introduced chemicals varies but herbicides are usually more mobile than fungicides and insecticides and should be applied with care in areas with very permeable soils.

Cover crops and no-till farming can be effective in reducing some nutrient loss through leaching. For example, the process of crop growth slows percolation and removes nitrate from solution by incorporating it into plant tissue as nitrogen. Unchecked, nitrate leaching through soils and sediments into water supplies is a serious human health and environmental contaminant.

As mentioned earlier, a common determinant for leaching is precipitation, its rate, intensity and duration. A lack of precipitation can lead to an accumulation of mobile constituents such as carbonate and other salts in soils and sediments of arid climates. Conversely, excess precipitation can lead to complete removal of these constituents from sediments in more humid environments. In artificially irrigated crop lands, a leaching requirement (LR) is usually recommended. The LR is the fraction of irrigation water that must be leached through the root zone to control soil salinity at a specific level (Foth 1984). The goal is to sustain a productive soil and produce no change in salinity during irrigation.

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CAROLYN G. OLSON

LEAST ACTION PRINCIPLE

Forms and patterns in natural systems are often the products of optimized circumstances related inherently to operational efficiency. In systems of motion, optimum operational efficiency requires, by the minimization of a system's 'action', the least expenditure of energy for completing a particular task. In other words, out of many possible alternatives nature follows the most 'economical' path. This least action principle (LAP) was formulated originally by numerous mathematical physicists, notably Pierre-Louis Moreau de Maupertuis (1698–1759), Leonhard Euler (1707–1783), Joseph Louis Lagrange (1736–1813), Sir Rowan Hamilton (1805–1865) and Carl Gustav Jacob Jacobi (1804–1851). It contains a curious and subtle twist on Newton's laws, for its variational formulation of motion does not use force and momentum but instead the physical quantities of energy and work. As a result, it often shows structural analogies between various areas of physics and has been found useful in unifying subjects and consolidating theories in various branches of science (Lanczos 1986).

By the end of the nineteenth century, LAP had become a very successful scheme applicable not only in classic mechanics, but also in electrodynamics and thermodynamics, typically through the work of Hermann von Helmholtz (1821–1894). During the twentieth century even more widespread applications were developed and in the 1910s, Albert Einstein (1879–1955) deduced the equations of *general relativity* from LAP. In the 1940s, Richard Feynman (1919–1988) identified the applicability of LAP in quantum physics and since then physicists have found that LAP also underlies the fundamental gauge theories of particle physics, leading to the establishment of what is termed *fundamental* physics (Brown and Rigden 1993). LAP has also been applied widely outside of physics. A notable example is the study of George Zipf (1902–1950), who, within the context of LAP, tried to derive the power-law form of his law for understanding the behaviour of humans on the basis of a principle of least effort (Zipf 1949). With the wide adoption of FRACTAL theory in the 1970s, Zipf's 'law' became ever popular and has been regarded as one of the essential phenomena of nature. It occurs not only in the distribution of words but also in the occurrence of cities, populations, wars, species, coastlines, floods, earthquakes and many other processes and behaviours (Schroeder 1992).

Huang and Nanson (2000, 2002) and Huang *et al.* (2002) have examined the applicability of LAP in geomorphology. Their theoretical inferences and evaluation of a wide range of case studies have shown that LAP governs fluvial systems in the form of MAXIMUM FLOW EFFICIENCY (MFE), providing a soundly based explanation as to why regular bankfull HYDRAULIC GEOMETRY relations occur in very different geographical regions. They also showed that MFE subsumes the previously proposed extremal models in geomorphology of maximum sediment transport capacity and minimum stream power. Further work is examining the application of MFE under various physical constraints (particularly available energy in the form of gradient) to explain the physical conditions for the formation of different river patterns and drainage networks.

SEE ALSO: fractal; hydraulic geometry; maximum flow efficiency; models

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HE QING HUANG AND GERALD C. NANSON

LEVEE

Among the most prominent products of overbank deposition on alluvial and deltaic FLOODPLAINS and submarine abyssal plains are natural levees bordering channels. A natural levee is a wedge-shaped ridge of water-laid, channel-derived sediment which elevation tapers gently into the flanking floodbasin. Moderately to well-developed levees are present along most channel reaches, the exceptions being new, rapidly migrating, or coarse-grained braided channels.

The morphology of a levee depends upon its age, channel size and, in the case of alluvial and deltaic channels, the maximum height to which waters are ponded during floods or storm surges. Vegetation type, grain size and rate of channel alluviation exert a secondary control. Levee heights above the adjacent floodplain range from a few centimetres on young creeks to 5 m along mature sections of large rivers. Heights of submarine levees typically are greater by a factor of at least 10. Levee widths are also highly variable, ranging from fractions of a channel width to over 10 for alluvial and deltaic cases and 10 to 20 widths for submarine levees. Although data are sparse, levee cross-sectional area appears to scale linearly with channel cross-sectional area. Levee slopes transverse to the channel range widely, from virtually horizontal to near-channel values of 6° for fluvial deltaic levees, and to at most 3.5° for levees on the Amazon submarine fan. In alluvial and deltaic levees one levee of a pair is often significantly higher, wider or steeper than the other, with no systematic variation along a channel; in submarine levees of the northern hemisphere, the right-hand levee (when looking downstream) is often higher, sometimes by as much as three-halves, due to Coriolis forces.

The sedimentological characteristics of levees are highly variable. Generally it can be said that levee deposits fine laterally from coarser sands and silts closer to the channel to distal silts and clays with no vertical textural trends. Sedimentary structures and stratification consist of climbing ripple cross-laminated sands alternating with lenticular and wavy-laminated muds and rhythmically bedded, laminated, thin silts. Levees bounding submarine channels generally consist of silty to clayey spill-over turbidites with occasional thicker fine-grained sandstone beds and hemipelagic and pelagic intervals.

Levees arise by the transfer of suspended sediment from the channel to the floodplain via two mechanisms: diffusion and advection. Diffusion occurs when turbid turbulent eddies along the channel-floodplain boundary spin off onto the floodplain and decelerate, allowing grains to settle at distances determined by channel geometry, floodplain roughness, particle size distribution and flow character. Advection occurs when turbid flows leave the channel as channel water overtops the banks during the rising limbs of floods and on the outsides of meander bends.

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RUDY SLINGERLAND

LICHENOMETRY

Lichenometry is a dating technique using lichen measurements to supply relative or absolute dates for rock surface exposure. Geomorphologists have chiefly applied it to MORAINES and periglacial (see PERIGLACIAL GEOMORPHOLOGY) landforms in arctic and alpine environments. Other dating applications have included MASS MOVEMENTS, SCREE and blockstreams (see BLOCKFIELD AND BLOCKSTREAM), NIVATION, avalanches (see AVALANCHE, SNOW) and snow cover, LIMESTONE PAVEMENTS, seismic movements, river channel deposits, lake levels, shorelines, storm beaches and archaeological features.

The technique's dating range depends on lichen form, competition and the local environment. In temperate regions some foliose forms might survive 150 years; minutely crustose forms can provide dating over 400 to 600 years and dating may exceed 1,000 years at high latitudes. All dates can only be empirically justified unless validated by an independent source. Claims have been made for a dating range of 8,000 to 9,000 years based on radiocarbon dates for organics beneath moraines, but such a range is unlikely due to rock WEATHERING, successive glacier advances and climate change. Lichens only provide minimum dating for the last period of surface exposure while radiocarbon determinations provide maximum dates.

Three main approaches have been developed: the original approach is based on size/age correlations of largest lichens; the other two approaches are based on population size–frequency distributions, one using measurements of populations of largest lichens on uniform-aged surfaces and the other, measurements of whole lichen populations, with a population defined by its individual rock surface.

The original approach

In the original approach, pioneered in 1950 by Roland Beschel, lichen species cumulative growth rates are derived indirectly from size/age correlations of near-circular lichen bodies (thalli) growing on known-age surfaces. The longest axes of largest lichens on each surface are plotted on a graph, thallus size against date. The species cumulative growth rate is assumed to be that described by the curve traced through the largest size/age plots in the distribution. There may be a delay before species start colonizing a fresh surface; establishing this period relies on projection of the plotted growth curve on to its time axis. Absolute dates for a rock surface are obtained by fitting the largest lichen measurement to the local growth curve and relative dating may be achieved by drawing lines of equal maximum growth around areas defined by largest lichens of equal size.

Lichenometry is frequently perceived as a quick, easy, cheap method for use in areas where other dating tools are lacking. However, it has often been misapplied and results (only justifiable empirically) have been at best questionable. A review of the technique (Innes 1985) describes the difficulties surrounding species identification, particularly those in the yellow-green *Rhizocarpon* section *Rhizocarpon* group containing the species most frequently used (and confused) in lichenometric studies and the various methodologies employed, many of which were developed in attempts to resolve uncertainties arising from Beschel's five initial assumptions:

1. Thallus size correlates with age.
2. Colonization occurs soon after rock exposure.

3 Largest thalli are founder members of their populations.
4 Variations in growth rate due to habitat differences are minimized by selection of largest thalli growing in optimal conditions.
5 Accurate species identification in the field.

These are approximate assumptions because species growth rates vary over time and space, sensitively depending on habitat and climate, and identification is notoriously difficult for the non-expert. In addition, although lichens can theoretically colonize almost any bare rock surface, this may not occur until surfaces are WEATHERED. Consequently, both growth rates and delay before colonization should be confirmed at each dating site, a difficult procedure in places where there are no dated surfaces for growth-rate calibration.

Additional problems for this approach introduced during its development, arose as a result of misconceptions of the nature of lichen growth and confusions in terminology. The two terms causing the greatest confusion are 'great period' and 'lichen factor'.

Beschel (1961: 1,046, 1,057) described lichen growth as sigmoidal: beginning very slowly then gathering speed with many thalli passing through a 'great period' (limited to a few decades) before dropping to a long-term constant value. However, geomorphologists, due to a scarcity of younger thalli on surfaces of known age in their field areas, have often missed the initial part of the growth curve and assumed that the 'great period' is represented by linear growth over the whole historical period, possibly lasting some 300 years, with this followed by apparently declining growth, the decline being suggested by a curve drawn through very few lichen-size/radiocarbon-age correlations. However, as noted above, radiocarbon cannot reliably be used to date the latest period of moraine exposure and protracted growth curves constructed on this basis are therefore questionable.

These misconceptions contributed to the confusion over interpretation of the 'lichen factor' a term employed in many studies to describe the average value of maximum growth over 100 years, despite Beschel's (1961: 1,055) assertion that when calculating a 'lichen factor' the 'great period' should not be taken into consideration and that averaging cannot describe growth rates. This is because a straight line drawn from its origin and projected over 100 years will only cut a sigmoid curve in one place. Beschel's intention was that the 'lichen factor' should be used as a measure of hygrocontinentality and he defined the term as a growth velocity gradient produced by low precipitation at high altitudes that could be a helpful indicator when planning mountain reservoirs. And he warned that no standard growth velocity could be expected across a large region.

Population size-frequency approaches

In attempts to resolve some of the uncertainties in the initial assumptions and supply a measure of dating independence, two population size-frequency approaches have been developed. In the first, a composite curve is computed for the frequency distribution of large samples of longest axial measurements of largest lichens growing on uniform-age surfaces. The curve can be decomposed into subpopulations describing discrete events in their relative-age order. Where absolute dates are required, means of the largest lichens on each event surface are either plotted or regressed against dates of historical events to obtain a regional growth curve. Such a curve is, however, questionable (as Beschel warned) and requires careful testing before acceptance.

The advantages of this approach are that it can be used to investigate the history of seismic and large diachronous surfaces; statistical methods, with error limits and modelling can be applied, and anomalous lichens (either coalesced thalli or survivors from an earlier population) are less likely to corrupt the data.

In the second size-frequency approach, longest axial measurements are taken of whole lichen populations, with populations growing on surfaces of similar aspect and lithology and containing not less than thirty individuals, but ideally over 100 (statistical tests can be used to show where the smaller populations may safely be grouped). Changes in the shape of unimodal population size-frequency distributions show the relative age of populations. Bi- or multi-modal distributions indicate surface changes that have partially removed lichen thalli creating space for new colonizers; isolated large thalli may either be survivors or anomalous growths.

The advantages of this approach are similar to those of the largest lichen size-frequency one, but in this case a single boulder's whole colonization history is reflected by its population distribution, and comparison of its history with the histories of other surrounding surfaces provides insights into micro-environmental effects; showing for example, colonization rates differing on surfaces of differing aspects. Hence, colonization delays should be investigated before dating the proximal and distal sides of moraines.

The limitations of these two approaches are that a very large number of lichen measurements are required and the quality of absolute dating depends on the reliability of the independent dating source. Consequently, neither size-frequency approach achieves the desired status of full independence in absolute dating. However, because of large data sets the uncertainties implicit in the initial assumptions are less important.

Lichenometry is a reliable relative dating tool and, used with proper care, can provide revealing insights into land-forming processes especially in locations where the technique can be used within some independent dating framework supplied, for example, by historical records or DENDROCHRONOLOGY. In these circumstances absolute dating, frequencies and rates of change can be established with a reasonable degree of confidence.

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VANESSA WINCHESTER

LIDAR

LIDAR light detection and ranging, is a form of airborne scanning altimetry which is of great importance for mapping landforms and landform change, especially at relatively local scales. It can be used to produce DIGITAL ELEVATION MODELS and provides high-resolution data on topography. It has been used in a number of geomorphological applications including cliff and landslide monitoring, the study of tidal channels, assessing subsidence risk, predicting areas subject to storm surges and tsunamis and changes in beach height (see, for example, Adams and Chandler 2002; Brock *et al.* 2002; Stockdon *et al.* 2002).

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A.S. GOUDIE

LIESEGANG RING

Liesegang rings are alternating concentric iron-rich and iron-depleted shells which are formed by differential chemical leaching and precipitation in rocks exposed to WEATHERING. The rings (shells) are developed in rock blocks, typically 2–20 cm in diameter and 10 cm thick (Liesegang blocks), bound by the tectonically induced joints at the periphery, and joint or bedding planes at the bottom. The Liesegang rings follow the configuration of the outer shape of the blocks. Structurally, the Liesegang blocks can be divided into four types: (a) primitive (single shell), (b) ordinary (multi-shell), (c) compound (multi-pattern), and honeycomb (cylindrical) Liesegang blocks. The most important factors involved in transformation of a bed of rock into Liesegang blocks are: a condensed grid of joint polygons, surface water, the appropriate topographic site of exposure, and composition, texture and thickness of the beds. Field, hand specimen, microscopic and geochemical studies suggest two opposing diffusion direction trends in the course of Liesegang ring formation: one mainly Si-trend (outward), and one mainly Fe-trend (inward).

The effect of recent tectonic movements on the Liesegang blocks manifests itself in various forms. Of interest is the partial replacement of the earlier Liesegang patterns by the later patterns along joints, and formation of compound Liesegang blocks (Liesegang blocks with more than one set of Liesegang patterns). From the study of Liesegang patterns within the compound Liesegang blocks, the configuration of the joint polygon related to each Liesegang pattern and thereby the sense of stress field migration in the area, can be deduced.

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SEE ALSO: WEATHERING

JAMSHID SHAHABPOUR

Plate 73 Stepped limestone pavement with well-developed karren features, Triglav area, Julian Alps, Slovenia

LIMESTONE PAVEMENT

Limestone pavements are stripped bedding-plane rock surfaces constituting complex polygenetic assemblages of KARREN landforms, or Karrenfeld. The limestone's bedding planes are scoured of mature weathering forms and debris producing a surface on which develop solutionally widened fissures (grikes, Kluftkarren) and residual limestone blocks (clints, Flachkarren).

Adjectives such as 'regular' or 'rectangular' popularly describe clint and grike patterns, reflecting DISSOLUTION along the two main joint directions at right angles usually found in limestone massifs. However, field situations show highly complex and variable dissection patterns. In addition, not all limestone pavement areas are on single bedding planes. They can be benched or stepped across several beds depending on the limestone outcrop (Schichttreppenkarst, sheet-stepped karst) (Plate 73, and see KARREN, Plate 67). Whilst the term 'pavement' implies a walkable surface, there is a spectrum of surface limestone bedding-plane landforms, from highly dissected, almost shattered strewn rock fields, to virtually undissected rock sheets.

Morphometric work (see GEOMORPHOMETRY) comparing numerous pavements identifies typical ranges of clint and grike dimensions (Goldie and Cox 2000). Clints up to 1 m long are commonest, but many are longer, typically up to several metres. Whilst most widths are up to 500 cm, many are over 1 metre. Most grikes are up to 50 cm wide, whilst grike depths have a wider range, between 20 and 150 cm. These ranges, however, do not encompass all limestone pavement surfaces. For example, although 95 per cent of measured grikes were less than 30 cm wide, the absolute range is from 1 cm to over 1 m; although a 'typical' grike width is 13 to 16 cm. Grike depths showed a 70-fold variation from 4 to 274 cm but focus on 100 cm. A rather weak relationship between grike widths and depths suggests that their horizontal and vertical development are not strongly linked.

Clint shape as demonstrated by a clint width to length ratio (theoretically ranging from 0 to 1) is interesting in relation to conventional ideas of pavements, as a ratio of 0.4 is four times more common than a ratio of nearly 1 (indicating square). Very elongated clints are also rare (ratios close to 0). This implies that one joint direction dominates the planform of limestone pavements. A complication, however, is that the two main joint directions in limestone massifs are frequently bisected, resulting in many triangular and diamond-shaped features, causing problems for measurement and interpretation of rectangularity.

Beyond measurability are features of extreme or minimal dissection. Where clints are too narrow or small they move easily and fall over, failing to satisfy the walkability concept. These should be excluded from the limestone pavement definition. Harder to exclude, however, is the undissected extreme where the surface lacks grikes but is certainly walkable. Although lacking the karren forms expected on limestone pavement, such bare rock expanses can really only be called pavement. The more dissected extreme shows a geomorphological transition to a rock field.

Plan morphometric data says nothing about cross-sectional form of clints and grikes, important for understanding pavement weathering sequences. Depth and intensity of surface weathering may not affect clint area but changes their vertical relief (see Figure 103). For example, clints become pinnacled or very well rounded where solutional weathering is advanced (Plate 74).

The main factors producing a good limestone pavement are strong limestone, with some fractures, but not so fractured as to weaken the beds too much. Scouring is necessary to clear weathered debris off bedding planes, and pavements then need to be in a favourable solutional environment, i.e. reasonably wet and possibly covered with soil and vegetation, to develop their characteristic karren. Then that cover needs removal.

Glaciation is the scouring mechanism creating most of the world's limestone pavements. Other scouring agents include marine stripping, scarp retreat in semi-arid conditions and even human removal. In the main, though, distribution of limestone pavements is related to glaciation and thickly bedded, strong, pure limestones. Thus there is much in the younger mountain ranges such as the Alps, Rockies and Himalayas and in shield areas, such as northern Canada. Vast expanses of pavement on dolomites are found in northern Canada.

Rock type influences pavements at several scales. The best, most persistent, pavements develop on pure, mechanically strong and massively bedded limestones and dolomites. Similar landforms can occur on sandstones but weathering and soil development here generally mean that the features do not last except in situations such as waterfall or river beds where constant fluvial action keeps jointed outcrop clear. Pure limestones stay clear as soil formation is exceptionally slow due to the lack of insoluble residue, and the

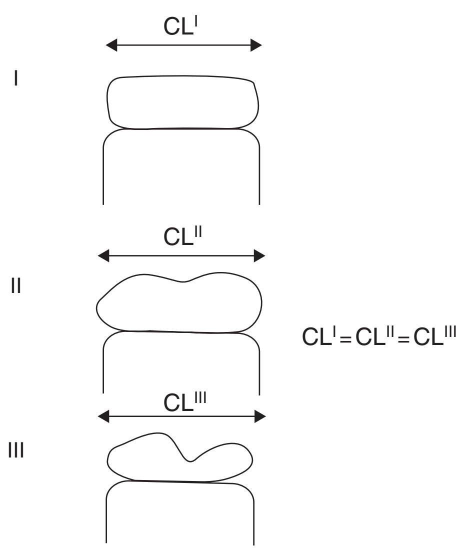

Figure 103 Simple cross-sectional diagrams of three clints demonstrating how plan measures of clint size (CL^{I}, etc.) can be identical on clints with very different degrees of down-wearing

Plate 74 Limestone pavement with a slight slope (5° to 8° down away from viewer) at Gaitbarrows, Cumbria, UK. There are several stages of grike development from relatively immature short slits to grikes running for tens of metres. Other karren features are less well developed

swallowing of material by developing solutional fissures. Thin or weak limestones (internal weaknesses, either horizontal weaknesses or frequent vertical joints within a given bed) favour weathering both by mechanical and solutional means, permitting easy rock breakup unable to sustain a pavement.

Slope significantly affects both how glacial scour operates and the subsequent pavement surface development by solution (see Plate 67). Sloping layered outcrops provide complex situations for glacial scour. The postglacial surface may be interrupted by areas where shelter from the ice allowed preglacial karstified features to survive. This is found in north-west England and in the Alps, for example. Some pavements may also have characteristics surviving from earlier phases of karstification (see PALAEOKARST AND RELICT KARST).

Another factor affecting pavement form and distribution has been human activity. Some limestone pavements are anthropogeomorphic (see ANTHROPOGEOMORPHOLOGY), indirectly because their exposure results from soil erosion, and directly because of clint removal by humans for resource reasons. Because their botanic as well as landscape value is threatened by this, limestone pavements in the UK are now legally protected (Goldie 1993).

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SEE ALSO: karren; karst; limestone pavement

HELEN S. GOLDIE

LINEATION

Lineations are linear patterns observed on imagery that represent fractures. The fractures may be either joints or faults (see FAULT AND FAULT SCARP). Each lineation does not necessarily represent an individual fracture, and typically represents a fracture zone. Lineations include, but are not limited to, actual joints in bedrock, straight stream segments, linear alignments of natural vegetation, aligned topographic features and linear changes in image tone or texture. Any feature thought to be a lineation should be one of a group of parallel features, and each lineation and group (set) may consist of different types of patterns. For instance, a line of vegetation may continue as a linear pattern of dark-toned soil across a light-toned field, and then as a stream valley.

Lineations are readily apparent on all scales, resolutions and types of imagery – aerial photographs, radar and various types of satellite imagery – on hard copy, as well as in digital form. The delineation of lineations is, however, both art and science (Wise 1982). For best results, and particularly if quantitative results are desired, the imagery should be vertical or near vertical. Certain features, such as those noted above, define lineations, but the interpretation of these features, particularly tonal and textural variations, can be difficult. The eye must be trained to see them, and the more experience one has, the greater the number of lineations that can be identified with confidence on any given image. All hard-copy images should be evaluated meticulously from different angles and with different kinds and angles of illumination. Some lineations will be obvious in sunshine, for example, but cannot be seen in natural light on a grey, overcast day. Topographic lineations are often more easily identified using stereo imagery, but those defined by vegetation or tonal variations can be readily – and more easily – identified monoscopically. Although attempts have been made to automatically delineate lineations on digital imagery, they have not been generally successful (e.g. Ehlen *et al*. 1995). Finally, the individual delineating the lineations should also be careful not to confuse linear man-made features, ancient or modern, with natural lineations.

Stereonet projections can be used to display lineation orientation data so that sets can be identified (dips are assumed vertical). Different patterns are indicative of fracture formation under different tectonic conditions, and the chronology of the different events can be identified using cross-cutting relations between individual lineations and lineation sets. When the lineations are in digital form, quantitative information on length and spacing (frequency) can also be obtained. These types of data are useful with respect to landform evolution and engineering geomorphology, and can provide information on fracturing in areas that are difficult to access in the field. Furthermore, lineations can be directly related to joint or fault data identified on the ground. Several studies have identified lineations on imagery, then located the individual lineations on the ground (e.g. Lattman and Parizek 1964). Other studies have shown statistical links between joints and lineations (e.g. Ehlen 1998, 2001).

Another term often used for lineation is 'lineament'. Lineaments are in fact subsets of lineations: they are topographic features and, for example, do not include linear patterns produced by tonal changes, which are among the most important indicators of fractures on imagery. Examples of lineaments include aligned saddles or ridge lines and aligned stream valleys.

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SEE ALSO: jointing; remote sensing in geomorphology

JUDY EHLEN

LIQUEFACTION

Liquefaction is the transformation of granular material from a solid to a liquid by an increase in pore pressure. Liquefaction is associated with earthquakes, with saturated ground, with granular sandy/silty soils, with some spectacular building collapses, and with submarine landslides and waste tip failures. During the 1964 earthquake at Niigata in Japan large apartment buildings were gently tilted by as much as 80º on to their sides, with no structural damage, as a result of liquefaction of a shallow sand bed underlying the structures.

In the Kobe earthquake of January 1995 widespread damage occurred. Massive soil liquefaction was observed in many reclaimed lands in Osaka Bay including two man-made islands. Both islands had been constructed from fill material derived from decomposed granite. The grain size of the fill varied from gravel and cobble-sized particles to fine sand, with a mean grain size of approximately 2 mm. In the Port Island ground, liquefaction of this fill material (the thickness of the submerged fill in this island was about 12 m) resulted in settlement of between 0.5 and 0.7 m. Soil liquefaction also occurred around the port of Kobe and caused extensive damage to many industrial and port facilities such as tanks, wharves, quay walls, cranes, and the collapse of the Nishinomiya Harbour Bridge.

A catastrophic liquefaction failure occurred in a coal waste tip at Aberfan in Wales in 1966. The tips from the Merthyr Vale Colliery were sited on the upper slopes of the valley sides, directly above Aberfan, and some 100 m up on a 13º slope. These slopes consisted of the permeable Brithdir sandstone, with many springs and seeps in evidence. Tip no. 7 was built over these springs and in October 1966 a flowslide developed in the saturated material – a classic liquefaction event. The liquefied tip material ran into the village and killed 144 people, most of whom were children in the village school.

In the Netherlands liquefaction events often concern flowslides in underwater slopes of loose sand. Flowslides are observed after sudden liquefaction of the sand under static loading. The liquefaction is caused by the contraction (packing collapse) occurring simultaneously in a large volume of sand and the impossibility of rapid drainage of the completely saturated pore fluid. The upper 10–30 m of soil in the western part of the Netherlands consists of Holocene layers of clay, peat and sand. Tidal estuaries were formed in Zeeland, the south-west part of the country, from the beginning of the Holocene. The position of the tidal estuaries, however, has shifted in an alternate process of rapid erosion and sedimentation. Rapid sedimentation has resulted in thick layers of loose sand at many locations. Median grain diameter is around 200 μm; with a silt content of perhaps 3 per cent. Liquefaction is accompanied by sudden pore pressure increase as the open sand packing collapses from a metastable state to a more stable state.

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SEE ALSO: quickclay

IAN SMALLEY

LITHALSA

The term lithalsa was used for the first time by S. Harris (1993) to describe mineral mounds in Yukon (Canada). They are mounds similar to PALSAS but without any cover of peat. The upheaval of the soil is the result of the formation of segregation ice in the ground. These mounds which, like the palsas, are found in the domain of discontinuous PERMAFROST, were successively called purely minerogenic palsas with no peat cover, mineral palsas, cryogenic mounds, mineral permafrost mounds and palsa-like mounds. Lithalsas are presently known only in Subarctic Northern Québec and Lapland in places where the temperature of the warmest month is between +9 and +11.5 °C and the mean annual temperature between −4 to −6 °C.

Remnants of lithalsas formed during the Younger Dryas exist in Ireland, Wales and on the Hautes Fagnes plateau in Belgium (Pissart 2000). They are closed depressions surrounded by a rampart. These features were first explained as remnants of pingos. They are now regarded as periglacial phenomena which provide more precise palaeoclimatic indications.

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ALBERT PISSART

LITHIFICATION

The process by which unconsolidated sediments become indurated sedimentary rock, originating from Greek and Latin for ‘to make rock’. The source of sediments is typically loose material that accumulates upon the surface as a result of subaerial weathering. These are then transported and deposited before the process of lithification begins. Lithification precedes DIAGENESIS and must not be confused with the term (diagenesis refers to the processes and their products affecting rocks during their burial). The process of lithification involves three main components. First, desiccation (drying of sediment) reduces the amount of pore space within the sediment by eliminating the water within the material. Desiccation is particularly significant in fine-grained sediments (clays and silts), where water cannot percolate easily through the pores. Compaction further reduces pore space, aided by the increasing weight from the materials above as surface deposition continues. Surface tension between the individual grains then allows them to act as a consolidated mass.

Second, cementation binds the loose sediment together by filling in the remaining pore spaces. It is typically done through precipitation of cementing agents in solution as water flows through the pore spaces. The most common cementing agents are calcite ($CaCo_3$), dolomite ($CaMg[Co_3]_2$), quartz (SiO_2), and iron oxide (Fe_2O_3) (haematite).

The pH of a solution plays a significant role in determining the cementing agent type. An acidic solution will tend to produce a quartz cementing agent, whereas an alkaline solution will invariably generate a calcitic or dolomitic cement.

Third, crystallization is a specific form of cementation. It is particularly effective on carbonate sediments, whereby crystals form in pore spaces from minerals in solution and bond onto existing crystals within the sediment. This often has the side effect of making some rocks harder.

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STEVE WARD

LOESS

Loess is a sedimentary deposit, largely composed of coarse and very coarse silt, which is draped over the landscape. The silt is largely quartz and, at the soil structure level, the loess particle system has an open packing, which is a result of aeolian deposition during Quaternary times. This open structure gives rise to the main practical problem; when loess ground is loaded and wetted the structure may collapse and subsidence occurs. Construction on loess has to be carefully designed in order to avoid these subsidence problems. The silt-sized particles and the open metastable structure are the main characterizing features of loess deposits.

The literature on loess is vast and complex; the major loess languages are Russian, Chinese, English, German and French. The largest literature is in Russian (see Kriger 1965, 1986; Trofimov 2001) but there are major works in many languages. Scheidig (1934) was for many years the standard work, possibly until supplanted by Pye (1987) and Rozycki (1991). Early literature has been collected (Smalley 1975) and an outline bibliography attempted (Smalley 1980). Loess investigation has had a major part in the activity of the International Union for Quaternary Research (INQUA), which publishes the one specialized journal – *Loess Letter*. The one institute in the world totally devoted to the study of loess is the Xian Laboratory for Loess and Quaternary Geology of the Chinese Academy of Sciences; there is a focus of geotechnical activity at Moscow State University and at the G.A. Mavlyanov Institute of Seismology in Tashkent.

Loess really is an aeolian material; it is the aeolian factor which gives it its defining characteristics, and separates it from other deposits. Loess is a wind-blown silt; the aeolian deposition accounts for the open metastable structure, which accounts for the propensity to collapse when loaded and wetted. The wind-blown deposition mode accounts for its characteristic geomorphological position – the 'draping' over the landscape – of the loess on the ground, thinning away from source. Loess in the air is a dust cloud, travelling in low suspension over a relatively short distance. In parts of the world where the geomorphology is favourable this dust has been depositing for millions of years (but see also Pecsi 1990).

Loess formation

An area rich in controversy and argument; how do loess deposits form? Even in the twenty-first century the tag-end of this debate still carries on – but only in the Russian literature. By the end of the nineteenth century it was fairly widely accepted that the key event in the formation of a loess deposit was deposition by aeolian action. This might be called the mainline loess view; loess is an airfall sediment.

But starting early in the twentieth century and continuing fitfully throughout the entire century there has been an alternative view of loess formation, which has generated a large literature, from proponents and opponents. This can be called the 'soil' theory of loess formation, or the '*in situ*' theory. At its heart is the idea that loess is formed where it is found, by a process of 'loessification', the conversion of non-loess ground into loess ground. This is largely a Russian theory and its chief, and very forceful protagonist, was L.S. Berg. The minor role was played by R.J. Russell in the USA who put forward a very similar theory to explain the Mississippi valley loess in 1944.

Another question, which arose somewhat later in the history of loess investigation, was that of the possible mechanism for the formation of the vast amounts of quartz silt that were required to make large loess deposits. What sources of geo-energy were available for large-scale silt formation? And did this affect the nature and distribution of loess deposits? This question connected in particular to the problem of 'desert' loess.

V.A. Obruchev, early in the twentieth century, placed loess at the fringes of hot, sandy deserts. This desert loess has been something of a problem ever since. B. Butler, a noted Australian soil scientist, suggested that it did not exist, because he could not find it in Australia; and Albrecht Penck, a famous geomorphologist, suggested that peri-Saharan deposits were essentially lacking. So the question arose, could large hot sandy deserts generate enough loess-sized material to form significant loess deposits? In geographical terms it appears that the Chinese and Central Asian deserts do have major loess deposits close by, but the Sahara and Australian deserts do not. There is a loess-like material, called 'PARNA' by Butler, in south-east Australia, but it has modest extent. It has been proposed that the Central Asian and Chinese loess deposits are also near very high mountain regions, and that these are the real source of the loess particles, with the deserts simply acting as holding areas or reservoirs.

Loess distribution

There are some very large deposits, and many widespread smaller ones. The classic deposit is in China, in the north and north-west of the country; this is the 'Yellow Earth', which gives colour to the Yellow River, and played a major part in the development of the Chinese civilization – the only one of the ancient civilizations which has lasted into modern times. Deposits in Lanzhou are believed to be over 300 m, perhaps over 400 m, thick. Although loess is thought of as a Quaternary material there are suggestions that loess deposition in China has been going on since the Miocene, i.e. for over 20 million years.

The other large deposits are in central North America, relating to the Great Plains, and the Dust Bowl; in South America, the Pampas; and in Europe. The European deposits are complex but might be divided into the northern glacially related deposits, and the Danubian deposits which derive their materials from mountain sources. Of the smaller deposits those in North Africa and in New Zealand are of interest. There are deposits near the coast in Tunisia and Libya which are certainly very loess-like, but the particle size is rather large. The Sahara is a great source of wind-blown dust but most of it is very small, which travels in high suspension for great distances. In New Zealand there are significant loess deposits in the North and South Islands, associated with the mountains. New Zealand is the only country which has a monograph devoted to the national loess (Smalley and Davin 1980) and it is to be hoped that more countries will follow this lead.

With the disappearance of the Soviet Union and the revision of geography in the fringe regions new loess-rich countries have emerged, in particular Ukraine and Uzbekistan. Ukraine has large loess cover and in many parts chernozem soils have developed in the loess providing top-grade agricultural land. Also, in Kyiv is the most amazing loess building, the Pecherskaya Lavra, the Caves Monastery, where the monks have excavated into the loess and developed a subterranean complex in the Dnepr loess. In Uzbekistan, in the eastern part, near the Tien Shan mountains, are major loess deposits. The capital city of Tashkent is built on loess and was largely destroyed by a large earthquake in 1966. Loess ground is very vulnerable to earthquake shocks.

Loess stratigraphy

The basic idea of loess stratigraphy was invented by John Hardcastle in Timaru, New Zealand in 1890; this was that a loess deposit can give a good indication of past climates. Loess acted as a 'climate register'. At the beginning of the twentieth century problems of climate change were exciting quite a lot of interest and this drove a large research effort devoted to loess stratigraphy.

Hardcastle's insights were ignored; as in the case of Mendel and genetics, the scientific culture was not prepared for them. They were addressing a problem that had no structure, no framework, no reference points. It appears that stratigraphic ideas re-emerge with Soergel in Germany in 1919 and developed slowly in Europe. The 'eureka' moment came at the INQUA Congress in 1961 in Poland when Liu Tung-sheng presented the work of the Chinese investigators which showed multiple palaeosols in the Luochuan loess. This thick loess section showed alternating layers of loess and palaeosols. The palaeosols indicated warm periods and the loess layers cold periods – a climatic indicator as Hardcastle had suggested. The data from Luochuan suggested many climatic oscillations in the Quaternary period and laid the foundations for continuing investigations. It was a giant step away from the simple four event Quaternary derived from Alpine observations.

Applied geomorphology

Loess drapes the landscape; it is the accessible ground in which engineers operate. The most significant and expensive of the loess ground engineering problems is hydroconsolidation and soil structure collapse, caused by loading and wetting, and leading to subsidence and structural failure. This was classically a problem in the Soviet Union and now occurs in post-Soviet states like Ukraine and Uzbekistan. The building of irrigation canals in loess in Uzbekistan during one of the early five-year plans alerted the Soviet authorities to the vast problems of subsidence. It was, more than anywhere, a Soviet problem. Therefore most of the subsidence literature is in Russian (see Trofimov 2001 for a good review) and tends to conform to the requirements of the *in situ* approach to loess formation.

A major topic in the Russian literature is 'how did collapsibility develop?' This has never been discussed in 'western' literature because the reason for collapse is implicit in the aeolian deposition mechanism which forms loess deposits. In the Russian literature it appears to be the applied geomorphologists and ground engineers who continue to cling to the *in situ* theories of loess formation.

The problem of soil erosion falls within the purview of applied geomorphology. Because of its silty nature loess soil is very prone to erosion, by wind and water. The classic wind erosion events (e.g. the Dust Bowl) tend to be the blowing away of loess soil material; and the major water erosion problems are often loess connected – in particular the loss of soil material in north China. It is soil erosion that makes the Yellow River yellow. In north-west China where the loess is spread over mountainous terrain, there is a considerable landslide problem (Derbyshire *et al*. 2000). In 1920 a large earthquake caused much loss of life because of widespread collapse of loess tunnels which housed a large proportion of the population.

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SEE ALSO: aeolian geomorphology; palaeoclimate; wind erosion of soil

IAN SMALLEY

LOG SPIRAL BEACH

There are several terms for the plan shape of asymmetrical beaches and bays, including log spiral, half-heart, crenulate, hook-shaped and zetaform. BEACHes between headlands consist of: a curved, nearly circular, portion in the lee of the headland, which may be absent in some areas; a logarithmic spiral section; and a nearly linear to curvilinear reach tangential to the downcoast headland. Yasso (1982) proposed that plan curvature is described by a logarithmic spiral law, with the distance from the centre of the spiral (r) to the beach increasing with the angle θ according to:

$$r = e^{\theta \cot \alpha}$$

where: θ is the angle of rotation, or spiral angle, which determines the tightness of the spiral; and

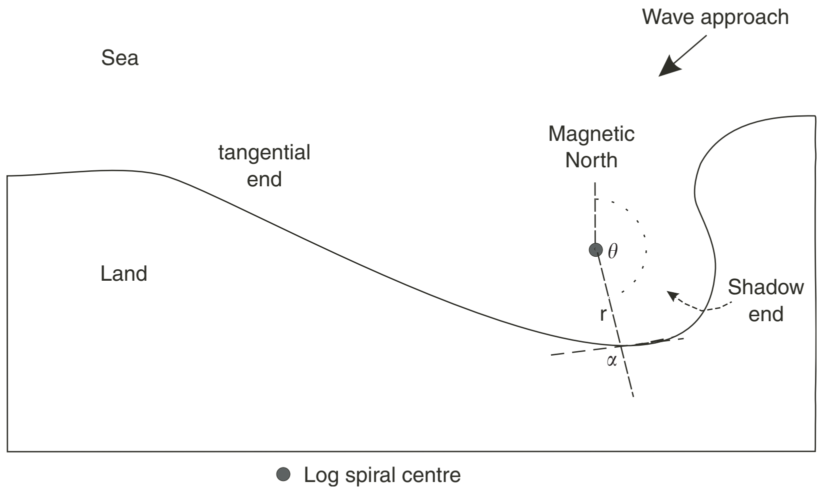

Figure 104 Log spiral beach and Yasso's (1982) logarithmic spiral law

α is the logarithmic spiral constant, the angle between a radius vector and the tangent to the curve at that point – this is constant for a given log spiral (Figure 104).

There may be systematic variations in beach sediment and morphology along log spiral beaches corresponding to longshore changes in wave height and energy. Beaches in California are finer grained and more gently sloping in the sheltered portions of log spiral bays, but they tend to be steep or reflective in southeastern Australia, while sections exposed to the dominant swell and storm waves are gentle or dissipative. The sheltered portion of mixed sand and coarse clastic bays in Alaska generally have low wave energy, small grain sizes, fairly well-sorted sediment, gentle beachface slopes and eroding shorefaces. The central portions have high wave energy, large grain sizes, poor sorting, moderate beachface slopes and shorelines transitional between erosion and deposition. The tangential ends of the bays are similar to the shadow ends, except that the beachfaces are steep and the shorelines depositional (Finkelstein 1982).

Progressive decline in beach curvature downdrift of headlands is usually thought to reflect increasing exposure to wave action, although wave energy is only one of the factors that must be considered. Shorelines try to attain equilibrium conditions determined by offshore wave refraction and diffraction, the distribution of wave energy, rates of longshore sediment transport and the relationships between beach slope, wave energy and grain size. Log spiral beaches formed by oblique waves are generally thought to be the most stable in nature. They are in static equilibrium when the tangential downcoast section is parallel to the wave crests, and as waves diffract into the bay they break simultaneously along the whole periphery. There is no longshore component of breaking wave energy and no littoral drift, and the plan shape, local beach slope and sediment size distribution are constant through time.

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ALAN TRENHAILE

LONG PROFILE, RIVER

A graph representing the relationship between altitude (H) and distance (L) along the course of a river, expressed by

$$H = f(L)$$

Any of three functions – exponential, logarithmic or power – can provide a reasonable fit to stream profiles and give rise to smooth, concave-upward curves (see GRADE, CONCEPT OF). Most long profiles tend to be concave, but they are not invariably smooth. Local steepening of channel gradient (see WATERFALL and KNICKPOINT) can result from such causes as more resistant bedrock strata, the introduction of a coarser or larger load, tectonic activity (see, for example, Riquelme *et al*. 2003) and BASE LEVEL changes (Knighton 1998) associated with REJUVENATION. Long profiles of glaciated valleys are often characterized by steps and overdeepenings and by hanging tributaries (MacGregor *et al*. 2000).

Some rivers have convex long profiles, and this may be a characteristic of updoming passive margins to continents or of discharge reductions downstream as is found in arid zone rivers (e.g. in the Namib).

Excessively or overconcave rivers occur in rivers with lower reaches that have become recently infilled by, for example, estuarine sedimentation or which have been affected by glacial diversion and subsequent lengthening of their courses (Wheeler 1979). At a local scale river long profiles may be punctuated by pool and riffle topography, and by the development of dune bedforms (see STEP-POOL SYSTEM).

The inverse relationship between channel gradient and discharge recognized by Gilbert (1877) helps to explain concavity, since tributary inflows cause a downstream increase of discharge which enables the stream's sediment load to be transported on progressively lower slopes. When discharge increases rapidly downstream with increasing contributing area, profile concavity is greater (Wheeler 1979). In addition, the calibre of sediment load is also related to stream gradient (see Richards 1982, for a review). However, whereas discharge is clearly an independent control of stream gradient, causation is less obvious in the gradient–sediment size relationship, for there are complex feedbacks between sediment size and gradient.

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A.S. GOUDIE

LONGSHORE (LITTORAL) DRIFT

Longshore sediment transport is the net displacement of sediment parallel to a coastline. Such transport is maximized where WAVES (especially *breaking waves*) and wave-induced, shore-parallel, quasi-steady *longshore currents* (see CURRENT) prevail, i.e. in the *surf zone*. A secondary peak also occurs in the *swash zone*. Here wave energy is finally dissipated in the reversing currents of the wave *uprush* (flow at an angle to the shoreline under oblique wave approach) and *backwash* (flow normal to the shoreline under the force of gravity). This gives rise to a 'zig-zag' motion of water and sediment along the beach face. In both cases the obliqueness of the angle of wave approach is critical to the transport rates. The volume or mass of sediment transported by *longshore currents* and *swash* is termed *littoral drift*. Sediment may move as BEDLOAD, in continuous contact with the bed, or as SUSPENDED LOAD. In theory, particles in suspension are supported by turbulent (Reynolds) stresses within the fluid, while bedload particles are supported by inter-granular impact forces (Bagnold 1963). Debate exists concerning their relative importance. For example, Komar (1998) suggests that bedload comprises upwards of 75 per cent of the total transport, while Sternberg *et al*. (1989) suggest that in the surf zone, where waves are the pre-eminent entraining mechanism, suspended load can account for virtually all the longshore transport. The distinction between the two is, however, purely theoretical; the time-averaged mass concentration of particles decreases rapidly away from the immobile bed without a significant break, as the two support mechanisms merge. Further, since wave-induced, reversing oscillatory motions are the main entraining force, both suspended and bedload reveal strongly episodic behaviour. In the swash zone, the thin flows make the relative role of bedload and suspended load even more difficult to define. The variable directions of wave approach result in reversals of the direction of transport by both the longshore currents and the swash; thus, the *gross* sediment transport alongshore may be very large, while the *net* transport may be significantly smaller. However, transport rates up to several million $m^3 a^{-1}$ are common and this transport has been studied extensively by geomorphologists, coastal oceanographers and engineers concerned with coastal forms and dynamics, harbour siltation and dredging, the effectiveness of groynes as shore protection, etc.

Longshore sediment transport rates

Longshore transport is driven fundamentally by the alongshore component of the *wave momentum flux* or *radiation stress* (Longuet-Higgins and Stewart 1964), which itself creates a quasi-steady longshore current, although the complex interaction between the waves, the currents and the sediment is far from understood. Transport takes place within a combined *wave-current boundary layer*. The wave boundary layer is relatively thin compared to that of the current; consequently the stresses generated by the waves are significantly larger. However, to a first order, waves are purely oscillatory and thus cannot induce significant transport. It was this assumption, that waves entrain (initiate motion) and currents advect sediment, that led to the most common formulation for sediment transport alongshore (Bagnold 1963).

Longshore transport models

The Energetics Model: Inman and Bagnold (1963) defined longshore transport rates as a simple function of the alongshore component of the incident wave energy flux:

$$I_l = (\rho_s - \rho_f)\, ga' Q_l = K\, P_L$$
$$= K\, (E\, C\, n)_b \sin \alpha_b \cos \alpha_b$$

where I_l is immersed weight transport rate, a' is a constant of 0.6, Q_l is longshore transport rate, ρ_s and ρ_f are solid and fluid mass densities, K is a dimensionless proportionality coefficient, E is the specific wave energy density, C is the wave celerity or speed, n is the ratio of wave group velocity to wave phase velocity, α is the angle of wave breaking and the subscript b refers to conditions at the point of wave breaking. The constant K, which was originally proposed as an efficiency factor ($\approx$0.77 when the root-mean-square wave height is used in the energy calculation; Komar 1971), has subsequently been the source of considerable debate. This proportionality factor is also thought to be dependent upon the grain size, wave steepness, the surf similarity parameter or Irribarren Number, bed slope, etc. The 'constant' K was originally conceived as applying to fully developed transport in an instantaneous sense. A number of authors have used a time-averaged relationship to predict the potential for a time-averaged rate of transport and thus a model for the long-term potential for erosion and deposition (see Greenwood and McGillivray 1980). Komar (1998) gives an extensive review of the energetics model.

Bailard (1981), following Bagnold (1966) and Bowen (1980), expanded the basic model and showed that the total longshore transport rate depends upon the steady current and a number of higher order moments of the velocity field:

$$\langle i_y \rangle = \rho\, c_f\, u_m^3 \left\{ \left[\frac{\varepsilon_b}{\tan \phi} \right] \left(\delta_v^3 + \frac{\delta_v}{2} + \frac{\tan \beta}{\tan \phi} (u3)^* \tan \alpha \right) \right.$$
$$\left. + \frac{u_m}{w} \varepsilon_s \delta_v (u3)^* + \frac{u_m^2}{w^2} \varepsilon_s^2 \tan \beta\, (u5)^* \tan \alpha \right\}$$

where c_f is a drag coefficient, u_m is oscillatory velocity magnitude, ϕ is angle of internal friction, α is angle of wave approach, ε_b is bedload efficiency, ε_s is suspended load efficiency, w is sediment fall velocity and the relatively steady currents δ, δ_u and δ_v are defined as:

$$\delta = \frac{\bar{u}}{u_m}\ ;\ \delta_u = \frac{\bar{u}}{u_m} \cos \theta\ ;\ \delta_v = \frac{\bar{u}}{u_m} \sin \theta$$

and the velocity moments ψ_1 and ψ_2 are defined as:

$$\psi_1 = \frac{\langle \tilde{u}^3 \rangle}{u_m^3}\ ;\ \psi_2 = \frac{\langle |\bar{u}|^3 \tilde{u} \rangle}{u_m^4}$$

and the integrals $(u3)^*$ and $(u5)^*$ contain the time-averaged magnitudes of the combined velocities associated with each of the higher moments and are defined as:

$$(u3)^* = \frac{1}{T}\int_0^T (\delta^2 + 2\delta \cos(\theta-\alpha)\cos \sigma t + cos^2\sigma t)^{\frac{3}{2}}\, dt$$

$$(u5)^* = \frac{1}{T}\int_0^T (\delta^2 + 2\delta\, cos(\theta-\alpha)\cos \sigma t + \cos^2 \sigma t)^{\frac{5}{2}}\, dt$$

where T is the wave period, θ is steady current angle and σ is wave frequency.

The Applied Stress Model uses the concept of 'excess stress' for alongshore transport, i.e. the stress in excess of that used to initiate motion of the sediment. The models use precise physical relationships coupled with semi-empirical expressions to describe the transport process. A typical model of the longshore component of transport, q_l, is that of Grant and Madsen (1979):

$$q_l(t) = 40wD\left(\frac{\frac{1}{2}\rho c_f\left(u^2(t)+v^2(t)\right)}{(s-1)\rho g D}\right)^3 \frac{v(t)}{\sqrt{u^2(t)+v^2(t)}}$$

where t is time, w is mean sediment fall velocity, D is sediment diameter, ρ is fluid density, c_f is coefficient of bed friction, u and v are horizontal cross-shore and longshore fluid velocities, s is specific gravity and g is gravitational constant.

Most of the models have been calibrated using sand-sized material, but clearly longshore transport also includes coarser grained materials. Van Wellen *et al*. (2000) reviews the longshore transport equations for coarse-grained beaches.

Geomorphological significance

Littoral drift is an important part of the *sediment budget*, which is based on mass conservation principles applied to the coastal zone. A major concept in coastal geomorphology is that of the littoral cell, consisting of a zone of sediment supply (may be rivers or shore erosion), a zone of erosion (net loss of sediment), a zone of longshore transport and a zone of accretion (net gain of sediment). An excellent review of the concept is available in Carter (1988). *Gradients* in the rates of longshore sediment transport can be used to model major aspects of shoreline erosion and/or accretion (see Komar 1998).

Several specific geomorphological forms owe their origin to longshore sediment transport. For example, a TOMBOLO is a strip of sediment accumulating between an offshore island and the main shoreline formed as a result of wave refraction around the island. Refraction produces waves, which approach from opposing directions in the lee of the island and induce currents and longshore transport, which converge from two directions. Spectacular *sand spits* may develop as a result of littoral drift being deposited at a re-entrant in the shoreline or across embayments and may stretch for several kilometres. Refraction of waves around the down drift terminus often produce *recurved* or *hooked* spits.

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BRIAN GREENWOOD

LUNETTE

Transverse and roughly concentric aeolian accumulations (the *bourrelets* of some French workers) that occur on the downwind margins of PANs. Although they had been described before, they were named as such in Australia by Hills (1940), though the basis of his etymology is unclear. They tend to occur in areas where present-day precipitation levels are between about 100 and 700 mm, but their stratigraphy can give a good indication of past changes in climate and hydrological conditions. Some basins may have two or more lunettes on their lee sides and these may have different grain size and mineralogical characteristics. Lunettes may be some kilometres long and in exceptional circumstances may attain heights in excess of 60 m.

Good regional descriptions of lunettes are provided for the High Plains of the USA by Holliday (1997), for Tunisia by Perthuisot and Jauzein (1975), for the Kalahari by Lancaster (1978) and for the Pampas of Argentina by Dangavs (1979).

The materials that make up lunettes can vary from clay-sized material (which in the case of clay dunes (Bowler 1973) can make up 30 to 70 per cent of the total) through to sand-sized material. Equally some lunettes are carbonate-rich (Goudie and Thomas 1986) whereas others are almost pure quartz. Lunettes may also contain appreciable quantities of evaporite minerals derived from the basins to their windward.

Various hypotheses have been put forward to explain lunette composition. Hills (1939) believed that the lunettes were built up when the pans contained water and that they were composed of atmospheric dust captured by spray droplets derived from the water body. Stephens and Cocker (1946) pointed out that this could not account for those lunettes that were not predominantly silty. They also suggested that many of the lunettes were built up of aggregates transported from the floors of pans. Campbell (1968) believed that this deflation hypothesis could indeed account for many lunette features. As she remarked (p. 104) 'the close similarity between the composition of the lunette and its associated lake bed suggested that the two are causally related, i.e. that the material in the lunette was derived from the lake bed'. However, she also recognized that some of the material could be derived from wave-generated beaches and so could be analogous to primary coastal foredunes.

This was a view that was developed by Bowler (1973), who saw sandy facies as being associated with a beach provenance (at times of relatively high water levels) whereas clay-rich facies, which also may have a high content of evaporite grains, formed during drier phases when deflation of the desiccated lake floor was possible. Lunettes can therefore provide evidence for understanding past hydrological changes (Page *et al*. 1994).

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A.S. GOUDIE

M

MAGNITUDE–FREQUENCY CONCEPT

When the activity of different geomorphic processes is compared on a given timescale some processes appear to operate continuously while others operate only when specific conditions occur (referred to as events). The term eposidicity refers to the tendency of processes to exhibit discontinuous behaviour and occur sporadically as a series of individual events. Episodicity occurs when discontinuity is inherent in the forcing process (e.g. discontinuous rainfall produces episodic gully erosion). It may also occur if the relationship between the forcing process and the geomorphic response is not constant (e.g. the strain resulting from continuous crustal plate convergence may be manifest as either continuous deformation or episodic earthquake activity and coseismic uplift). The point at which the resistance to stress imposed by a forcing process is overcome is marked by a discernible geomorphic response (event) and is referred to as a threshold (see THRESHOLD, GEOMORPHIC). Because of hysteresis effects, the initiating threshold for a geomorphic event may be of a different magnitude to the terminating threshold.

Like many concepts in geomorphology, interpretations of eposidicity are scale dependent. Even processes that are sometimes considered continuous can be interpreted as episodic on different timescales. For example, SOIL CREEP in a given area is commonly portrayed as continuous and ubiquitous in terms of landform evolution. However, on a diurnal or seasonal scale, some forms of soil creep are evidently episodic, operating only when certain temperature or moisture conditions are met. For a given process, large events involving high amounts of concentrated energy (high magnitude) are rare (low frequency) and small events are common and tend to be less episodic.

Historically geomorphology has gone through a number of major paradigm shifts involving the eposidicity and the effectiveness of processes. The earliest interpretations of how landforms were created were based on catastrophic formative supernatural or natural events unrelated to the work of contemporary processes. As the science of geomorphology developed, increasing recognition of the age of the Earth allowed for the possibility that slow acting, contemporary processes, with sufficient time, could account for the development of much of the form evident in the landscape. Paradoxically the ability to date accurately key deposits revealed that certain landforms had been formed under specific and unusual circumstances. This gave rise to the prospect of NEOCATASTROPHISM. The consequential question that heightened interest in the study of magnitude–frequency relations was: were the effects of a rare catastrophic event 'permanent' or, given time, were they overwhelmed by 'normal', small but constantly operating processes responsible for producing a characteristic form? The prospect of identifying a 'characteristic' form in a geomorphic system however requires stability in boundary conditions such as climate, tectonic activity and vegetation cover and thus the existence of some form of DYNAMIC EQUILIBRIUM. Clearly the shorter the period of observation the more likely it is that these conditions pertain.

From the point of view of landform development, the RELAXATION TIME associated with a geomorphic event may be a more appropriate parameter than magnitude for identifying the formative or geomorphically effective event. The relaxation time is the length of time over which the effects of an event can be discerned in the landscape. Although this is clearly a function of event magnitude it also incorporates the influence of terrain resilience, LANDSCAPE SENSITIVITY and ambient conditions (Crozier 1999). The degree of equilibrium and the prospect of identifying characteristic form can then be determined by the 'transient form ratio' (Brundsen and Thornes 1979) which is the ratio of relaxation time of a geomorphic event to its recurrence interval (frequency). Values equal or greater than one indicate a constantly changing system while values less than one represent some form of dynamic equilibrium.

Many studies on this question, including the seminal work by Wolman and Miller (1960) have drawn conclusions based only on the period of instrumental record and the extrapolation of such results to landform evolution must be done with care (Wolman and Gerson 1978). In Wolman and Miller's original study it was shown that most of the work of sediment transportation in rivers was carried out by moderate flow events of a magnitude that recurs on average at least once every five years. The analytical approach leading to this conclusion is illustrated in Figure 105. Curve (a) indicates that the rate of sediment movement is a power function of applied stress, or in this case a surrogate such as discharge. Curve (b) shows a log normal frequency–magnitude distribution of measured flow events. Curve (a × b) is the product of frequency and rate of movement which attains a maximum work value which can be related back to the discharge axis at (x) to identify the magnitude of the event performing the most work. The frequency of this event (y) can be determined with reference to the frequency–magnitude distribution curve. Wolman and Miller's work has been extended to a number of different processes and environments, in some cases producing conclusions at variance with the original findings. Selby (1974), for example, argues that in high-energy hillslope environments and the headwaters of drainage basins, high magnitude landslide events occurring at frequencies less than once every five years dominate geomorphic activity on slopes and low-order stream channels. Similar conclusions have been made by a comprehensive comparison of fluvial transport and mass movement in different parts of the drainage basins (Trustrum *et al.* 1999).

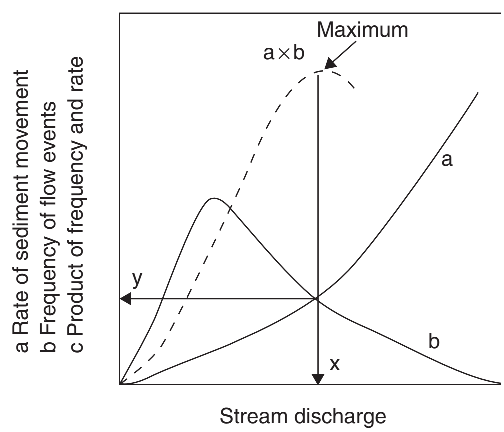

Figure 105 Relations between rate of sediment transport, discharge and the frequency–magnitude distribution of discharge events, c = a × b (after Wolman and Miller 1960)

In strictly geomorphic terms, the magnitude of an event generally refers to the amount of work carried out (e.g. mass of sediment transported) or the degree of landform change experienced (geomorphic response). However, because of difficulty in directly measuring geomorphic events many magnitude–frequency studies are approached indirectly. This involves characterizing events from the behaviour of the forcing agent, rather than from the geomorphic event itself. However, the indirect approach implies a known relationship between geomorphic work and the behaviour of the forcing agent. For example, if this relationship and the initiating thresholds are known, the frequency and magnitude of soil erosion and landsliding can be determined from the rainfall record, aeolian transport from wind speed, coastal changes from the wave regime and fluvial transport from the streamflow regime.

A major problem with the indirect approach is that relationships between the forcing process and geomorphic work are not always close or temporally or spatially stable. For example, stream discharge may not relate closely to sediment transport in a supply-constrained system. Indeed in some streams, sediment load may be related to stream power at certain times and to rates of hillslope sediment supply at others. The nature of sediment control can also change throughout the catchment to the extent that sediment load in steep upper catchments may relate to the magnitude and frequency of landslide events. Furthermore, the parameter used to represent magnitude also needs to be carefully chosen if it is to reflect accurately and consistently the energy of the forcing process. In the case of landsliding for example, the groundwater content of a hillslope is directly responsible for initiating movement, however lack of adequate databases dictates that magnitude–frequency analysis is often carried out using rainfall values; a more appropriate analysis would involve the magnitude and frequency of groundwater levels. The use of arbitrarily defined events such as daily rainfall as opposed to event rainfall introduces another level of inaccuracy into the analysis, as important parameters such as actual intensity and duration are obscured. Besides magnitude and frequency, duration is another parameter of the forcing processes that influences geomorphic response. Clearly, in power-constrained systems, high effective streamflow, maintained for long periods of time, will accomplish more work than the same levels of flow occurring as short duration events. Similarly, studies of the relationship between the magnitude and intensity of earthquakes and the degree of land deformation have shown that the duration of shaking is important in evoking a geomorphic response.

For a given process, average event frequencies can be established from the number of times an event initiating threshold is exceeded or the number of times the threshold is exceeded by a specified magnitude, in a unit of time. More commonly, following the practice of flood frequency analysis, frequencies are expressed as recurrence intervals (return periods) determined by the relationship of the length of record to the ranking or relative magnitude of the event in a series of events. A range of statistical models for extreme value probability distributions can be used for depicting the declining exponential function between magnitude and frequency. Little confidence can be placed on derived event magnitudes with recurrence intervals greater than the length of record.

An important property of recurrence intervals (sometimes overlooked when regional comparisons are made) is that they are generally a function of the size of the area from which they are derived. In other words, the larger the area, all other things being equal, the more likely it will be to experience a specified event and therefore the shorter the derived recurrence interval. Another point of caution is that the statistical derivation of frequency may obscure the clustering of events, which is a behavioural property that can be significant in generating a geomorphic response. Because variation of frequency through time may signal significant environmental change the identification of clustering is an important aspect of magnitude–frequency analysis.

For some processes, magnitude and frequency of occurrence in space (Innes 1985; Hovius *et al.* 1997) mirrors temporal magnitude–frequency relationships. In the case of landslides for example, episodes producing numerous landslides on the one occasion show that large landslides are greatly outnumbered by smaller landslides. These spatial magnitude–frequency relationships are used in hazard assessments as analogues for temporal magnitude–frequency relationships. Establishing magnitude–frequency relationships from landscape evidence needs to take account of the longevity of landscape evidence; clearly the signature of large landslides persists longer than that of smaller landslides.

The magnitude–frequency concept and the analysis it has provoked have been important in informing the discipline about the variability and behaviour of geomorphic processes. It reminds us that within one lifetime we are unlikely to experience the whole range in magnitudes that a particular process is capable of generating. The concept provides a rationale for extrapolating short-term measurements over longer periods, as a way of assessing the long-term rates of geomorphic processes. However changes in boundary conditions limit the extent to which the relationships can be extrapolated. Magnitude–frequency analysis also provides a method for statistically identifying the key events in terms of work and landform response, thereby providing a key variable for characterizing geomorphic systems and predicting other system characteristics. Finally, from a pragmatic point of view, the concept enables the identification of a design or planning event for use in engineering decisions and hazard management.

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MICHAEL J. CROZIER

MANAGED RETREAT

Usually defined as the backward realignment of coastal defences so as to allow the reformation of SALTMARSH or mud flat (see MUD FLAT AND MUDDY COAST) on the seaward side. It may also, rarely, be used where other coastal features such as dunes are expected to form in front of a retreating line of defence. The term has now largely been superseded by the preferred phrase '*managed realignment*' which is effectively synonymous but avoids the negative implications to many people of the word 'retreat'.

In low-lying tidally dominated coastal areas, especially in northern Europe and North America, large areas of saltmarsh have been 'reclaimed' by the building of sea walls for agricultural use or for urban and industrial development. Such land claim alters the dynamics of the local tidal flow and sediment budget and may lead to localized increase in tidal energy and to EROSION. Reclaimed areas are often in estuaries; the natural funnel shape of an ESTUARY is transformed to a narrower more parallel one providing less space for the dissipation of tidal energy. The tidal flow tends to move further upriver, the river channel deepens and widens and the mudflat/saltmarsh complex migrates also, maintaining its relation to the local energy gradient (Pethick 2000). Although seawalls were often originally positioned so as to leave in front some marsh and mudflat, these often become eroded away. High tides come up to the sea defences and the threat of flooding at spring tides and during storms increases; the walls themselves may be destroyed. If also SEA LEVELS are rising, the effect is magnified.

Appropriate responses to the threat depend on the type of infrastructure in the hinterland. Raising and strengthening seawalls, while expensive, may be thought the best option if valuable built-up or agricultural land is protected. Where the hinterland consists of low value agricultural land, managed retreat is seen as the best solution to the erosion problem. Reinstatement of the coastal marshes should provide a dispersal area for the tidal flow, a supply of sediment for local coastal processes, and a buffer zone for dispersion of wave and tide energy. Given time and space, it is hoped that the coastline will develop its own contours and maintain itself, altering to suit changing conditions without the need for expensive engineering works.

However, breaching or removal of seawalls does not always lead to the reestablishment of saltmarsh. Some low-level sites inundated either through storm damage or by the deliberate realignment of walls have remained as largely unvegetated mudflats (French *et al*. 2000) over a period of as much as fifty years. While popular with environmentalists and relatively inexpensive compared with hard engineering solutions, managed realignment is less attractive to the owners of the land to be sacrificed.

The success of realignment programmes has been shown to depend on several factors. It appears only to be successful on sites where saltmarsh once existed and where marshes survive nearby as sources of suitable plant propagules. There should be traces of the original saltmarsh creek system (though a channel system may initially be artificially created). The site should be sufficiently high that it is only inundated 400 to 450 times annually – in practice this means above 2.1 m OD or only mud flat is likely to form. It should preferably be neither completely flat nor steeply sloping but with a gentle land to sea slope (Burd 1995).

At least in the EU, managed realignment schemes can attract funding under the saltmarsh option of the Countryside Stewardship scheme (which replaced the Habitat Scheme in January 2000). Detailed Coastal Habitat Management Plans need to be drawn up in all cases before action is taken.

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PIA WINDLAND

MANGROVE SWAMP

Mangroves are trees or shrubs which grow in sheltered, or low-energy, upper intertidal zones in the tropics and subtropics, replacing SALTMARSH which is found along muddy temperate shorelines. There are more than fifty species of mangroves occurring in two distinct provinces, the Indo-west Pacific which contains the greatest diversity of species, and the West Indian region with far fewer species. In the West Indies, three species of mangrove occur, each tending to occupy a discrete location, *Rhizophora mangle* to seaward, *Avicennia germinans* in more landward locations, and *Laguncularia racemosa* mixed with the other species, or in areas that have been disturbed. In the Indo-west Pacific province there are more species of mangroves, with up to thirty species in the most diverse locations. There is generally a seaward zone, composed of species of *Avicennia* or *Sonneratia*, an intermediate zone dominated by species of *Rhizophora* and *Bruguiera*, and more landward zones in which *Ceriops* occurs with genera such as *Lumnitzera, Heritiera* and stunted forms of *Avicennia*. In areas of heavy rainfall mangrove forests merge into tropical forests, whereas in arid regions they often contain a hypersaline mudflat and are flanked by bare or samphire-covered supratidal flats.

Mangroves can grow in freshwater, but they appear to have a competitive advantage over other vegetation when the substrate is saline. Mangroves perform best at salinities that are less than that of seawater, but can survive in salinities of up to 90 parts per thousand. They show a series of adaptations to saline sediments. The root systems of mangroves contain breathing cells called lenticels, and are often distinctive, enabling the plants to survive in upper intertidal habitats which are subject to frequent inundation and are composed of muds or sandy muds which are anaerobic. The enormous prop root systems that characterize *Rhizophora* (Plate 75) are almost impenetrable. *Avicennia* and *Sonneratia* have pencil-like roots termed pneumatophores, whereas other species have knee or buttress roots. In addition many species of mangroves produce viviparous seedlings, the fruit already producing a well-developed root before falling from the tree.

These adaptations to the inhospitable environment in which mangroves live, and the zonation of species which is apparent on many shorelines, led to an early interpretation that mangroves undergo a succession of species culminating in a terrestrial climax vegetation. The root systems slow the movement of water and may trap sediment promoting accretion along the shore. There have been various attempts to measure sedimentation rates beneath mangrove forests. Direct measurements have been made by using systems

Plate 75 The interior of a mangrove swamp in the West Indies. The mangrove *Rhizophora mangle* has a network of prop roots that extend two or more metres up the trunks of the trees. These root systems can substantially reduce the flow of water across the substrate encouraging deposition of mud where suspended in the water column. In other cases, such as the forest shown here, the fibrous roots themselves contribute to a mangrove peat substrate

of stakes throughout the mangroves, or placing a marker layer. These approaches have generally met with less success than in saltmarshes because of the active bioturbation of the muds by a rich and diverse fauna, including crabs and mud skippers. Bioturbation also limits the efficacy of determining sedimentation rates using ^{210}Pb or ^{137}Cs isotopes, but rates of up to $3\,mm\,yr^{-1}$ are indicated (Lynch *et al.* 1989). Longer term sedimentation rates have been determined using radiocarbon dating and indicate that rates of up to $6–8\,mm\,yr^{-1}$ can occur. However, the short-term and long-term estimates may not be directly comparable because it is often difficult to discriminate whether root material was formed close to, or at some depth below, the surface. It is also difficult to take compaction into account, or to allow for the variation in sedimentation rate that is likely across the intertidal zone.

Zonation of mangrove species, even if it does occur, need not indicate a temporal succession that involves replacement of successive zones through time. Patterning of species may be a static equilibrium in relation to environmental gradients in habitat factors, such as salinity or water-logging (Lugo 1980). It has proved useful to distinguish various geomorphologically defined habitats within which mangroves grow (Thom *et al.* 1975; Semeniuk, 1985).

There is considerable variation both in the mangrove species that are found, and the elevation at which they grow, up estuaries. Particularly extensive mangrove forests occur in the abandoned parts of deltas where tidal processes are important, such as the Sundarbans west of the mouth of the Ganges–Brahmaputra rivers. Where mangrove swamps are associated with the mouths of large rivers, the substrate on which they are established is generally composed of terrigenous sediments washed from the catchments. The patterning of mangrove species in such deltaic environments reflects an ever-changing series of geomorphological habitats in which mangroves respond opportunistically to habitat change induced by geomorphological processes, such as channel migration and avulsion (Thom 1967).

Mangroves also occur in reef environments where their extent is often a function of stage of reef evolution (Stoddart 1980). In these settings, mangroves either establish over calcareous sediments formed from reef organisms, or they develop over mangrove peat which is derived from the roots of the mangroves themselves. Mangrove islands within the Belize barrier reef, termed mangrove ranges (Plate 76), are complex. Although these islands are dominated by tidal overwash, the patterning of species is not straightforward, with *Rhizophora* adopting a dwarfed scrub form in some locations, and elsewhere reaching a woodland of several metres tall, intergrading into areas of more open *Avicennia* woodland. In places the mangrove forests are dissected by sinuous creeks, and elsewhere there are bare areas which may result from storm damage, but where soil chemistry presently appears to prevent recolonization by mangroves.

It is clear that mangrove swamps have altered in extent considerably. The stratigraphy of mangrove-dominated coasts records the Holocene evolution of coastal environments and there have been significant adjustments as a result of changes in sea level. Throughout much of the Indo-west Pacific region, where sea level achieved a level close to present around 6,000 years ago and has been relatively stable or fallen slightly since, former mangrove sediments often underlie extensive Holocene plains upon which freshwater wetlands or peat swamp forest have established (Woodroffe *et al.* 1985). Former mangrove sediments often represent potential acid-sulphate soils in which pyrite can become oxidized, resulting in extremely acidic waters if drained or excavated. In the West Indies and the Everglades of Florida, where the history of relative SEA LEVEL change through the Holocene appears to have

Plate 76 A mangrove island, called a mangrove range, on the Belize barrier reef. Although dominated by two mangrove species, *Rhizophora mangle* and *Avicennia germinans*, there is a complex vegetation pattern with areas that are bare of vegetation and a network of creeks that dissect the island

been characterized by rise to present, intertidal mangrove peat overlies previously terrestrial environments, such as freshwater sedge peat.

These productive and protective forests are also likely to undergo changes in distribution as a result of sea-level change in the future (Woodroffe 1990). Although this has attracted considerable attention, it is the case that human impact, clearing forests for other land use, or for timber, and most recently for shrimp farming, have generally already had far-reaching impacts. Mangroves are subject to various natural disturbances, for instance storm impact. However, they play an important role as storm protection, especially where surges are experienced as in the Bay of Bengal.

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COLIN WOODROFFE

MANTLE PLUME

'A blob of relatively hot, low-density mantle that rises because of its buoyancy' (Condie 2001: 1). Their existence was suggested by J.T. Wilson (1963) who sought to explain the progressive change in age and form of islands along oceanic chains, such as the Hawaiian–Emperor Chain of the Pacific. He proposed that as a lithospheric plate moves across a fixed hotspot (the mantle plume), volcanic activity is recorded as a linear array of volcanic seamounts and islands parallel to the direction in which the plate moves (Plate 77).

As rising mantle plumes reach the base of the lithosphere, they spread laterally to produce plume heads which may have diameters that reach 500 to 3,000 km. The tails of the plumes are only 100–200 km in diameter. The surface manifestations of plumes are large hotspots, which are zones with active volcanism. Those plumes that are 1,500–3,000 km in diameter are termed superplumes (Condie 2001: 2).

Mantle plumes are of fundamental geomorphological importance. First, their associated hotspots help to explain the distribution of volcanic activity in intraplate situations. They also explain the development of volcanic chains on the Pacific plate and elsewhere, and also the development of some seamounts. Second, broad zones of uplift (swells) are sometimes associated with mantle plumes. These can create large domes that then dominate drainage patterns at a regional and subcontinental scale. Cox (1989) argued that the drainage directions and patterns of India, eastern South America and southern Africa were good examples of this effect (Figure 106). Third, many large igneous provinces (LIPs) have a mantle

Plate 77 The Spitzkoppje in central Namibia is a granite mass that was emplaced in the early Cretaceous as a result of magmatic activity associated with the presence of a mantle plume that contributed to the rifting of Gondwanaland. The hotspot is now located in the Southern Ocean in the vicinity of Gough and Tristan da Cunha

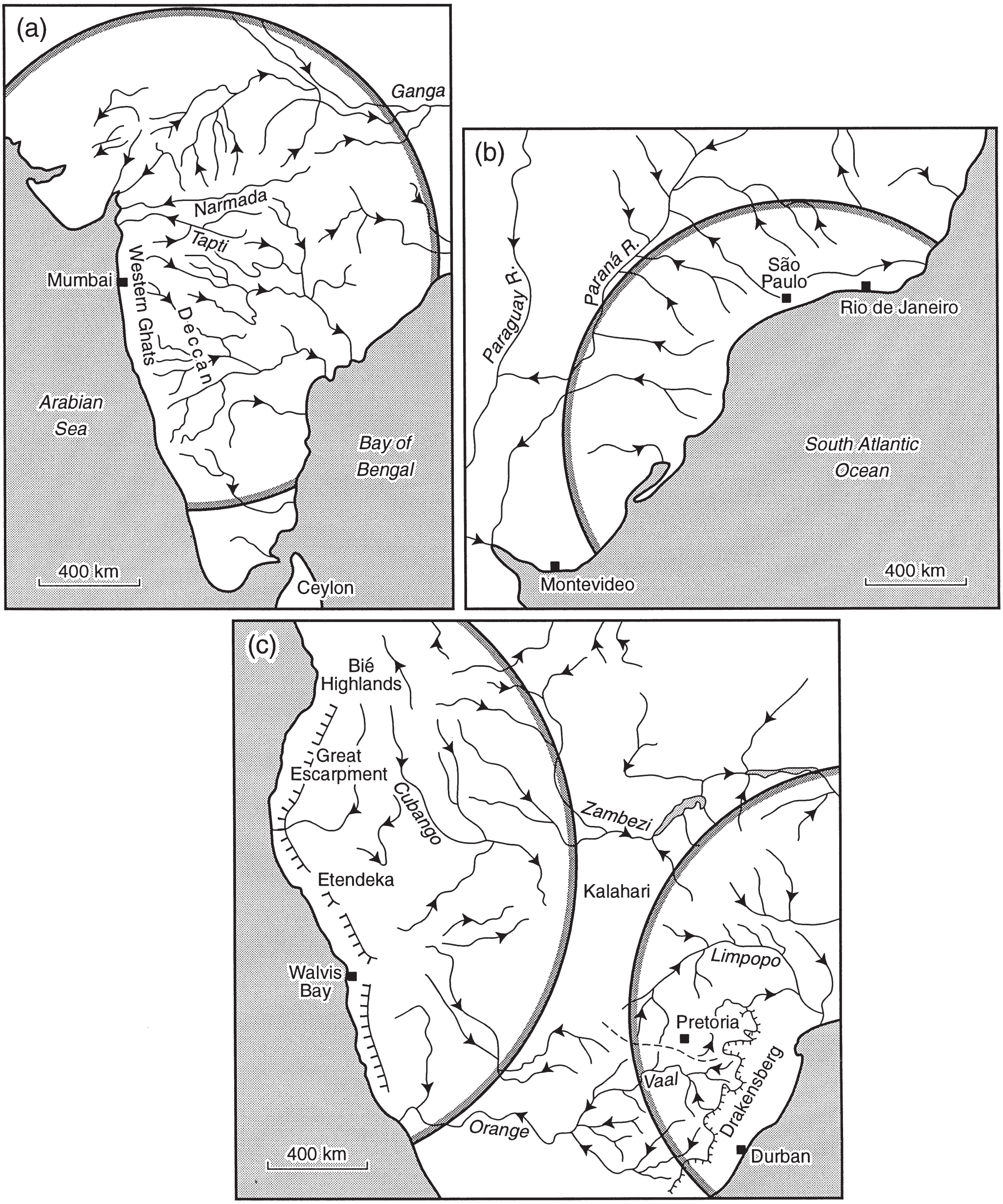

Figure 106 Postulated locations of major domes associated with mantle plumes and showing their relationship to drainage patterns (modified after Cox 1989)

plume origin, and phenomena associated with them include such features as continental flood basalts (e.g. in the Deccan of India), giant dyke and sill swarms (Ernst *et al.* 1995), and large layered intrusions (e.g. the Bushveld Igneous Complex in South Africa). Fourth, mantle plumes may play a major role in the breakup of supercontinents (e.g. Gondwanaland), the formation of passive margins like those in Namibia (Goudie and Eckardt 1999), and the development of basins (e.g. Red Sea, Gulf of Aden, etc.).

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A.S. GOUDIE

MASS BALANCE OF GLACIERS

The mass balance of a glacier is the sum of all processes which add mass to a glacier and remove mass from it. Accumulation or addition of mass most commonly takes place in the form of snowfall, often modified by wind and avalanches (see AVALANCHE, SNOW). Melting of snow and ice is the predominant form of ablation or removal of mass, but calving of tidal glaciers, ice avalanching at steep hanging glaciers or evacuation of windblown snow in dry areas can locally be of high relative importance. The resulting loss or gain of mass is the direct, undelayed response of a glacier to climatic conditions and must be considered to be a key indicator of climate change (IPCC 2001; Haeberli *et al*. 2002).

Techniques applied to determine glacier mass balance include (a) the direct glaciological method (snowpits and ablation stakes), the geodetic/photogrammetric method (repeated precision mapping), the hydrological method (difference between measured precipitation minus evaporation and runoff) and index methods (snowline mapping, etc.). Long-term mass changes can also be inferred from cumulative glacier length changes using continuity approaches or flow models (Haeberli and Hoelzle 1995; Oerlemans *et al*. 1998).

In humid maritime regions large amounts of ablation are required to compensate for heavy snowfall: the equilibrium line which separates accumulation from ablation areas on glaciers remains at altitudes with relatively warm air temperatures, enabling intense sensible heat flux and strong ice melt during extended ablation seasons. Temperate glaciers at melting temperature, exhibiting high mass turnover and rapid flow, dominate these landscapes. The lower parts of such temperate glaciers commonly extend into grassland and forested valleys where summer warmth and winter snow accumulation prevent the development of PERMAFROST. Ice caps and valley glaciers of Patagonia and Iceland, the western Cordillera of North America and the coastal mountain chains of New Zealand and Norway are features of this type. In contrast, dry continental conditions such as exist in northern Alaska, arctic Canada, subarctic Russia, parts of the Andes near the Atacama desert or in many central Asian mountain chains, force the EQUILIBRIUM LINE OF GLACIERS to elevations with cold air temperatures, short ablation seasons, reduced sensible heat flux and limited amounts of ice melting. In such regions, polythermal or cold glaciers, lying far beyond the treeline and often even beyond tundra vegetation, have a low mass turnover, less rapid flow and are associated with severe periglacial conditions and permafrost (Shumskii 1964).

The goals of long-term mass balance observations are (1) to determine annual ice loss/gain as a regional signal and (2) to better understand processes of energy and mass exchange. A change (δ) in equilibrium line altitude (ELA) induces an immediate change in specific mass balance (b = total mass change divided by glacier area). The resulting change in specific mass balance (δb) is the product of the shift in equilibrium line altitude (δELA) and the gradient of mass balance with altitude (db/dH) as weighed by the distribution of glacier surface area with altitude (hypsometry). The hypsometry represents the local/individual or topographic part of the glacier sensitivity whereas the mass balance gradient mainly reflects the regional or climatic part (Kuhn 1990). As the mass balance gradient tends to increase with increasing humidity and mass turnover (Kuhn 1981), the sensitivity of glacier mass balance with respect to changes in equilibrium line altitude is generally much higher in areas with humid/maritime than with dry/continental climatic conditions (Oerlemans 1993). Rising snowlines and cumulative mass losses lead to changes in average albedo and continued surface lowering. Such effects cause pronounced positive feedbacks with respect to radiative and sensible heat fluxes. In areas of cold firn, atmospheric warming first induces firn warming; mass loss only sets in when the firn has reached melting temperatures and water can leave the system instead of refreezing.

A Global Terrestrial Network for Glaciers as part of the climate-related Global Terrestrial Observing System (GTOS/GCOS) is operated by the World Glacier Monitoring Service (WGMS) which co-ordinates worldwide compilation and dissemination of standardized data on glacier fluctuations. Mass balance measurements are reported in a biennial bulletin (IAHS(ICSI)/UNEP/UNESCO/WMO 2001; http://www.geo.unizh.ch/wgms/). Overall ice loss appears to be strong and probably even accelerating. Anthropogenic greenhouse forcing could have started to exert a predominant influence on this development and may lead to complete deglaciation of many mountain regions of the world within decades (Dyurgerov and Meier 1997a,b; Haeberli *et al*. 2002).

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WILFRIED HAEBERLI

MASS MOVEMENT

A mass movement is the downward and outward movement of slope-forming material under the influence of gravity. The process does not require a transporting medium such as water, air or ice. The term LANDSLIDE often is used as synonymous for mass movement phenomena. However, in a pure sense the term landslide is used as a generic term describing those downward movements of slope-forming material as a result of shear failure occurring along a well-defined shear plane.

There are numerous classifications of mass movements. Most are based on morphology, mechanism, type of material and rate of movement (e.g. Varnes 1978; Hutchinson 1988; Cruden and Varnes 1996). The classification and description used in this entry was developed by a European research group (Dikau *et al*. 1996) (Table 27). The terminology is based upon the classifications of the International Geotechnical Societies' UNESCO Working Party on World Landslide Inventory (WP/WLI) (UNESCO 1993).

Fall

Alternative terms: rockfall, stone fall, pebble fall, boulder fall, debris fall, soil fall.

A fall is a free movement of material from steep slopes. Different types of falls are described by material and failure processes. The term rockfall is often used as the general term without further reference to the material involved. Whalley (1984) and Flageollet and Weber (1996) provide summaries.

Falls occur in various sites such as coastal cliffs, steep riverbanks, edges of plateau, mountain faces or escarpments. They may also occur on artificial embankments (road outcrops). Joints and faults produce planar sheets to form wedge-shaped hollows and boundary vertical joints. Falls can form a fan-shaped cone at the base of the slope. These

Table 27 Classification of slope movements

Type	Rock	Debris	Soil
Fall	Rockfall	Debris fall	Soil fall
Topple	Rock topple	Debris topple	Soil topple
Slide (rotational)	Single (slump) Multiple Successive	Single Multiple Successive	Single Multiple Successive
Slide (translational)	Block slide Rock slide	Block slide Debris slide	Slab slide Mudslide
Lateral spreading	Rock spreading	Debris spread	Soil (debris) spreading
Flow	Rock flow (sackung)	Debris flow	Soil flow
Complex (with runout or change of behaviour downslope, note that nearly all forms develop complex behaviour)	e.g. rock avalanche	e.g. flow slide	e.g. slump-earthflow

Source: Dikau *et al*. (1996)
Note: A compound landslide consists of more than one type e.g. rotational-translational slide. This should be distinguished from a complex landslide where one form of failure develops into a second form of movement i.e. a change of behaviour downslope by the same material

TALUS slopes should be distinguished from accumulations arising from a large fall, a complex rock avalanche (STURZSTROM), which result in an accumulation of debris of all sizes that can block a river, leading to flooding hazards.

Falls are influenced by slope aspect and angle, size and shape of jointed rocks, strike angle, status and deformation of rocks and vegetation cover. Debris and soil falls originate in material which has already been detached from the bedrock. In solid rock the separation process may take time and it arises from internal and external factors which are often combined.

The main implication of fall processes for a planner is to ensure that there are suitable surveys of those areas which are likely to produce rockfalls (Table 27). The main product will be a planning hazard zoning map and suitable monitoring systems. Warning systems are also in operation on various other sites with this type of hazard.

Topple

Alternative terms: rock topple, debris topple, soil topple, tilting blocks.

A topple consists of a forward rotation of a mass of rock, debris or soil about a pivot or hinge on a hillslope. The toppling may culminate in an abrupt falling or sliding, but the form of movement is tilting without collapse. Goodman and Bray (1976) and Dikau *et al*. (1996) provide summaries.

There are several processes responsible for toppling failure such as progressive weathering or erosion resulting in a weakening or loss of elastic underlying material, swelling and shrinking of clay-rich material because of soil moisture changes, and deepening or undercutting of slopes by erosion providing sufficient stress to cause unloading decompression.

The primary driving force for topple failure is the detachment of a column so that the load is transferred to a narrow base of weaker rock. Slope height is an important controlling parameter as is the width of the supporting base. Topples in rocks usually require high cliffs, whereas topples in debris and soil fail on lower cliffs. The formation of tension cracks is caused by severe undercutting by fluvial agents, sea wave action or man-made scarps. Joint and bedding plane water pressures are a vital contribution to failure at the base of the column.

Slide

The term 'slide' is used for a movement of material along a recognizable shear surface. The shear surface type and the number of shear surfaces are used to divide the slide group.

SLIDE (ROTATIONAL)

Alternative terms: slump, rotational slip, rotational slide.

Rotational slides occur as a rotational movement on a circular or spoon-shaped shear surface. They differ in the degree of disintegration in the slide masses and the depositional features in the toe areas. Varnes (1978) and Buma and van Asch (1996) provide summaries.

Varnes (1978) defines a single rotational slide or slump as a 'more or less rotational movement, about an axis that is parallel to the slope contours, involving shear displacement (sliding) along a concavely upward-curving failure surface, which is visible or may reasonably be inferred'. A rotational slide has a small degree of internal deformation although sometimes soil slump material liquefies and transforms into a flow at its toe. In the terminology the American usage 'earthflow' is replaced in European literature by 'mudslide'. Rotational slides can vary from an area of a few square metres to large complexes of several hectares.

Given the relatively small degree of internal deformation, the slump matrix will essentially be the same as the surrounding undisturbed slope. Soil slides generally consist of fine-textured, cohesive materials, like consolidated clays, weathered marls and mudstones. Rotational rock slides often develop in formations of interbedded strong and weaker materials, e.g. marls and limestones or sandstones. In general, rotational sliding produces a disrupted, anomalous drainage pattern. Ponds and peaty areas may develop in depressions between slump units.

Movement of rotational slides starts with initial failure followed by rotation. It may disintegrate into several discrete blocks. In the head area, these blocks tilt backwards while sliding downhill and often flattening or even slope reversal occurs. Sliding along the flanks causes longitudinal and diagonal shear stresses. The lowest part of the slump mass moves over the toe of the failure surface, thereby bulging, cambering, overriding and producing transverse tension cracks.

Movement rates of slumps can vary by several orders of magnitude, between a few centimetres a year to several metres per month, while soil slumps can attain velocities up to 3 metres per second. Tilted trees (generally backwards in the head area, forwards in the foot and toe areas) can reveal the presence of rotational slides.

Typical causal situations are undercutting by waves or streams, excavation and other construction activities. Common triggering mechanisms are earthquakes, explosions and sudden increases of overburden or high water tables following periods of rainfall or snowmelt.

SLIDE (TRANSLATIONAL)

A translational slide is a non-circular failure which involves translational motion on a near planar slip surface. The movement is largely controlled by surfaces of weakness within the structure of the slope-forming material. Translational slides may occur in three types of material: rock, debris and soil. Depending on the slope angle and the velocity, slides will either stay as a discrete block on the failure surface or break into debris.

Block slide

Alternative terms: planar rock slide (rock block slide), slab slide (common usage for soil/earth block slide).

Large block slides are often part of extensive compound landslides involving rotational slides either at the toe or the head and occasionally mudslides at the edges of the landslide. Due to the geometry of the slip surface the mass may only move through the development of internal shears and displacements. Hutchinson *et al.* (1991) and Ibsen *et al.* (1996b) provide summaries.

Block slides are found in stiff, fissured or overconsolidated clays often in combination with stronger rock formations. The basal surface needs only a very low-angle for displacement, because the driving forces are usually very large. Large block slides are particularly sensitive at the toe and sometimes require only slight toe erosion to trigger failure. Movements involve deep settlement at the head, whilst the margins of the block slide are generally totally destroyed.

Block slides can be initiated or reactivated where construction loads the slope or where excavation undercuts or unloads the toe of an area of strongly developed joints or bedding planes that dip outward toward the natural slope. Block slides may move continuously in frequent pulses. In large slides movement is primarily controlled by wet year sequences and extreme or intense rainfall events. Though velocities are frequently low, the masses involved can be very great and extremely difficult or impossible to stabilize.

The primary cause of block slides is the presence of an abrupt change of rock type or a bedded rock sequence which provides a weak stratum dipping gently towards the slope. Strong discontinuities parallel to and into the face which clearly define a potential movement area are also helpful. The second condition concerning block failure is that the slope may be unloaded by erosion or excavation to a point where the potential failure surface crops out above or close to the base level.

Slab slide

Alternative terms: debris block slide, soil block slide, earth block slide, sheet slide (shallow translational failure in dry, cohesionless soils).

Slab slides are translational failures in slopes composed of coherent, fine soils or coarser debris with a fine matrix. Weathered soils, especially those derived from clays, mudrocks and silty-clays, are commonly involved. The weathered material normally moves on a shear zone close to a surface of unweathered or lightly weathered bedrock, a pedogenic horizon or a structural surface. A slab slide is dominated by the pedological or geological structure and frequently fails along discontinuities. Hutchinson (1988) and Ibsen *et al.* (1996a) provide summaries.

Slab slides are extremely susceptible to seasonal changes in groundwater levels or to loading at the head or unloading at the toe. Movement increases in wet months and may cease in dry periods. If the ground freezes and seasonally thaws the saturated conditions can initiate movement. Slab slides also occur at permeable/impermeable soil junctions. Movement takes place on low angle shear surfaces and is normally parallel to the ground surface.

Slab slides are caused by geometrical changes to the slope, water regime changes or human activity. They are dominated by the relationship between regolith depth and slope angles which determines the critical depth for failure. Melting of PERMAFROST is a special cause of saturated ground overlying an impermeable horizon.

Rock slide

A rock slide is a translational movement of rock which occurs along a more or less planar or gently undulating surface (Varnes 1978). It is typical for mountain slopes or rock exposures where the slope angle is close to, or parallel to, the dip of the rock. The movement is controlled by planar structural discontinuities, such as faults, joints and layering and the presence of weaker formations within the rock mass. Sorriso-Valvo and Gulla (1996) and Erismann and Abele (2001) provide summaries.

Rock slides are characterized by well-defined head scarps and flanks, a pronounced scar generally left with little or no debris, and a mass of debris that accumulates in the track or at the base. In case of a sliding rock mass a planar slope surface is developed. If the rock slides well away from the depletion zone, the scar and flanks may remain visible.

There are different mechanisms of movement of rock slides. If the movement is slow (mm to m/day) the whole mass may disaggregate because of differences in velocity along the yield surface. The frequency of events and magnitude of each single slide may vary. In velocity slides the mass disaggregates during movement, transforming it into a rock avalanche or a debris flow.

The essential prerequisites are steep slopes, intense jointing, bedding or fault planes dipping towards the open face and the preparation of the slope by unloading and weathering and the development of joint water pressures. The triggers are the undercutting of the toe support and earthquakes. The fundamental cause of a rock slide is the presence of a rock mass which produces such a stress that the resistance of the intact rock or the friction mobilized on existing discontinuities is exceeded.

Rock slides have a wide range in volume and velocities and pose considerable hazards to human settlements and lives. The destructive power of rock slides can be enormous in the case of rapid rock slides on steep slopes.

Debris slide

Alternative terms: shallow translational slides, sheet slides, soil slips.

Debris slides are failures of unconsolidated material which breaks up into smaller parts as the slide advances downslope. The material involved is mostly COLLUVIUM and weathered material of fractured rocks masses (i.e. flysch formations, shales and slates). The failure surface usually develops at the contact between the REGOLITH cover and the bedrock, and is roughly parallel to the ground surface. Clark (1987) and Corominas *et al.* (1996) provide summaries.

The speed of sliding and degree of runout tend to increase with slope angle and decrease with clay content (Hutchinson 1988). Velocities of up to 16 m/sec have been recorded. Many translational debris slides turn into debris flows. This occurs where water is available, and where the topography favours the convergence of both debris and water into concavities and channels. On very steep slopes, debris slides can reach high velocities. This is common in valleys shaped by glaciers in which morainic sediment is located high above the present river.

Debris slides are often triggered by intense rainfall or by earthquakes. The probability of a debris slide occurring is greatly increased by the destruction of vegetation cover by fires or logging. Sites most likely to provide failures are first-order basins with hollows where regolith can reach the maximum thickness and high slope angles. Failures are often caused by an increase in PORE-WATER PRESSURES following heavy rains which reduce the shear strength of the material. After failure, the breakage of the sliding mass allows the water to escape and the debris to stop. Debris slides can burst explosively out of a slope. Hazard assessments should be based on regolith depth and slope angle.

Mudslide

Alternative terms: earthflow (US usage), mudflow (redundant usage also climatic and temperate mudflow), slump-earthflow (complex, lobate mudslide form with minimal track).

Mudslides are a form of mass movement in which masses of softened silty or very fine sandy debris slide on discrete boundary shear surfaces in relatively slow-moving lobate or elongate forms. Brunsden (1984) and Ibsen and Brunsden (1996) provide summaries.

A mudslide is divided into source, track and lobe and accumulation zone units. The source has a bowl-shaped head. The material in this section is usually soft, weathered debris often with depressions containing water. The track evolves as an elongate or lobate channel through which the material moves. The accumulation zone develops at the base of the slope and normally consists of lobes of debris. Mudslides usually occur in saturated clays which have been described as fissured, mudstones, siltstones and overconsolidated clays with a generally medium plasticity.

Mudslide movement rates range from about 1–25 m/yr and are generally classified as slow mass-movement types. Extreme events range from hundreds of metres per day. Mudslide movement is normally seasonal, as the wetter weather increases the water content to the point where pore water becomes sufficient to generate movement. Mudslides in temperate areas display a pronounced winter–summer cycle. Movement usually commences in the late autumn, peaks in mid-winter and comes to a slow halt by late spring and summer. Heavy rainfall will frequently result in a mudslide surge. Movement can also be generated by undrained loading when there is rapid supply of debris at the head. Other causes are associated with the supply of water such as snowmelt or permafrost melt. Finally unloading of the toe area is important since it presents the development of passive resistance at the toe.

The essential plannning and engineering implication is that this form of slide requires water. Planning must include surveys for drainage installations. The removal of the source of water influx is critical, as is the reduction of pore-water pressure from the mudslide mass. Roads and other linear features are most vulnerable to movement at the lateral shears.

Lateral spreading

Lateral spreading describes the lateral extension of a cohesive rock or soil mass over a deforming mass of softer underlying material.

ROCK SPREADING

Alternative terms: gravitational spreading, gravity faulting, block-type slope movement, cambering and valley bulging.

Rock spreading is the result of deep-seated, plastic deformation in the rock mass, leading to extension at the surface. It may take place in a relatively homogeneous rock, or there might be a fractured capping stratum leading to gravitational stresses. Where the rock is relatively homogeneous, the moving mass breaks up into successive units arranged like horsts and grabens. Rock spreading in homogeneous rock masses is characterized by double ridges, trenches and uphill facing scarps which occurs often in high mountains. Uphill facing scarps evolve from the combination of ridge spreading and erosion of the uphill sides of the tension cracks. Zaruba and Mencl (1982) and Pasuto and Soldati (1996) provide summaries.

Rock spreading is very often associated with toppling, rock falls, slumps and mudslides. The rock masses involved are huge and generally more than one million cubic metres. The velocity is particularly low in comparison with other types of mass movement and seasonality is rarely observed. Lateral spreading phenomena are highly controlled by geological structures and often connected with deep-seated gravitational slope deformations.

Rock spreading is caused by an outward and downward movement of both valley sides along low-angle shear planes causing rock spreading at the ridge top. There may be sliding of blocks with respect to the whole ridge with minor crushing and redistribution of material at the point of the wedges. Alternative wedging along the centre of the ridge in combination with rotation of the peripheral blocks together with crushing at the areas of greatest stress may take place. Rock spreading rates range between 10^{-4} and 10^{-1} metres per year. Large-scale rock spreading is generally not seasonal.

The process depends on the local geological and topographical conditions. The horizontal or subhorizontal overposition of a thick slab of competent rocks, affected by a dense network of tectonic joints, on clayey materials which may behave as a visco-plastic medium, is a prerequisite for the process.

SOIL/DEBRIS SPREADING

Alternative terms: sudden spreading failure, lateral soil spreading, quick clay sliding, quick clay flow, soil liquefaction sliding.

Soil spreading is defined as the collapse of a sensitive soil layer followed by either settlement of the overlying more resistant soil layers, or progressive failure throughout the whole sliding mass. The duration is generally only a few minutes. Considerable lateral movement along the basal mobile zone occurs. The material deformations accompanying the spreading often cause loss of life and severe damage to roads, buildings and embankments. The areas involved are low angled and are ideal for both agricultural use and urban development. Areas of particularly intense hazards are often situated close to the shoreline or the banks of rivers, as toe erosion is a significant factor in destabilizing these slopes. Bjerrum (1955) and Buma and van Asch (1996) provide summaries.

QUICKCLAYS are found in areas which have been depositional marine environments adjacent to Pleistocene ice margins. Morphological features of quickclay slides which have their whole slide mass liquefied are the pear-shaped scar, and the presence of an extensive flow lobe on the lower side of the scar. This lobe can attain lengths of up to 1,000 m. Quickclay slides start in lower slope regions and extend upslope by retrogression. Due to loss of horizontal support, retrogression proceeds rapidly, both upslope and to the flanks. Due to remoulding the clay already subject to movement loses its internal strength and descends the slope as a more or less viscous slurry, and can travel up to several kilometres. In (quick) clay soils, initial failures are mostly caused by extremely high water contents (30–40 per cent) within the clay, oversteepening of the local slope and/or stress caused by loading.

Quickclay slides are initiated within a body of sensitive (‘quick’) clay. Sensitive clays are defined as clays which need only very little remoulding to transform into a viscous slurry. However, the whole sliding mass does not necessarily liquefy. It may be restricted to a thin layer of sensitive clay at some depth. The sensitivity of the clay is influenced by the dominant type of clay minerals and the depositional environment. Marine sedimentation results in flocculation and an open ‘card-house’ porosity structure. Strength reduction through interparticle repulsion is caused by leaching of salt in pore water over time.

Sand liquefaction slides are situated on coasts and adjacent to rivers or lakes. Especially sensitive are slopes already subject to mass movement, with disturbed drainage and corresponding high water contents in the soil layers. Varved clay formations are also sensitive to sand liquefaction failures along the boundary between impermeable clay layers and water-bearing fine sand or silt layers under high pore-water pressures. Tension fractures will develop at the head of the slide, dipping towards the sliding mass. The overhanging wall of the sliding block collapses to form a graben. The sliding blocks become subject to tilting, internal fracturing, subsidence, heaving and overthrusting, producing a very hummocky topography. The liquefied layer material escapes through tension cracks and may cause sand boils.

A triggering mechanism of sand liquefaction is a combination of prolonged periods of heavy rainfall or snowmelt, causing high initial pore pressures, and earthquakes. During an earthquake a soil undergoes cyclic stress loading and unloading, and pore-water pressures increase with each cycle. After a certain number of cycles the pore-water pressures become equal to the existent confining pressures and the soil suddenly loses its strength, suffering from considerable deformations without exhibiting resistance. This process is restricted to cohesionless soils such as sand and silt.

Flow

A flow is a landslide in which the individual particles travel separately within a moving mass. They involve highly fractured rock, clastic debris in a fine matrix or small grain sizes. Flow in its physical sense is defined as the continuous, irreversible deformation of a material that occurs in response to applied stress. They are, therefore, characterized by internal differential movements that are distributed within the mass.

ROCK FLOW

Alternative terms are: SACKUNG, sagging, rock creep, deep-seated gravitational creep.

Rock flows are creeping flow-type, deep-seated gravitational deformations affecting homogeneous rock masses. Rock flows are characterized by the high volume of the rock mass (several thousand to millions of cubic metres) and small total displacement rates. Structural elements and landforms associated with rock flows are high angle extensional shear planes in the upper part of the deforming slopes, producing graben-like depressions (trenches), double ridges, ridge depressions and troughs. The slope foot often shows compressional features such as bulging and, sometimes, low-angle shear planes. Mencl (1968) and Bisci *et al.* (1996) provide summaries.

The mechanisms of rock flows are not well known. Mencl (1968) postulates that in the central part of the slope, where the confining pressures are high and the deviator stress is too small to produce shearing, the rock mass deforms through viscous flow. At the uppermost and lowermost parts of the slope, where the confining pressures are low, the rock mass moves along the shearing surfaces. At depth, the high pressures induce plastic deformation without necessarily creating a proper slip surface.

Some rock flows display evidence of a constant rate of creep deformation. Others show a rapid reactivation phase in connection with extreme rainfall or earthquakes. Some rock flows may be characterized by a step-like evolutionary behaviour, including short active phases related to critical events and long-lasting dormant phases involving creep deformation at extremely slow rates. Under particular circumstances, the slow slope deformation can be turned into a catastrophic event leading to the collapse of a large-scale rock avalanche or debris flow. Rock flows can be considered as extremely slow preparatory stages of huge landslides which only in a few cases reach their final evolutionary step.

Rock flows are produced only where slopes are high enough to induce strong gravitational stress in the bedrock. Such conditions are typical of valley slopes in mountain areas and high coastal cliffs. Rock flows do not normally represent a major problem to planning and engineering. In recent years, however, there is a concern that large structures such as dams or hydrological sites may be placed at the toe of these phenomena.

DEBRIS FLOW

Alternative terms: mudflow (old usage), lahar (volcanic mudflow).

Debris flows consist of a mixture of fine material (sand, silt and clay), coarse material (gravel and boulders), with a variable quantity of water, that forms a muddy slurry which moves downslope. The flow moves in surges and includes the erosion of the channel bed and the collapse of bank material. Debris flows usually take place on slopes covered by unconsolidated rock and soil debris. Debris flows are characterized by the source area, the main track and the depositional toe. The flows commonly follow existing drainage ways. Some of the coarse debris will be deposited at the side of the track to form lateral ridges (levees). Deposits are accumulated where the channel gradient decreases or at the toe of mountain fronts. Successive surges will build up a debris fan. Some debris flows are exceptionally large, fluid, and can reach long distances beyond the source area. Debris flow is a gravity induced mass movement between landsliding and water flooding, although with mechanical characteristics very different from either of these processes. Pierson and Costa (1987) and Corominas *et al.* (1996) provide summaries.

Debris flows are a very destructive type of mass movement caused by heavy rain or snowmelt. In alpine environments debris flows are composed of coarser material from mechanical weathering and glacial deposits. Debris material in melting permafrost (near the lower limit of the discontinuous permafrost) has also been considered a debris flow source area. Debris flows which originate on the slope of a volcano contain vulcaniclastic materials and are called LAHARS.

Debris flows consist of large coarse material embedded in a fine-grained matrix. The coarse material is randomly distributed and individual beds are generally poorly sorted. Buoyant forces and dispersive pressures may concentrate boulders at the top of the deposit, forming reverse grading. The frequency of debris flow events is controlled by the rate of accumulation in hollows or channels, and by the recurrence of climatic triggering events.

Debris flows have well-graded deposits with a small clay content, generally less than about 5 per cent. They have a range in volume concentration of solids from approximately 25 to 86 per cent. Sediment concentration is the primary criteria for classification of flows given by Pierson and Costa (1987). There is a continuum from sediment movement in rivers to debris flows. Fluids with large sediment concentrations do not deform until a threshold strength is exceeded and they behave like a non-Newtonian fluid.

Debris flows are commonly triggered by an unusual presence of water created by intense rainfall, rapid snowmelt, and glacier or lake overflows. Debris flows are frequent in topographic concavities or hollows at first-order watersheds. This geometry supports the accumulation of colluvium and the convergence of groundwater flow necessary to cause the failure. Many debris flows start as a translational or rotational slide and then turn into a debris flow.

Series of debris flow waves are frequent and can be produced by breaching of the temporary dams or obstructions in the channel. The front lobe is composed of coarse blocks occasionally mixed with trees. During the progression of the debris flow through a channel, lateral ridges (levees) can develop. Overflowing of channel banks is a significant natural hazard of debris flows. The velocity of the flow depends on the size and sediment concentration, and on the geometry of the path. Observed velocities range from 0.5 to about 20 m/s. Large lahars can travel over a distance of more than 100 km and the rate of movement may reach more than 50 km/h. The erosion that occurs on both the channel floor and banks, cause some debris flows to significantly increase their volume.

The socio-economic impact and the loss of life, property and agriculture can be catastrophic. Even smaller mudflows and debris flows may cause serious damage, especially in mountainous regions. The deposits are also responsible for severe indirect damage and hazards such as damming of rivers or sudden debris supply to river systems. It is essential that potential source areas and runout zones are assessed.

SOIL FLOW (MUDFLOW)

Alternative terms: mudflow, alpine mudflow, earthflow, sandflow.

Soil flows may occur in three forms, wet mudflow, wet sand flow and dry sand flow. The wet forms are special categories of debris flow where the material is of a single and fine grain size and coarse clasts are rare. They are very mobile and can flow downslope quickly. They tend to follow gullies or shallow depressions and then to spread out into a flat fan or even a thin sheet when they reach low gradients. Soil flow conditions are abundant water, unconsolidated material and insufficient protection of the ground (i.e. lack of vegetation). Pierson and Costa (1987) and Schrott *et al.* (1996) provide summaries.

A wet soil flow contains relatively cohesive earth material that has at least 50 per cent sand, silt and clay. Thus, the term soil flow should be used for a flow with a significant lack of coarse-grained material. Typical source areas and starting zones are steep (25°–40°) slopes (e.g. moraines, proglacial zones), volcanic environments and mountain torrents.

A characteristic of soil flows is their ability to travel long distances (some kilometres) over even low slopes, usually following pre-existing drainage patterns. They are often termed viscous slurry flows, but they can be either viscous or inertial flows, depending on the driving forces. The flow behaviour is normally that of a visco-plastic type. The change from a slow creep to visco-plastic flow in clay-rich soils supersedes the destruction of strong bonding and the subsequent decrease in viscosity. Continuous undrained loading causes rapid readjustment of this mass and velocities of up to 10 m per second have been recorded.

Very rarely dry sand flows may occur. These form when a large mass of dry material falls from or over a steep slope and fluidizes on impact. Flow is then of 'rock avalanche' type with a track of uniform depth. Dry sand flows may also occur from dunes or similar sandy deposits. They require the sliding of material at the head and then either the transfer of momentum by cohesionless grain flow (i.e. grain upon grain momentum transfer) or descent over a cliff and fluidization. The first type can be observed in fluviglacial deposits or riverbanks.

Complex

It is common for mass movement processes to combine together and complex landslides occur where the initial failure type changes into another as it moves downslope. Compound landslides include two types of movement which occur concurrently within the same failure.

ROCK AVALANCHE

Alternative terms: STURZSTROM, rockfall avalanche, rock-slide avalanche.

FLOW SLIDE

A flow slide is a structural collapse of slope-forming material with momentary fluidization and is usually referred to as a high magnitude event both in terms of velocity and destruction. The high energy is capable of causing incredible devastation not merely through its impact on humans, by cutting communication and power lines or diverting a river, but also environmentally by littering the surrounding valley with debris. Flow slides are often associated with man-made tips and spoil heaps, although this type of failure may also occur in rock debris of geological origin. The internal structure has very little cohesion and the matrix ranges from clay-size particles to large blocks. Flow slides are a subclass of debris flows. The various causes of flow slides are an initial rotational failure at the head, vibrations or shocks, heavy rainfall, loosely deposited spoil heaps, removal of lateral support and rapid loading. Bishop *et al*. (1969) and Ibsen *et al*. (1996c) provide summaries.

The key characteristics of a flow can lie in their origin in artificial spoil materials but their behaviour is also found in many natural debris flow events. A flow slide is generally composed of loose material, which losses its cohesion with a reduction in strength, becoming a fluidized mass. The fluid may be air or water and, therefore, the dominant mechanism of a flow slide is fluidization or liquefaction. Sliding may occur at the head of the flow slide perhaps in the form of a rotational slide, but there is usually little indication of shearing at the subsequent stages of movement. Flow slides not only fluidize very quickly, but also rapidly consolidate and become solid when they cease moving, creating an additional hazard in the depositional area.

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SEE ALSO: factor of safety; fluidization; lahar; landslide; landslide dam; method of slices; quickclay; riedel shear; sackung; sensitive clay; sturzstrom

RICHARD DIKAU

MATHEMATICS

Mathematics provides an essential set of tools for the study of geomorphological phenomena. Aspects of mathematics particularly relevant to geomorphology include analysis of numerical relationships among measured quantities, the use of differential equations to describe processes, and descriptive mathematics related to form, such as geometry and topology. The former two aspects are critical when deriving equations to describe geomorphological processes or statistical relations between system variables. The latter is relevant when considering morphology of landscape features.

Prior to the twentieth century, the study of Earth sciences in the western world consisted mainly of qualitative consideration of early Earth history and landform origins. Until the mid-twentieth century, mathematics in geomorphology was limited to a few pioneering studies or engineering problems. Davis (1899), following earlier traditions, influenced geomorphological research for many decades by adopting a qualitative framework to describe regional landscape evolution in his 'CYCLE OF EROSION'. At around this time, Gilbert (1877) developed an innovative, quantitative research agenda based on the application of Newtonian mechanics to the study of landscape processes. This latter approach was not generally adopted at the time, perhaps because of difficulties in measuring processes at the large scales of study that dominated the discipline.

Around the mid-twentieth century, the 'quantitative revolution' in geomorphology emerged. Landmark papers such as Horton (1945) and Strahler (1952) signalled the shift to a quantitative approach. Morphometrics and numerical analysis of processes became dominant research themes. Scales of study decreased, as it is generally easier to measure system parameters at smaller scales. In the ensuing decades, many studies focused on establishing empirical relations between system variables. Physically meaningful expressions, in contrast to empirical relations, must be dimensionally balanced. In fact, what may appear to be fundamental relations in geomorphology do not actually fit this definition. Such equations, therefore, represent scale relations and are not true representations of the underlying physics. Physically based attempts to describe processes make use of mathematical tools, such as differential calculus, to portray fundamental conservation principles (mass, momentum, energy) in temporally evolving geomorphological systems. The Buckingham Pi Theorem, which is based on formal dimensional analysis, can be used in an attempt to derive rational equations that encompass essential system attributes.

The use of mathematical techniques is now firmly entrenched in geomorphological research. Continuing advances in technologies, such as remote sensing and radiometric dating, and the availability of improved data sets, such as DIGITAL ELEVATION MODELS (DEMs), have made it possible to measure and describe geomorphic phenomena at large spatial and temporal scales. Although complexities in the natural environment prohibit exact quantification of geomorphic processes, simplified numerical models of landscape processes have emerged due to advances in computing technologies. Sensitivity analysis, which examines the significance of changes in controlling variables on process operation, can be undertaken within a modelling framework.

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YVONNE MARTIN

MAXIMUM FLOW EFFICIENCY

The operation of fluvial systems can be described by the physical relationships of flow continuity, resistance and sediment transport. However, these three relationships are applicable to any alluvial-channel section and their solution involves more than three variables (width, depth, velocity and slope). As a result, the number of sections satisfying the three relationships is generally very large. In an attempt to understand the physics behind this problem of non-closure solution, Huang and Nanson (2000, 2002) propose a mathematical analytical approach that identifies a mechanism of self-adjustment. Among the numerous channel sections that satisfy flow continuity, resistance and sediment transport, a unique solution of HYDRAULIC GEOMETRY occurs when flow reaches the following inherent, optimal state defined as maximum flow efficiency (MFE):

$$\varepsilon = \frac{Q_s}{\Omega^{\lambda}} = a\ maximum$$

where ε is flow efficiency factor, Q_s is sediment discharge, Ω is stream power or $\Omega = \rho gQS$, and exponent λ has a value of 0.65–0.85. For two reasons this state can be regarded as a fundamental law governing the adjustment of fluvial systems. First, MFE can be directly derived from the widely applied LEAST ACTION PRINCIPLE (LAP) (Huang *et al.* 2002). Second, mathematically derived MFE channel geometry relations are highly consistent with empirical relations developed from a wide range of observations for stable canals and relatively stable river channels (Huang and Nanson 2000, 2002).

While recognition of the applicability of LAP to river systems in the form of MFE provides a soundly based and computationally economical method for determining stable alluvial channel geometry, its physical implications are much more profound. First, it is an advance over the thermodynamic analogies and the empirical formulas previously used to justify numerous extremal MODELS in geomorphology, for it clarifies confusion regarding which of these hypotheses should be regarded as rationally based. Indeed, MFE subsumes the earlier hypotheses of maximum sediment transport capacity and minimum stream power. Second, it identifies a fundamental cause for the formation of different river channel patterns. It is the balance between available stream power and imposed sediment load that ultimately determines equilibrium channel form. This balance has been shown to exist in the ideal case of a single-thread, straight and fully adjustable channel system (Huang and Nanson 2000, 2002). Ongoing research suggests that when the balance in the ideal system cannot be maintained due to the effect of physical restrictions, such as the imposed valley gradient, then planform or cross-sectional changes will occur to either consume excess energy or to increase transport efficiency over low gradients. As a consequence, meandering, braiding, anabranching and wandering channel patterns will be formed (Huang and Nanson 2002).

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SEE ALSO: hydraulic geometry; least action principle; models

HE QING HUANG AND GERALD C. NANSON

MEANDERING

Phenomenology of meandering

Meandering refers to the spontaneous evolution of a single channel to high values of SINUOSITY, or to a channel that shows this pattern (Plate 78). Meandering is one of three basic types of river planform, of which the other two are straight and braided (see BRAIDED RIVER) (or, more generally, anabranching). Individual channels of anastomosed rivers can show meandering (see ANABRANCHING AND ANASTOMOSING RIVER). Meandering with planform shape comparable to that in rivers also occurs in tidal channels, in marine channels produced by density CURRENTs, and in geostrophic ocean-surface currents like the Gulf Stream.

Plate 78 The meandering Pembina River, Alberta, Canada. Flow towards top of image

In classical meandering the width of channel remains constant as the sinuosity increases, so the channel planform is well described by a single sinuous line. The simplest function that gives an adequate description of this curving line is a relative of the common sine wave known as a sine-generated curve. A sine-generated curve is one in which the local direction of flow varies as the sine of distance along the channel. As meandering reaches high sinuosity values, the symmetric sine-generated curve develops an asymmetry that gives the channel path a more saw-toothed shape (Parker and Andrews 1986). The sharp 'corner' of the tooth is on the upstream side of the bend.

Origin of meandering

Observationally, meandering is favoured in fine-grained (sand and finer) rivers with high suspended load relative to bedload and low slopes, but it is known from steep and/or coarse-bedded rivers as well, so the former conditions are apparently not fundamental. Meandering river channels are also often associated with fine-grained and/or vegetated banks and usually have well-developed floodplains.

The first step towards understanding the mechanics of meandering was to understand the effects of streamline curvature on flow in channel bends. Rozovskii was the first to show mechanistically how streamline curvature leads to a secondary flow, that is, a closed circulation of the water across the direction of the vertically averaged flow. The secondary circulation moves fast-moving surface water toward the outer bank and slowly moving bottom water toward the inner bank. The net effect of this circulation pattern is erosion on the outer bank and deposition on the inner bank, leading to development of a bank-attached bar (a point bar) on the inner bank. Thus channel curvature tends to be self-amplifying – an example of positive feedback. As the bend grows, deposition on the inner bank often maintains a rough balance with erosion of the outer bank, keeping the width constant. The growth of the point bar is often recorded in the surface morphology as a set of scroll bars that trace previous positions of the inner bank (Plate 78).

Later work has elaborated on this simple model considerably. First, the dynamics of bend flow is influenced at least as much by mean-flow inertia and by bottom topography as by secondary circulation. Interaction of the secondary circulation and the mean flow also tends to displace the thread of maximum velocity downstream relative to the bend apices, leading to a tendency of the bends to migrate downstream.

A major next step was formal mathematical analysis of the stability of a straight channel, with the aim of providing a mechanistic basis for the occurrence of straight, meandering and braided channel patterns (Fredsoe 1978; Parker 1976) (see BRAIDED RIVER). Meandering is thought to correspond to cases where the main mode of instability is alternate bars, which are bank-attached bars on alternating sides of the channel. Alternate bars are predicted to develop in channels with widths of roughly 15 to 150 times the flow depth. The bars alternately deflect the flow from one bank to the other, producing a sinuous thalweg (planform trace of the deepest part of the channel). This initial sinuosity leads to fully developed meandering via the positive-feedback mechanisms discussed above. A more elaborate stability analysis (Blondeaux and Seminara 1985) shows that the meandering instability is actually a kind of 'resonance' between the original alternate-bar instability and a planform instability of the channel curvature. Given that they apply strictly only to the initial growth of the bend, it is remarkable that these stability analyses correctly predict the wavelength of fully developed meanders: about seven times the channel width.

Stability analysis suggests that the main control on channel pattern is the aspect ratio of the channel. Although the dynamics of channel width are still not entirely understood, it is clear that one of the main controls is the total effective sediment flux that must pass through the channel. Apart from directly controlling the width, the effective sediment flux also influences the aspect ratio indirectly: the slope is proportional to the ratio of sediment discharge to water discharge; as slope decreases, depth tends to increase. Hence reducing the effective sediment flux relative to water discharge reduces the width and increases the depth. The central role of the sediment/water discharge ratio in controlling the plan pattern is consistent with the empirical observation that meandering tends to occur in low-gradient rivers with relatively fine, suspension-dominated sediment flux. However, neither the gradient nor the sediment grain size *per se* is the real controlling factor.

Stability analysis helps explain the origin of meandering. However, since it treats only the initial instability, it does not constrain the final amplitude of the fully developed meanders. The ratio of amplitude to meander wavelength sets the sinuosity of the channel. There is still no complete analysis of the controls on the amplitude of fully developed river meanders. The main process that limits meander growth in rivers is cutoff of the bend by formation of a new, shorter channel across the bend. Thus one could view meander geometry and average sinuosity as the outcome of a competition between sinuosity production by meander growth and sinuosity destruction by cutoff (Howard 1992). The ease with which cutoff occurs appears to be controlled by the erodibility and resistance to flow of the point bar surface. Vegetation on the point bar helps prevent cutoff in several ways. Stems and leaves block flow and provide baffling that aids in the deposition of fine, cohesive sediment, while roots bind deposited sediment. It would be helpful if techniques could be developed to reproduce steady-state, fully developed meandering experimentally, but so far this has proved extremely difficult to do (Smith 1998).

Submarine meandering

Submarine channel systems formed by density currents produce meander patterns that are very similar to river meanders in planform geometry. However, the scale of submarine channels is much larger than that of river channels: depths are typically 100–200 m and widths typically several kilometres. The mechanics of submarine meanders are similar to those of subaerial meanders; the larger scale results mainly from the fact that the density difference between turbid and clear water is much less than that between water and air. Hence much deeper flows are required to provide sufficient force to move sediment particles.

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CHRIS PAOLA

MECHANICAL WEATHERING

Mechanical weathering causes disintegration of rock, without substantial chemical and mineralogical alteration or decomposition. The culmination of mechanical weathering is the collapse of parent material and diminution of clast size. Rock disintegration is caused by stresses exerted along zones of weakness in the material, which may include pre-existing fractures, bedding planes or intergranular boundaries.

Mechanical weathering processes promote rock breakdown by inducing stresses within the rock; these stresses may be produced by volumetric change in the rock itself or by deposition of, and/or volumetric change in, material introduced into voids in the rock. Volumetric expansion of rock may be induced by temperature changes, wetting and drying or pressure release. Volumetric expansion of material within rock pores and cracks predominantly involves salt and ice crystals. Although mechanical weathering processes may cause rock disintegration by themselves, they often operate in association with CHEMICAL WEATHERING and biological weathering processes.

Expansion of rocks and minerals

A rock unit beneath the land surface is subjected to high compressive stress from the rock and sediment layers above. Once surface erosion removes the overlying rock layers, the rock unit will tend to expand in the direction of stress (or pressure) release. This may result in sheet jointing parallel to the unloading surface. As the rock surface expands in response to the PRESSURE RELEASE, the tensional strength of the rock may be exceeded by the tensile stress due to expansion, causing cracks to develop perpendicular to the rock surface. In this way, the rock unit may be broken into smaller slabs, which increases the surface area available for attack by other weathering processes. Pressure release cracking is most common in granites and metamorphic materials, and may also be seen on massive sandstones. As the surface sheets are lost, curved surfaces are formed; on a large scale (outcrop size) this process is termed exfoliation, on a smaller scale (boulder size) it is termed SPHEROIDAL WEATHERING. Exfoliation plays an important role in the creation of some landforms such as INSELBERGS, tors, arches and natural bridges.

Expansion of a rock surface creates tensile stresses and contraction of the surface creates compressive stresses. This may be induced by changes in the temperature of a rock surface. In a similar fashion to the stresses induced by pressure release, temperature-induced stresses decline in magnitude with increasing distance from the exposed rock surface. Stresses are restricted to the outer few centimetres due to the low conductivity of rock, which prevents inward transfer of heat (Hall and Hall 1991). Physical disruption of rock due to thermal shock occurs during forest fires, although the effectiveness of this process depends on rock composition (Ollier and Ash 1983). Whether or not receipt of insolation and diurnal temperature cycles can drive this process has long been debated. Daily temperatures in hot desert environments may reach 50 °C, and rock surface temperatures can reach 80 °C, with rapid cooling at night. Early anecdotes about rocks cracking in the desert were dismissed as hearsay when early experimental studies suggested the process was not viable. More recent research has suggested that thermal expansion, or insolation weathering, may indeed cause rock disintegration, though its effectiveness may largely be dependent on sufficient moisture within the rock. Igneous rocks, which contain many types of minerals with differing coefficients of thermal expansion, may experience stresses as a result of differential thermal response of minerals to heating and cooling cycles.

Rock minerals may expand when water is introduced into their structure; certain clay minerals typically behave in this way. Clays such as smectites and montmorillonites have the capacity to absorb water into the mineral during periods of wetting, which causes the mineral to swell. Bentonite (Na-montmorillonite), for example, may increase in volume by up to 1,500 per cent due to hydration and swelling. The species of clay mineral determines the degree to which it will expand and contract on wetting and drying (Yatsu 1988).

Expansion of material in voids

Exposed rock surfaces experience cycles of WETTING AND DRYING WEATHERING related to rainfall events and periods of evaporation. Simple wetting and drying of some rocks may cause their breakdown. When water enters a crack, or void, on a rock surface, it will become adsorbed by minerals lining the crack which may show unsatisfied electrostatic bonds. Further ingress of water may induce a swelling pressure within the void. Evaporation will remove all water molecules except those strongly bonded to the mineral surfaces; the sides of the crack, or void, may be pulled together by the attractive forces between water molecules on opposing faces. In this way, cycles of wetting and drying may induce expansion and contraction, which can split susceptible rocks such as shale, schist and even sedimentary rocks. Rocks that may be affected by wetting and drying usually have high clay mineral content, structural weaknesses, high permeability and low tensile strength. Wetting and drying may increase the size and/or number of pores and microcracks in rock, which has important implications for both frost (see FROST AND FROST WEATHERING) and SALT WEATHERING; an increase in water absorption capacity and a reduction in rock strength will accelerate the action of these processes.

Frost weathering involves the breakdown of rock as a result of the stresses induced by the freezing of water. Water experiences a 9 per cent increase in volume upon freezing and, in a closed system, may create pressures (theoretically 250 MPa) that exceed the tensile strength of rock (typically 25 MPa). A closed system may be produced if water in a rock freezes rapidly form the surface downwards, allowing ice to seal water in surface cracks and pores in the rock. Experimental work has shown that numerous mechanisms are involved in frost weathering, not only volumetric expansion of water; the most significant include adsorptive suction, as pore water moves toward the freezing front, and expansion (0.6 per cent from $+4\,°C$ to $-10\,°C$) of absorbed water. Many experimental studies have shown the importance of rapid freezing rates (at least $0.1\,°C$ per minute), low minimum temperatures ($<-5\,°C$) and high rock moisture content in determining the efficiency of frost weathering in causing shattering of rock samples (McGreevy and Whalley 1985). Moisture content is particularly important as the presence of air in pore spaces in unsaturated rock allows ice to expand into empty pores and voids and prevents crack growth. Rock properties exert an important control on the efficiency of frost weathering, as texture and structure determine both water absorption capacity and strength. Igneous and metamorphic rocks tend to be most resistant to frost shattering, while shales, sandstones and porous chalk tend to be least resistant. Frost weathering is likely to be most effective in alpine and cold temperate environments where freeze–thaw cycles occur frequently and abundant moisture is available, rather than cold deserts and polar areas. Angular rock fragments produced by frost shattering are termed felsenmeer.

Salts, the chemical compounds formed from reactions between acids and bases, are very important in causing rock breakdown. The effects of salt attack can best be seen in arid, coastal and urban environments, where salts are available and rocks routinely experience desiccating conditions that allow the salts to crystallize. The most common salts found in rocks are halite (NaCl), gypsum ($Ca_2SO_4 \cdot 2H_2O$), sodium sulphate (Na_2SO_4), magnesium sulphate ($MgSO_4$), sodium carbonate (Na_2CO_3) and sodium nitrate ($NaNO_3$) and their hydrated forms. The stresses causing rock breakdown are produced by three mechanisms: crystallization of salts in rock cracks and voids, expansion of salt crystals on hydration and thermal expansion of crystals (Cooke and Smalley 1968).

The most potent cause of salt weathering is salt crystal growth (Goudie and Viles 1997). Crystal growth occurs as a result of a saline solution becoming saturated as evaporation occurs and/or temperature changes, or by mixing of salts in solution, termed the 'common ion effect' (Goudie 1989). The role of salt crystallization in causing breakdown depends on the pore-size distribution and the pore connectivity of the material, as well as its overall strength. A second cause of stress arises from the capacity of certain common salts to take significant quantities of water into their structure. This hydration causes volumetric expansion of the salt and may exert pressure on crack and pore walls. Sodium sulphate, for example, will expand by 313 per cent on hydration. Hydration expansion may occur in response to changes in relative humidity, which, as this is closely related to temperature, may be diurnal. The extent to which salts expand when they are heated depends on their thermal characteristics and the temperature ranges to which they are subjected. Most commonly occurring salts have coefficients of expansion higher than those of rocks, yet simulation studies have not demonstrated differential thermal expansion to be a very effective weathering mechanism in isolation from other effects of salts. One reason may be that rocks only experience thermal cycling in a shallow surface layer, so subsurface salts probably do not experience significant diurnal temperature changes.

Evidence exists for the severe damage caused by salts to many rock types in a range of environments. Weathering forms produced will vary with lithology, though landforms produced by weathering are usually small or minor forms. Salt weathering processes may produce CAVERNOUS WEATHERING, flaking, scaling and GRANULAR DISINTEGRATION of the surfaces of most rock types; porous sedimentary rocks are particularly susceptible. Decay to stone used in buildings and monuments is commonly caused by salt attack, induced by salt-rich environmental conditions (Cooke *et al*. 1993) and by polluted conditions in urban atmospheres (Cooke and Doornkamp 1990).

Other material which can expand in rock voids, causing internal stresses and eventual breakup, include plant material. Growth of plant roots or lichen thallus, for example, in cracks in rock may have a biophysical effect in creating growth stresses. The likely tensile stresses created are, however, smaller (3 MPa) than the tensile strength of most rocks. The impact of ORGANIC WEATHERING is the result of a complex suite of biochemical and biophysical processes, and cannot be explained by mechanical weathering processes alone.

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SEE ALSO: deep weathering; frost and frost weathering; honeycomb weathering; hydration; organic weathering; pressure release; salt weathering; wetting and drying weathering

ALICE TURKINGTON

MECHANICS OF GEOLOGICAL MATERIALS

Newton's laws of motion provide the basis for the science of classical mechanics, and the mechanics of geological materials involves mostly a special branch of classical mechanics called continuum mechanics. Continuum mechanics utilizes the same physical laws that govern motion of discrete bodies such as planets and billiard balls, but continuum mechanics addresses the internal deformation of bodies that cannot be idealized as discrete points.

Familiar geomorphological examples of deformable bodies that can be analyzed using continuum mechanics include water flowing in streams and soil moving down slopes. In treating such bodies as continua, a basic assumption is that the elemental constituents of the bodies (e.g. molecules of water or grains of soil) are minuscule relative to the scale of observable phenomena of interest. Observable motion of streams or slope debris typically involves momentum exchange amongst billions of water molecules or soil grains, and it is consequently impractical to analyse the movements of individual molecules or grains to predict the behaviour of the aggregate. A logical alternative is to analyse the aggregated behaviour of a mass of grains or molecules by treating them as a single deformable body, or 'continuum'.

Numerous textbooks provide excellent introductions to continuum mechanics. Examples range from mathematically rigorous treatises (e.g. Malvern 1997) to introductory books aimed at the interests of Earth scientists (e.g. Middleton and Wilcock 1994). An oft-overlooked but very informative book is the small classic by Jaeger (1971).

Continuum conservation laws

The central principles in geological continuum mechanics are conservation of mass, momentum and energy. Although both momentum and energy conservation apply in every geomorphological setting, momentum conservation is generally the more useful principle, because momentum is a vector quantity that includes information about the direction of motion, whereas energy is a scalar. However, if thermal effects or phase changes are important (as in melting or formation of ice, for example) energy conservation must be considered explicitly, in addition to conservation of momentum and mass. The discussion below assumes that thermal effects are negligible, and emphasizes purely mechanical behaviour that arises solely from conservation of momentum and mass.

The basic equations of mass and momentum conservation describe behaviour in four dimensions (space + time), and they apply to any continuous body, regardless of its composition or state (solid, liquid or gas). The equations can be written in mathematical vector notation (with a brief English translation beneath) as:

$$\text{mass conservation: } \frac{\partial \rho}{\partial t} + \nabla \cdot \rho \vec{v} = 0$$

local rate of mass increase + rate of mass efflux due to deformation = 0

$$\text{momentum conservation: } \rho \frac{\partial \vec{v}}{\partial t} + \rho \vec{v} \nabla \cdot \vec{v} = \rho \vec{g} + \nabla \cdot T$$

local rate of momentum increase + rate of momentum efflux due to deformation = force imposed by gravity + internal reaction force (stress).

In these equations the dependent variables are ρ, the local mass density within the continuous substance, and $\vec{v}$, the local vector velocity, which can vary as a function of position and time, t. In many geomorphological phenomena, the only imposed driving force is the so-called 'body force' $\rho\vec{g}$ due to gravitational acceleration, $\vec{g}$. Additional imposed forces can be included if necessary.

The final quantity in the momentum-conservation equation is the stress, T, which represents the reaction forces (per unit area) that develop within a deformable body as a consequence of the interaction between the driving force and local accelerations. Unlike a rigid body, in which the action-reaction involving gravitational driving force and acceleration can be represented by a familiar form of Newton's second law, $\rho(d\vec{v}/dt) = \rho\vec{g}$, a deformable, continuous body can react to external forcing by generating internal stress. Thus, the concept of stress is a key one in continuum mechanics, and it should be mastered by all students of physical geomorphology. In general stress is a second-order tensor quantity (commonly represented mathematically by a 3×3 matrix), and the stress tensor is symmetric (which eliminates the need for a separate angular momentum equation in addition to the linear momentum equation above). Textbooks that explain stress and tensors in the context of geological sciences include an excellent one by Means (1976).

An important general feature of stress is known as 'static indeterminacy'. This condition dictates that even if a continuous substance is motionless and the momentum equation above reduces to $\rho\vec{g} + \nabla \cdot T = 0$, the stresses cannot be calculated without specifying a 'constitutive equation' that summarizes the mechanism of stress generation. The only mechanical analyses in which static indeterminacy and the need for constitutive equations can be circumvented are analyses in which stress is assumed to vary in one direction only. These 'one-dimensional' analyses can be useful for building insight but seldom provide accurate models of multidimensional geomorphic phenomena.

Constitutive equations and rheology

Two main branches of continuum mechanics, solid mechanics and fluid mechanics, have developed from observations of the fundamentally different mechanisms by which solids and fluids (liquids and gases) generate stress. In solid mechanics a pivotal observation is that, for sufficiently small deformations, stress is simply proportional to the magnitude of deformation (or, more precisely, proportional to the magnitude of strain, which may differ subtly from deformation). First quantified by Robert Hooke (1635–1703), this observation led to the theory of linear (or 'Hookean') elasticity. A similarly pivotal postulate, first made by Isaac Newton (1642–1727), was that fluids deform such that stress is simply proportional to the *rate* of deformation. Subsequently confirmed by experiments, this postulate led to the linear (or 'Newtonian') theory of viscous fluid flow.

Equations that relate stress to deformation are known as constitutive equations, because they express the influence of a body's constitution (or rheology) on the internal reaction forces that generate stress. In effect, constitutive equations (i.e. rheological models) serve as surrogates for momentum exchange that occurs at scales too small to be resolvable in a particular setting or observation (for example, momentum exchange by colliding water molecules in a stream observed by eye). The constitutive equations for linearly elastic solids and linearly viscous fluids may be written in simple forms as

$$\mathrm{T} = \epsilon D \text{ (linearly elastic behaviour)}$$

$$\mathrm{T} = \eta \dot{D} \text{ (linearly viscous behaviour)}$$

where ϵ represents elastic moduli, η represents dynamic viscosity, D represents deformation, and $\dot{\mathrm{D}}$ represents deformation rate. Whole treatises expound the detailed meaning of these equations and the detailed definition of D and $\dot{D}$ (which are tensor quantities, like stress). Here it suffices to emphasize that these well-known constitutive equations express straightforward connections between stress and a measurable macroscopic quantity such as D or $\dot{D}$.

Stresses in Earth-surface fluids such as water and air can be represented with fair accuracy by a simple constitutive equation describing linearly viscous behaviour, and stresses in a solid rock can be represented with similarly good accuracy by linearly elastic behaviour (provided the rock does not fracture). However, many materials encountered in geomorphology are not so simple. Rocks, soils and sediments that undergo large, irreversible deformation (as might occur in a landslide, for example) exhibit neither linearly viscous nor linearly elastic behaviour. Diverse constitutive equations have been proposed to represent such behaviour, but the discussion here introduces only the most significant equation, the Coulomb model.

Experiments demonstrate that stresses in soils, sediments and fragmented rocks undergoing large deformations adjust plastically to maintain limiting values, independent of the deformation magnitude or rate. These limiting 'plastic yield' stresses depend principally on friction due to rubbing and interlocking of adjacent grains, and to a lesser degree on cohesive bonding between grains. On planes of shearing the limiting shear stress τ is described by

$$\tau = \sigma \tan \phi + c \text{ (Coulomb plastic behaviour)}$$

where σ is the normal stress on the shear plane, φ is the angle of internal friction, and c is the cohesion. This equation, first posited by Charles Augustin Coulomb (1736–1806), has withstood repeated testing but does not provide the same straightforward interpretation as the linearly elastic and viscous equations noted above. Nonetheless, the Coulomb model has proved very useful for analysing phenomena such as LANDSLIDES, DEBRIS FLOWS and incipient motion of BEDLOAD particles in streams.

Initial and boundary conditions

In addition to conservation laws and constitutive equations, continuum mechanics requires specification of initial conditions that isolate a phenomenon in time as well as boundary conditions that isolate it in space. (For example, to mechanically analyse the behaviour of a flood, one must specify the initial channel geometry and the distributions of water-surface elevations and velocities prior to the flood onset.) Appropriate specification of these 'auxiliary conditions' can be the crux of successful mechanical modelling, because geomorphological phenomena seldom occur in isolation from the surrounding environment. Nonetheless, mechanical analyses of such 'open' geomorphological systems can provide key insights if auxiliary conditions are specified with care and precision (Iverson 2003).

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RICHARD M. IVERSON

MEGAFAN

Large fluvial depositional features that have been defined thus (Horton and DeCelles 2001: 44):

> Fluvial megafans constitute volumetrically significant depositional elements of sedimentary basins adjacent to mountain belts. A fluvial megafan is a large ($10^3 - 10^5$ km^2), fan-shaped (in plan view) mass of clastic sediment deposited by a laterally mobile river system that emanates from the outlet point of a large mountainous drainage network. Modern fluvial megafans have been recognized in nonmarine foreland basin systems at the outlets of major rivers that drain fold-thrust belts, particularly in the Himalayas and northern Andes. Although fluvial megafans are similar to sediment gravity flow-dominated and stream-dominated alluvial fans in terms of their piedmont setting, planform geometry and sedimentation related to expansion of flow downslope of a drainage outlet, fluvial megafans are distinguished by their greater size (alluvial fans rarely exceed 250 km^2), lower slope, presence of floodplain areas and absence of sediment gravity flows. The term 'terminal fan' is commonly used for a large, distributary fluvial system in which surface water infiltrates and evaporates before it can flow out of the system. A fluvial megafan therefore may be considered 'terminal' in cases where fluvial channels run dry before reaching bodies of water downstream. Although fluvial megafans are clearly related to the emergence of a large mountain river onto a low-relief alluvial plain, their stratigraphic evolution in nonmarine foreland basins may also be critically dependent on variables such as sediment flux, water discharge, drainage catchment size, catchment lithology and subsidence rate, factors that are ultimately controlled by tectonic, climatic and geomorphic processes.

Alternative names include megacone, inland delta, wet alluvial fan and braided stream fan. Some especially impressive megafans occur on the north side of the Ganga plain in India. They are fed from the Himalayas and may have formed in the Late Pleistocene when coarser grained sediment and high sediment and water discharges were available (Shukla *et al*. 2001).

Not all megafans occur in such dramatic settings as those of the Andean and Himalayan forelands. For example, the Okavango Fan of northern Botswana has accumulated in a graben and has been deposited by a low sinuosity/meandering river rather than a braided stream system (Stannistreet and McCarthy 1993).

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A.S. GOUDIE

MEGAGEOMORPHOLOGY

The term 'mega-geomorphology' was introduced in 1981 to designate a London conference of the British Geomorphological Research Group (Gardner and Scoging 1983). The original intent was for the term to apply to geomorphology on the scale of plate tectonics, biological evolution and macro-climatic change. Thus, it was to be concerned with entire landscapes, through histories of millions of years, in the context of continental or macro-regional evolution.

In 1985 another conference, 'Global Megageomorphology', was organized in Oracle, Arizona, and this meeting further refined geomorphological study at the largest spatial and temporal scales (Baker and Head 1985). Particularly relevant issues for the Oracle conference were the use of orbital remote sensing procedures to produce global mapping and analyses, studies of continental-scale denudation, the relationship of geomorphology to regional tectonics, global environmental change, and the geomorphology of other rocky planets besides Earth. The last issue is particularly important in both philosophical and historical perspectives. The history of geomorphological study of Earth began with scientists observing their immediate surroundings, and then generalizing to explanations that placed those observations in a larger context. In contrast, the study of other rocky planetary surfaces always begins at the largest spatial scales, through the remote sensing instrumentation of flyby planetary missions. The geomorphological understanding of Mars and Venus, the most Earthlike of the known planets, began at the megascale, while understanding of Earth's surface began at small scales and only evolved to the megascale after a long history of observations at the scales that were most accessible to human observers. A result of this history is the set of theories that currently constrains geomorphological inquiry to somewhat limiting viewpoints (Baker and Twidale 1991; Baker 1993). Megageomorphology affords an opportunity to break the constraints and develop new sets of theories.

The combined issues of scale and viewpoint are highlighted by a survey in the United Kingdom that found for the early 1980s that 75 per cent of geomorphological research concerned small-scale, modern process studies and another 15 per cent concerned Quaternary studies (Gardner and Scoging 1983). This emphasis on small scales of time and space is indicative of a reductionistic epistemological perspective in which one presumes that small-scale studies will integrate over time to generate a theory of the whole. Small-scale, modern process studies also tend to minimize the geomorphological role of rare processes of extreme magnitude because these are both remote from possible direct observation and destructive of most attempts to measure (Baker 1988). In contrast, as shown in Figure 107, the processes responsible for landforms and landscapes operate over a broad range of spatial and temporal scales. Moreover, the responses to these processes add further extensions of time and space to the zone of relevant natural operation for geomorphological change. One must conclude, 'Contemporary process studies are of little worth in evaluating landscape evolution' (Church 1980).

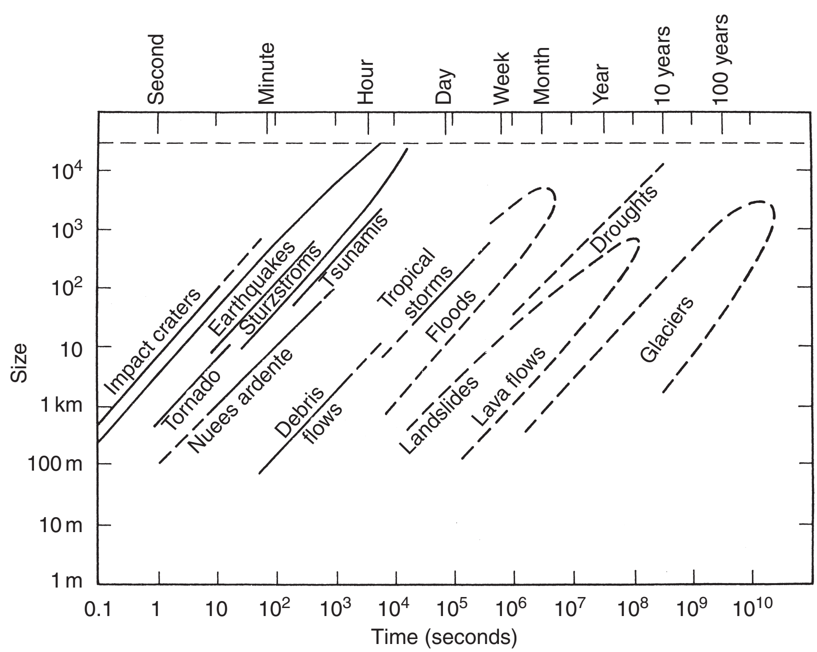

Figure 107 Scale relationships for various geomorphological processes, based on Carey (1962)

Many geomorphological systems are nonlinear, with thresholds that involve negligible responses at the process scales most easily measured in the field, while the relevant processes are both rare and of extreme magnitude (Baker 1988). The debates over the origin of the Channeled Scabland in eastern Washington in the 1920s illustrate this issue because a very large-scale process of immense magnitude was involved (Baker 1978). It is only because the process of megaflooding was recognized and understood for the Channeled Scabland that subsequent work has been able to show that cataclysmic megafloods played the dominant role in landscape development for other parts of Earth, and, somewhat surprisingly, for parts of Mars as well (Baker 2002).

Modern megageomorphology makes extensive use of global observations from spacecraft that employ a variety of imaging and remote-sensing instrumentation, including multispectral imaging, radiometers and radars. Image processing of digitally formated data has revolutionized the ability to study landscapes at the largest spatial scales. The theme of global-scale remote sensing is developed in the book *Geomorphology from Space* (Short and Blair 1986). A full scheme of geomorphology at large spatial scales is readily achieved in regard to the new technologies (Baker 1986).

Other technological advances that are stimulating megageomorphology include the quantitative geochronology of geomorphic surfaces and deposits; the widespread availability of digital topography; and the mathematical modelling of landscape evolution. These elements are creatively employed in the rapidly developing subfield of tectonic geomorphology (Burbank and Anderson 2001). In essence, the tectonic geomorphologist performs thought experiments with the computer, then tests those notions against the topographic response and the ages of elements in the landscape, such as glacial moraines, stream terraces and denudation surfaces. This can be done for entire regions that comprise the major tectonic elements of the planet.

The two major factors for geomorphological evolution at very large scales are tectonic factors arising from forces inside Earth (with the occasional imposition of extraterrestrial impacts) and the denudational factors arising from Earth's atmosphere, as largely summarized in its climate. Both these concerns are critical to any modern sense of global geomorphology (Summerfield 1991). The understanding of Earth tectonics was revolutionized in the 1960s and 1970s by the plate-tectonic theory. During the 1980s and 1990s climate became the central issue for international initiatives to understand global environmental change and the operation of Earth as a system. With the theoretical underpinnings of these two major elements, megageomorphology is now poised for major development as a science.

It is particularly interesting that a form of megageomorphology developed after 1945 in the former Soviet Union. Because of its practical application to mineral and petroleum exploration, the details of this science were subject to state secrecy, and its community of practitioners became somewhat isolated from the greater world scientific community. The focus of this science was the study of morphostructures, which are linear and circular elements of continental landscapes controlled by tectonics and denudation. The morphostructures exhibit hierarchical relationships with one another, and they evolve over geological timescales (Baker *et al.* 1993). Morphostructural analyis deciphers the complex interaction of long-term endogenetic processes with surface relief. It proved especially useful in identifying various concentric, circular or oval, and linear hidden dislocations of basement rocks that on Earth are commonly obscured by sedimentary or volcanic cover, deformations and intrusions. Though not tied to the modern plate-tectonic ideas about Earth's large-scale evolution, morphostructures are surprisingly similar to various quasi-circular shaped upland regions discovered on Venus by the planetary missions of the 1980s and 1990s. These also exist in hierarchical arrangement and may be related to mantle plume tectonics (Finn *et al.* 1994). Merely thinking at the megascale led to some surprising results.

Sharp (1980: 231) observes, 'One of the lessons from space is to "think big"'. This theme, combined with new analytical tools for geomorphological study at very large scales affords an opportunity for discovery and scientific excitement. Unifying models of global tectonics and climate dynamics afford the scientific framework for large-scale studies. Orbital remotes sensing, digital topographic data and geochemical tools for dating Earth history all permit the quantification of geomorphological parameters in greatly expanded temporal and spatial domains. Moreover, the science is immensely stimulated by the discovery of entire new landscapes on the surfaces of other rocky planetary bodies.

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SEE ALSO: extraterrestrial geomorphology; global geomorphology; tectonic geomorphology

VICTOR R. BAKER

MEKGACHA

A Setswana term for the DRY VALLEY systems which traverse the flat, sandy terrain of the Kalahari region of southern Africa. These broad, shallow, drainage features contain CALCRETE and SILCRETE in their floors and flanks and are also referred to as *laagte, omuramba* or *dum* in other regional languages. The origins of *mekgacha* (singular *mokgacha*) have been ascribed to episodes of permanent or ephemeral fluvial activity during wetter periods of the Quaternary, and, at longer timescales, to RIVER CAPTURE, groundwater sapping or DEEP WEATHERING focused along geological lineaments (Nash *et al*. 1994). Two groups of valleys can be identified (Shaw *et al*. 1992). The first are the exoreic (externally directed) Auob, Nossop, Kuruman and Molopo systems which drain the southern Kalahari and connect to the Orange River. The second are the endoreic (internally directed) systems which focus upon the Okavango Delta and Makgadikgadi Depression. Surface runoff is comparatively rare within the Kalahari and, as a result, flow within *mekgacha* is unusual. However, most exoreic systems are spring-fed and contain water in their headwater sections, with more extensive flooding following prolonged rainfall. In contrast, the endoreic systems are effectively fossil networks and only contain water in seasonal pools. Floods may occur in such networks after exceptional rainfall, but only two have been documented in the historical period (1851 and 1969), both in the Letlhakane valley (Nash and Endfield 2002).

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DAVID J. NASH

MELTWATER AND MELTWATER CHANNEL

Meltwater can be produced from the melting of snow (nival meltwater) or GLACIERS (glacial meltwater). The nival and glacial meltwater regimes represent two ends of a spectrum, as in many instances there may be contributions of both within the same catchment. The amount of meltwater produced is determined by the energy balance, a key component of which is solar radiation, thus most melt will occur at the ice or snow surface. Melting can also occur at the base of a glacier due to either geothermal heat or the effects of high pressure. However, melting will predominantly be related to air temperature, thus the temporal pattern of meltwater generation is not constant and varies over daily, annual and longer timescales. In nival regimes highest flows tend to occur in spring (the start of the melt season) as temperatures begin to rise above freezing. Meltwater discharge will usually exhibit a diurnal pattern, with a peak in the afternoon and a low around dawn, related to the cycle of daily temperature change. Over time, as less snow remains to be melted, diurnal fluctuations become more muted and flow will eventually cease altogether. Glacial meltwater production also exhibits a similar type of pattern, except that the melt season is longer and peak meltwater discharge is later. Glacial meltwater discharge displays a gradual rise through the melt season as the melting of snow on the ice surface is then followed by glacier melt. Peak flows in mid/late summer when the entire glacier, devoid of the insulating surface snow cover, contributes to melt. Outside the summer melt season flows can become very low or cease altogether. The specific meltwater regime is very closely tied with the environment, for example, glaciers in more polar environments will tend to have a shorter melt season and more muted diurnal discharge signature than those of mountain glaciers in temperate locations (Tranter *et al*. 1996). In addition to these more regular and predictable discharge regimes, meltwater can also be released catastrophically in OUTBURST FLOODS.

Once snow or ice has melted it can take several pathways with respect to its exit from an ice mass: as surface (supraglacial) channels, pipes or conduits within (englacial) the ice or as flow at the base (subglacial) of the ice (Shreve 1972). Supraglacial channels can be up to a few metres deep and because the ice is smooth water velocities are high (3–7 ms^{-1}). Supraglacial water will either flow off the ice surface or descend vertically into the ice via holes called MOULINS where the water connects with the pipes or conduits of the englacial system. Englacial pipes range in diameter from a few millimetres to a few metres, water can also flow through thin veins between ice crystals, but at a greatly reduced rate. Englacial water will often connect to the subglacial flow system at the base of the glacier. Where a cold-ice basal thermal regime predominates the supraglacial drainage system is often dominant, in a warm-ice situation a complex system involving all three types of drainage is often present. However, the drainage system is not fixed within the ice mass and will change and develop through the melt season as some channels or pipes open up and others close. In addition to this pattern of flow routes, meltwater can also be stored in supraglacial, englacial or subglacial lakes and ponds.

Subglacial meltwater has the biggest significance in terms of glacier dynamics and also in terms of providing evidence of meltwater flow in the landscape once the ice has receded. Subglacial meltwater significantly increases the rate of basal sliding and hence how fast the ice will move. However, if there is a highly permeable substrate at the base of the ice then very little meltwater flow may occur at the ice–bed interface, in which case basal sliding will not be greatly enhanced. A key factor about subglacial meltwater (and also englacial) is that due to the pressure of the weight of the overlying ice, the water can be flowing at much greater pressures than would be experienced under normal atmospheric pressure conditions at the surface. The direction of flow is determined by the ice surface slope and to a lesser extent the bed topography, such that uphill flow is possible (Shreve 1972). As the ice surface steepens, subglacial flow becomes less influenced by bed topography. Flow at the base will thus trend in the general direction of ice flow, often following valley floors and crossing divides at the lowest point.

Subglacial meltwater flow can occur in a distributed or discrete system. Distributed systems include sheet flow or linked cavity networks, discrete systems encompass the full range of channelized flows. Sheet flow is where meltwater exists as a thin continuous layer between the ice and the bed. Research has shown that this type of flow is relatively unstable and that flow is more likely to occur in channels (Shreve 1972). The linked cavity system is where basal hollows are linked by narrow short connections. Channelized systems can be cut down into the bed (Nye channels or N-type), cut up into the overlying ice (Röthlisberger channels or R-type) or display characteristics of both. The discrete (channelized) drainage systems transport meltwater more efficiently than the more circuitous pathways of the distributed system. The type of basal meltwater system thus has an important influence on water pressure and hence glacier motion. Meltwater can switch between these different flow systems and it has been suggested that this can trigger GLACIER surging.

Nye channels cut into bedrock or consolidated sediments leave the most distinctive imprint on the land surface once the ice has retreated. Due to the pressure conditions experienced by subglacial meltwater and the nature of meltwater supply to the subglacial system Nye channels often display very different characteristics to a normal subaerial fluvial system. For example, they can possess an undulating long profile, very steep gradients, lack any significant drainage basin and have an abrupt inception or termination (Glasser and Sambrook Smith 1999). The nature of the substrate beneath the ice will have an important influence on the overall channel morphology. For example, it is not uncommon for a channel to change its morphology over very short distances as it goes from being deep, narrow and incised in bedrock and then wide and shallow when passing over a deformable bed (Plate 79). Channels can range greatly in size, from tens of metres up to 100 kilometres long or from tens of metres to several kilometres wide. Once the channels reach the kilometre scale they are often referred to as TUNNEL VALLEYs rather than Nye channels. Channels can occur as either isolated features or part of a much larger channel network. The most spectacular form of isolated meltwater channel is what is referred to as a chute channel; these often occur on the flanks of a bedrock slope and are very steep and incised. They are thought to form as rapidly descending water within an ice mass reaches the glacier bed and cuts down into it. In contrast to these isolated channels, networks of large tunnel valleys can extend over very large areas and have been reported from many parts of North America and Europe that were covered by ICE SHEETs. The origin of such large features is thought to be due to either the catastrophic release of water stored in large subglacial lakes or that they were cut by normal meltwater flows over extended time periods. By mapping networks of exposed meltwater channels and coupling this with the theory of meltwater flow under pressure it has been demonstrated that reconstructions of the likely extent and dynamics of an ice mass is possible (Sugden *et al.* 1991).

Plate 79 A meltwater channel (flow away from camera) incised ~7 m into bedrock, Cheshire, England. Downstream the channel becomes wider and shallower as it passes over unconsolidated sediment before disappearing completely 500 m from where this photograph was taken. Photograph taken by N.F. Glasser

As well as being erosive, meltwater can also be a significant agent of deposition. For example, an ESKER is a long narrow ridge of sediment often deposited in subglacial channels as a result of high rates of meltwater flow. When subglacial channels completely fill with meltwater they can behave like pipes, allowing water under pressure to move, and subsequently deposit, large volumes of sediment in a single event.

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SEE ALSO: glacifluvial; subglacial geomorphology

GREGORY H. SAMBROOK SMITH

MESA

Mesa (the word is of Spanish origin) is a steep-sided and flat-topped hill or ridge rising above a flat plain, usually built of flat-lying soft sedimentary rocks capped by a more resistant layer, e.g. of shales overlain by sandstones. In volcanic terrains a former lava flow may act as a caprock; likewise, this role may be assumed by a blanket of DURICRUST. Mesas originate due to unequal scarp retreat, in the course of which parts of the plateau become isolated and remain standing in front of the retreating scarp. Lava and duricrust-capped mesas may form due to dissection of a plateau or through the mechanism of relief inversion (see INVERTED RELIEF), then the distribution of mesas indicates the position of a former valley floor or lava flow.

Mesas steadily decrease in size through slope retreat accomplished mainly by various kinds of mass wasting and gully erosion; however, due to the presence of a resistant cap they may be very durable landforms surviving long after the initial scarp retreated. With time, they are reduced to BUTTEs, but there are no agreed criteria as to when a mesa becomes a butte. Mesas are climate-independent landforms although those in desert areas have the most distinctive appearance.

SEE ALSO: caprock; sandstone geomorphology; structural landform

PIOTR MIGOŃ

METHOD OF SLICES

Slope stability is at present routinely analysed by the Limit Equilibrium approach, derived somewhat loosely from the Theory of Plasticity. It is based on the assumption that the pattern of stress in a failing slope can be determined from static equilibrium, without the need to consider stress redistribution due to elastic and inelastic straining. The stresses acting on the boundary of the sliding body ('rupture surface') can then be compared with available strengths, to evaluate equilibrium. The 'FACTOR OF SAFETY' is usually defined as a ratio by which available soil or rock strength may be reduced, without causing failure.

The Limit Equilibrium approach was originally applied to the sliding of rigid portions of a slope, along assumed rupture surfaces. Coulomb was the first to calculate the Factor of Safety of a block above a planar surface. Later, circular ('Swedish Circle') surfaces were analysed, as it was observed that slides in clay are often rotational.

In the 'Method of Slices' (Fellenius 1927), the sliding body, viewed in cross section, is divided into vertical slices. An approximate assumption is made that the vertical stress at the base of each slice is constant and equal to the weight of the material column above it. In the 'Ordinary' ('Fellenius') method, all stresses acting at the vertical boundaries between adjacent columns are neglected. The equilibrium problem then becomes statically determinate and could be solved by simple vector analysis. The Factor of Safety is equal to the ratio of the sum of available strengths on all column bases, to the sum of applied shear stresses. The results are often excessively conservative.

Bishop (1955) realized that, with a circular rupture surface, it is unlikely for the inter-slice *shear* stresses to be very high. He therefore assumed these to be zero and derived the normal and shear stresses at the base of each column from vertical force equilibrium. The Factor of Safety can then be evaluated from the moment equilibrium of the slice assembly, without the need to neglect the normal inter-slice forces. This method neglects horizontal force equilibrium, but nevertheless produces very accurate results when compared with more sophisticated approaches, for circular and some non-circular surfaces.

More detailed methods have been derived, taking into consideration inter-slice shear and all three equilibrium conditions. Recently, a much more sophisticated approach, based on numerical stress–strain analysis of the slope, has been developed, so called 'Stress Reduction Method' (Dawson *et al.* 1999). One of the advantages of this method is that it removes the need to predetermine the shape of the rupture surface. However, it is much more difficult to implement and lacks the long track record of practical experience, inherent in the Method of Slices. Thus, for the foreseeable future, the Method of Slices will continue to be an important tool for the analysis of slope stability. An extension to a three-dimensional 'Method of Columns' has now also been developed (e.g. Hungr *et al.* 1989).

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OLDRICH HUNGR

MICRO-EROSION METER

Many theories of landform evolution over time can only be tested through a knowledge of erosion rates. Earlier ideas of erosion sequences were constructed through logic and deduction from available morphological and sedimentary evidence. However, the answer to the question: 'could the landform have actually evolved in the timescale envisaged?' must involve a knowledge of rates. Two sets of endeavours at different scales devolve from this question, first large-scale measurements, for example on drainage basins, and smaller scale measurements. The latter may be performed more precisely and over a short space of time but the issue then remains of how far the results may be influenced by conditions at the time of measurement and of how they may be scaled up over a longer time span and a larger scale. Thus many of the data derived have been useful in short-term experimentation but extrapolation over longer time spans, over which conditions may be different to those obtaining at the present, remains an issue (Trudgill *et al.* 1989, 2001).

It was realized by High and Hanna (1970) that a micrometer dial gauge, used in engineering, could provide measurements of up to 0.0001 mm or even greater precision. A measurement is made of the height of a rock surface relative to some fixed datum. The instrument consists of a micrometer dial gauge and attached micrometer probe which is mounted onto a tripod framework. The tripod gives the instrument stability and it rests on three reference studs drilled securely into the rock. The measurement of the surface height of the rock relative to the studs can be made at repeated intervals, yielding results of surface lowering in mm a^{-1}. Initial meters had probes which were fixed and the tripod could be rotated to three positions, yielding three measurement points. Later meters (Trudgill *et al.* 1981; Trudgill 1983) used a traversing mechanism where not only was there a tripod base plate which could be rotated, additionally the dial gauge could itself be placed in several reference positions enabling a much larger number of points to be measured.

In terms of methodological limitations, Spate *et al.* (1985) suggest that erosion of the rock by the tip of the probe could lead to a fictitious annual loss of around 0.019 mm a^{-1}. They reported that when repeated measurements were made successively then surface lowering was recorded, the more so in softer limestones. For replicate readings at one site, on a harder (Buchan) limestone their data showed differences of 0.0001–0.0052 mm and for a softer (Gambier) limestone 0.0090–0.0284 mm. This may be initial surface compaction rather than actual erosion. Their data suggest that up to 0.0126–0.0284 for softer limestones and 0.0016–0.0052 for harder limestones will be an artefact of probe impact for any one measurement rather than actual erosion.

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STEVE TRUDGILL

MICROATOLL

Microatolls are discoid colonies of massive corals that grow in the lower intertidal zone on shallow reef flats. Corals are organisms that secrete a calcareous exoskeleton and they are major contributors to a CORAL REEF that may be preserved in the geological record as limestone. Microatolls are corals that grow at the reef-atmosphere interface, adopting a predominantly lateral growth form, constrained in their upward growth by exposure at lowest tides. This upper limit to coral growth reflects physiological factors that inhibit coral polyps when exposed.

Microatolls can be formed by several species of corals, but massive corals, such as *Porites*, are particularly prominent and more likely to be preserved (Plate 80). These growth forms have received particular attention because their flat upper surface is limited in terms of upward growth by exposure and is therefore related to regional sea level. There was initially debate as to whether the distinct form of microatolls might be due to sediment accumulation on top of the coral or nutrient limitation, but it has been demonstrated that it is water level that constrains vertical coral growth (Stoddart and Scoffin 1979). It is possible to examine the pattern of growth because corals form growth bands which are generally annual and which can be detected using X-radiography. The banding within microatolls confirms that growth has been primarily lateral and can indicate periods during which the limit to coral growth has been temporarily raised or lowered, preserved as undulations on the upper surface of the colony (Plate 81).

Plate 80 Microatolls on the reef flat in the Cocos (Keeling) Islands, Indian Ocean. The two large massive corals in the foreground comprise colonies of *Porites* that are no longer living on top, but which have live polyps confined to the margin. Their upper surface is constrained by exposure during lowest tides, and the concentric rim around the margin records a time at which this water level was slightly higher. The microatolls are approximately a metre in diameter and have been growing for several decades

Microatolls have been particularly important in the reconstruction of mid and late Holocene SEA-LEVEL change. The significance of microatolls was recognized during the 1973 Royal Society expedition to the Great Barrier Reef, northeastern Australia. Microatolls were surveyed on the surface of several different reefs and their growth form was used to indicate that sea level during the mid-Holocene had been higher than present (Scoffin and Stoddart 1978). Surveying of sequences of microatolls across transects along the mainland shore of Queensland, and cross-correlation using radiocarbon dating, provided convincing evidence that relative sea level had been above present level by more than a metre around 6,000 years ago, and demonstrated that it had undergone a smooth fall since that time (Chappell 1983). Much of the Indo-west Pacific reef province has experienced relative sea levels in mid and late Holocene that have been slightly above present. Microatolls have sometimes been preserved at a height presently above that of their living counterparts on reef flats in the eastern Indian Ocean, Southeast Asia, northern Australia, and across much of the equatorial Pacific Ocean, and provide evidence of sea-level change particularly on atolls (Smithers and Woodroffe 2000).

However, it is also clear that the banding structure of microatolls can preserve details of other events in the life history of the coral. In areas where storms are experienced, overturning of the colony can occur during individual storms, or microatolls may have responded to the moating of water that can occur behind boulder ramparts formed as a result of storms. In these cases their upper surface may record elevation of water level within impounded moats above that of regional

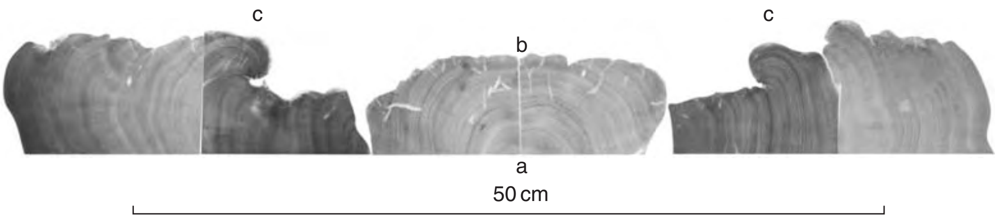

Plate 81 An X-radiograph of a vertical slice through a *Porites* microatoll. The banding indicates the growth form of the coral as the colony has aged. Initially the coral was hemispherical (a), but when it reached a level at which it was exposed too frequently during the lowest tides (b) it ceased upward growth and has extended laterally. Undulations on the surface, which occur symmetrically about the centre of the coral (c), record periods during which this upper limit to coral growth has been slightly higher

sea level. Although in open-water situations microatolls can enable centimetre-scale reconstructions of former sea level, they are subject to misinterpretation if moating has occurred, and it is therefore important to assess the geomorphological setting within which these corals grew before using fossil specimens to draw conclusions about past sea levels.

Fossil microatolls are often more accessible across the Indo-west Pacific reef province than massive hemispherical corals which are likely to have become buried by subsequent reef growth. Microatolls can be sampled along their horizontal growth axis in order to reconstruct a palaeoclimatological proxy record of the chemistry of surface waters during mid and late Holocene. Within living microatolls from the central Pacific Ocean interannual variations in water level indicate patterns of sea-level variation associated with EL NIÑO EFFECTS. Modern microatolls also enable oxygen isotope analyses of their skeleton preserving an important proxy of sea surface temperature which varies in association with the El Niño-related oscillations of sea level (Woodroffe and Gagan 2000). The application of these techniques to fossil microatolls offers the prospect of an insight into the palaeoclimatology of surface waters of more extensive areas within the tropical Indian and Pacific Oceans.

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SEE ALSO: atoll; reef

COLIN WOODROFFE

MICROMORPHOLOGY

The investigation of sediment and soil sequences has relied heavily on characterization of field morphological properties supplemented by data derived from laboratory analysis. Micromorphology, also called micropedology, extends this approach to the microscopic scale through the collection and analysis of undisturbed samples. The manufacture of soil thin sections involves the drying of samples (by air and oven drying or acetone replacement or freeze-drying), impregnation of resin under reduced pressure and curing of blocks, and sawing, lapping and mounting on glass slides. The final results are thin sections *c*.30 μm thick, which can be examined using a petrological microscope. It is possible to examine in thin sections the nature and spatial arrangement of such features as rock fragments, individual mineral grains, weathering products, void space, coatings of clay, organic material, precipitation of carbonate and excrement from soil animals.

W.L. Kubiena is widely regarded as the father figure to soil micromorphology and is best known for his book *The Soils of Europe* (Kubiena 1953). He very much established micromorphology's contribution to soil classification and our understanding of soil formation processes. Other benchmark contributions have been by Brewer and Sleeman (1988) in terms of the first attempt to systematize terminology, FitzPatrick (1993) for pedological interpretation and Bullock *et al*. (1985) for devising an international description system. The provision of full descriptions is extremely instructive since the investigator has to systematically examine all thin section attributes. Basic concepts from the international descriptive system are summarized in Tables 28 and 29. The application of micromorphology can be illustrated by summarizing some examples from soil science, archaeology and Quaternary science.

The application of micromorphology and its associated techniques provide distinctive insights into soil processes as induced by physical, biological and land use mechanisms (Miedema 1997). It has been used to determine the effects of different cultivation techniques on soils (Drees *et al*. 1994) and the interactions of soil and the soil biota (Kooistra 1991). As one example, Davidson *et al*. (2002) investigated the impacts of fauna on an upland grassland soil using micromorphological analysis. The incidence of eight different types of excrement in upper soil horizons was quantified using point counting. In both the organic horizon (H) and the underlying organo-mineral horizon (Ah), the bulk of the soil volume consisted of excrement derived primarily from enchytraeids and earthworms. Micromorphological analysis was thus able to demonstrate that the organic matter in these horizons was primarily derived from a limited range of soil fauna.

The early work on micromorphology as applied to archaeology very much focused on the palaeoenvironmental interpretation of buried soils as a means of providing environmental contexts for archaeological sites. This remains a key concern, but increasing attention is being given to site taphonomy and the wider impact of anthropogenic activity on soil landscapes. Courty *et al*. (1994) highlight the ways by which micromorphology can assist with understanding the relationships between environment and human

Table 28 Terminology in micromorphology

Soil fabric
Total organization of a soil as expressed by the spatial arrangements of the soil constituents (solids and voids)

Soil structure
Size, shape and arrangement of primary particles and voids in both aggregated and nonaggregated material and size, shape and arrangement of any aggregates present

Soil microstructures
Structures evident at magnifications $>\times 5$

Coarse and fine material
A division between coarse and fine material – division e.g. at 10 μm or 2 μm

Basic components
These are the simplest mineral and organic particles seen in thin section. They form the building blocks to the soil organization

Groundmass
General term to describe the coarse and fine material which forms the base material of soil

Micromass
General term for finer material

Pedofeature
A distinct fabric unit; stands out in contrast to adjacent soil material, e.g. clay coating in void, nodules of iron oxides

Table 29 Descriptive framework for micromorphology

1 *Structure*

2 *Groundmass*
Subdivided into coarse material and micromass

3 *Organic components*
(a) plant residues (>5 cells connected in original tissues)
(b) organic fine material (<5 cells and includes amorphous components)
(c) organic pigment staining – whole or part of micromass

Excrement pedofeatures can also be described under this heading

4 *Textural concentration features*
Features associated with an increase in concentration of material of a particular size, e.g. coatings, infillings, cappings

5 *Amorphous concentration features*
Appear amorphous in plane polarized light (PPL), isotropic in cross polarized light (XPL). Three types identified under oblique incident light are:
(a) white or dark brown colours – organic components
(b) black to yellowish brown – oxides and hydroxides of manganese
(c) yellowish brown to reddish brown – oxides and hydroxides of iron

These occur as e.g. nodules, segregations (e.g. mottles) which are impregnative

6 *Crystalline concentration features*
E.g. features consisting of Fe oxides, gypsum, gibbsite

behaviour. They summarize intra-site analysis (anthropogenically structural transformations, animal effects, anthropogenic deposits and spatial and temporal variability) and off-site analysis (land use practices and human-induced soil alteration and effects on landscape dynamics). As an example, they list micromorphological features and fabrics associated with such land use practices as slash and burn, up-rooting, ploughing, manuring, irrigation, horticultural practices and pasturing-herding. Thus ploughing, for example, results in fragmented slaking crusts, dusty silty clay intercalations and coatings, mixing of horizons, loss of the fine fraction, decrease in biological activity and changes in biological fabrics.

Micromorphology has contributed to Quaternary science, primarily through the investigation of buried soils. As an example, Zárate *et al.* (2000) demonstrate the particular contribution that micromorphological analysis can make to the investigation of a 5 m Holocene alluvial record in Argentina. Changes in the ratio of shells:diatoms:non-bioclastic coarse materials (e.g. quartz, feldspars) are used to propose different modes of deposition over time, for example the upper aeolian and alluvial units are dominated by non-bioclastic coarse particles. Two palaeosols are distinguished in thin sections by the presence of black isotropic, partially degraded root fragments, derived from the original vegetation cover. Microstructural changes down the section indicate the extent of pedogenic modification of the sedimentary fabric. The presence of partially welded excrements in the palaeosols indicates the former effect of soil fauna. Overall, the field micromorphological data provide the basis to a pedosedimentary reconstruction for the Holocene in this site in the Argentinian pampas. An alternative approach is to investigate soil formation on surfaces of different age. Srivastava and Parkash (2002) demonstrate the polygenetic nature of soils on the Gangetic Plains through micromorphological analysis of samples collected from surfaces ranging in age from 135,000 to <500 BP. Of particular interest was the degradation of early clay pedofeatures by bleaching, loss of preferred orientation and the development of a coarse speckled appearance and fragmentation; in contrast, clay pedofeatures from more recent soils were thick, smooth, strongly birefringent and microlaminated.

Kemp (1998) overviews the contribution of micromorphology to palaeopedological research and he stresses the importance of relying on a combination of such features as channels, faunal excrements, calcitic root pseudomorphs, and illuvial clay coatings. He highlights fundamental challenges posed by the polygenetic nature of soils and the fact that in contemporary soils, attributes may not be in equilibrium with current environmental conditions. This leads him to discuss the equifinality problems – the same end result can come from varying combinations of processes. He illustrates this with reference to argillic horizons; in Quaternary studies the traditional view is to regard the accumulation of illuvial clay as occuring under temperate (interglacial), seasonally dry climates under stable forest cover. Kemp (1998) argues that such illuviation can occur under a range of environmental conditions and thus it would be erroneous to propose a particular palaeoenvironmental condition because of evidence of translocated clay.

In summary, micromorphological analysis can yield distinctive information on past processes of soil and sediment formation. The continuing development of image analytical techniques is encouraging more quantitative approaches. However, research using micromorphological analysis is usually combined with other approaches, for example, soil physical and chemical analysis, pollen analysis, magnetic susceptibility, organic geochemistry or microprobe analysis.

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DONALD A. DAVIDSON

MILITARY GEOMORPHOLOGY

Military geomorphology refers to the application of geomorphic concepts, principles and technologies to military operations. This subfield links geomorphology and military science. Traditionally, military geomorphology is viewed from a perspective of the powerful influences terrain morphology has on military operations (see Winters 1998). Comparatively little attention is given to the profound effects armed conflict has on the physical landscape.

Warfare causes rapid and widespread terrain alteration. Physical landscape modification by munitions, intense vehicular movement, construction of obstacles and fortifications, and deliberate destruction are common consequences of war. Some of these activities leave erosional or depositional landforms similar to natural processes (see EQUIFINALITY) (Table 30).

Munitions and vehicle manoeuvers can alter the upper soil profile, destroy vegetation and change natural drainage patterns. These effects may persist for decades. The clay-rich landscape near Verdun, France for example, remains a pockmarked anthropogenic surface resembling GILGAI, CRATERs, DRUMLINs and HUMMOCKs nearly a century after artillery pounded the terrain during the First World War. Military defensive structures including castle moats, tank ditches, trenches and bunkers remain on the physical landscape long after their military usefulness is past. CAVEs built or modified for military purposes such as in the Tora Bora region of Afghanistan or Gibraltar, remain an intricate part of contemporary landscape morphology.

Deliberate destruction of terrain is not uncommon in war. Governments in conflict, including the Russians in 1812 and 1941–1942, and the

Table 30 Some possible geomorphic effects of military operations

Military activity	Possible geomorphic effects	Example
Vehicular movement	Causes COMPACTION OF SOILS, decreases soil infiltration rates, destroys vegetation and changes erosion and deposition patterns. Forms RILLS, gullies, WADIS, ARROYOs, and track scars.	Tank track scars are preserved in desert pavement of the southwestern United States over 60 years.
Use of artillery, bombs, minefields	Creates blast craters, alteration of the upper soil profile, and destruction of vegetation with subsequent changes to erosion and deposition patterns. Mined areas may remain relatively unchanged by further anthropogenic alteration.	Over 20 million blast craters were produced during the Vietnam conflict. Over 100 million mines remain from conflict in more than 90 countries worldwide.
Construction of bunkers, trenches, defensive fortifications	May form mounds, gullies, wadis, arroyos, moats, CAVEs and canals.	Mounds marking defensive fortifications remain along the Normandy coast of France from the Second World War.

United States Union Army during the American Civil War, employed 'scorched earth' tactics. Crops, vegetation and structures were purposefully destroyed to deny the enemy their use, generating changes to erosion and deposition patterns across large areas. In the Second World War, Allied forces destroyed dams on the River Ruhr in Germany, drastically changing downstream fluvial morphology. Millions of gallons of defoliants devastated tropical rainforest and cropland soils during the Vietnam conflict. In 1991, retreating Iraqi forces set fire to over 730 oil wells in Kuwait, creating oil lakes and a durable 'tarcrete' surface of petroleum sludge, an artificial DURICRUST.

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SEE ALSO: anthropogeomorphology

DANIEL A. GILEWITCH

MIMA MOUND

Also called prairie mounds and pimple mounds, Mima mounds take their name from Mima Prairie, Thurston County, Washington, USA. Such mounds are characteristically up to around 2 m in height, 25–50 m in diameter, and occur at a density of 50–100 or more to the hectare. There are many hypotheses for their origin (Cox and Gakahu 1986), including that they are erosional residuals, result from depositional processes around vegetational clumps, are the product of frost sorting, have been formed by communal rodents, are degraded termitaria, or have been created by seismic activity or groundwater vortices (Reider *et al*. 1996).

They are found from the Gulf Coast to Alberta, while in southern Africa, where they are termed *Heuweltjies*, they are widely distributed in the drier, western parts (Lovegrove and Siegfried 1989). Similar forms are also known from Argentina and Kenya. These mounds probably have many different origins, but the role of such beasts as mole rats, prairie dogs and gophers should not be underestimated.

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A.S. GOUDIE

MINERAL MAGNETICS IN GEOMORPHOLOGY

Mineral magnetic (or environmental magnetic) analysis provides a means of characterizing soil, sediment or rock samples on the basis of their magnetic properties. As demonstrated by rock magnetic research (Dunlop and Özdemir 1997), soils, unconsolidated sediments and solid rocks all display magnetic properties. These properties can be quantified to reveal information about the types of magnetic minerals present, their concentration and, in some circumstances, their magnetic grain size. The method shares many of the underlying principles of other methods of material characterization such as size, shape, colour, mineralogical or chemical composition and, for the geomorphologist, therefore offers a similar range of potential applications.

The magnetic behaviour of materials can be broadly classified into three types: diamagnetic (i.e. quartz, feldspar, water), paramagnetic (i.e. olivine, pyroxene, biotite) and ferromagnetic (i.e. magnetite, haematite). The first two types are relatively weak phenomena. While such materials produce a measurable magnetic response in the presence of an artificial external magnetic field (due to changes induced in electron motions within the constituent atoms), they are not capable of holding a remanence of that field. Thus, when the field is removed, the electron motions return to their previous behaviour and their magnetic properties cancel, resulting in no net spontaneous magnetization in the material.

In contrast, ferromagnetic materials display much stronger magnetic response in the presence of a magnetic field and can, in some circumstances, retain a memory (remanence) of that field after it is removed (spontaneous magnetization). Ferromagnetic behaviour can, in turn, be classified into sub-types. Of most interest in a geomorphological context are ferrimagnetic (i.e. magnetite, maghemite, greigite) and imperfect (canted) antiferromagnetic (i.e. haematite, goethite) types. These two groups show contrasting magnetic behaviour when subjected to specific laboratory measurements, although their magnetic response also varies with their exact composition and grain size. In addition, many natural samples will contain assemblages of mixed mineral types, concentrations and grain sizes, resulting in potentially complex 'bulk' magnetic behaviour.

Routine measurements (Table 31) are made at room temperature and provide (a) an insight into the magnetic mineralogy, concentration and grain size within a sample and (b) a characterization (or fingerprint) of the sample. In more advanced studies, temperature dependent magnetic properties may also be measured as these may provide more conclusive evidence in terms of (a) above. The advantages of mineral magnetic techniques include the ease of measurement, the ability to process large numbers of samples relatively quickly, their non-destructive nature (for room and low temperature measurements at least) and relatively low cost. However, the most significant advantage is the sensitivity of the instrumentation. In the majority of cases, differences in iron oxide concentrations can be detected that are well below the resolution of other methods such as X-ray diffraction, differential thermal analysis or differential chemical extractions.

Mineral magnetic analysis has a range of applications in studies of the environment (Table 32). In a geomorphological context, two such applications have received particular attention: (1) studies of WEATHERING/pedogenesis and (2) sediment tracing. The alteration and redistribution of iron that takes place within the soil environment is often reflected in corresponding changes in magnetic properties and certain soil types may show diagnostic variations in their magnetic signature with depth (Maher 1986). A common phenomena is the 'enhancement' of topsoil magnetic properties (e.g. higher values than the subsoil), where fine-grained magnetite and maghemite are formed as a result of biogeochemical transformations of iron weathered from other minerals (Dearing 2000). Like other changes induced by weathering and/or pedogenesis, all other things being equal, the degree of alteration of the REGOLITH may increase with time (although not necessarily in a linear fashion). In some circumstances therefore, the level of magnetic alteration may be used as a relative dating method (e.g. White and Walden 1994; Walden and Ballantyne 2002).

Differentiation of topsoil and subsoil materials on the basis of their magnetic properties has been widely used in sediment tracing studies

Table 31 Common room-temperature mineral magnetic parameters and their basic interpretation

Parameter	Interpretation
χ ($10^{-6}\,m^3\,kg^{-1}$)	Initial low field mass specific magnetic susceptibility. This is measured within a small magnetic field and is reversible (no remanence is induced). Its value is roughly proportional to the concentration of ferrimagnetic minerals within the sample.
χ_{fd} ($10^{-6}\,m^3\,kg^{-1}$)	Frequency dependent susceptibility. This parameter measures the variation of magnetic susceptibility with the frequency of the applied alternating magnetic field. Its value is proportional to the amount of magnetic grains whose size means they lie at the stable single domain/ superparamagnetic ($<0.1\,\mu m$) boundary.
χ_{ARM} ($10^{-6}\,m^3\,kg^{-1}$)	Anhysteretic Remanent Magnetization (ARM) is proportional to the concentration of ferrimagnetic grains in the 0.02 to 0.4 μm (stable single domain) size range. The final result can be expressed as mass specific ARM per unit of the steady field applied (χ_{ARM}).
SIRM ($10^{-6}\,Am^2\,kg^{-1}$)	Saturation Isothermal Remanent Magnetization (SIRM) is the highest amount of magnetic remanence that can be produced in a sample by applying a large magnetic field. The value of SIRM is related to concentrations of all remanence-carrying minerals in the sample but is dependent upon the assemblage of mineral types and their magnetic grain size.
Soft IRM ($10^{-6}\,Am^2\,kg^{-1}$)	The amount of remanence acquired by a sample after experiencing an applied field of 40 mT. At such low fields, the high coercivity, canted antiferromagnetic minerals such as haematite or goethite are unlikely to contribute to the IRM, even at fine grain sizes. The value is therefore approximately proportional to the concentration of the low coercivity, ferrimagnetic minerals (e.g. magnetite) within the sample, although also grain-size dependent.
Hard IRM ($10^{-6}\,Am^2\,kg^{-1}$)	The amount of remanence acquired in a sample beyond an applied field of 300 mT. At fields of 300 mT, the majority of ferrimagnetic minerals will already have saturated and the value is therefore approximately proportional to the concentration of canted antiferromagnetic minerals within the sample.
IRM backfield ratios	Various magnetization parameters can be obtained by applying one or more magnetic 'reverse' or 'backfields' to an already saturated sample. The magnetization at each backfield can be expressed as a ratio of IRM_{field}/SIRM and can discriminate between ferrimagnetic and canted antiferromagnetic mineral types.

Sources: After Thompson and Oldfield (1986); Maher and Thompson (1999); Walden *et al*. (1999)

within lake and fluvial sediment systems (e.g. Dearing *et al*. 1985; Dearing 2000) and at a catchment scale, soils based upon different parent lithologies can also be distinguished. Considerable potential also exists for magnetic properties to be used in studies of SOIL EROSION and redistribution on hillslope systems, where they may complement other methods such as artificial radionuclides. Despite the advantages of the method, the user must also be aware of the underlying assumptions and potential problems. Two key issues are (1) the ability to identify and fully characterize the variability of the potential sediment source types and (2) the validity of assuming that the magnetic properties remain unaltered during sediment transport and deposition (Dearing 2000).

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Table 32 Environmental applications of mineral magnetic analysis

Application	Sedimentary environment	Example
Sediment correlation	Lacustrine, glacial, loess, fluvial, marine, soil erosion, etc.	Correlation of Heinrich layers in N. Atlantic sediments.
Tracing sediment provenance	Lacustrine, glacial, loess, fluvial, marine, soil erosion, etc.	Source areas of glacial sediment sequences. Soil erosion into river/lake sediment systems. Catchment fire histories.
Weathering/soil forming processes	Contemporary soils, colluvial deposits, PALAEOSOL identification, ALLUVIAL FAN surfaces.	Relative dating of weathered/pedogenic surfaces. Quaternary climate change record in Chinese LOESS sediments.
Artificial tagging of sediment	Fluvial, estuarine.	Tracing movement of fluvial sediments.
Pollution monitoring	Recent organic sediments, urban drainage, atmospheric pollution.	Industrial emissions from coal-fired power stations.

Sources: After Thompson and Oldfield (1986); Maher and Thompson (1999); Walden *et al*. (1999); Dearing (2000)

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SEE ALSO: soil geomorphology; tracer

JOHN WALDEN

MINING IMPACTS ON RIVERS

Mining for heavy metals such as lead, zinc, copper, gold and silver has affected river environments since the advent of metallurgy. Pollution issues associated with heavy metal extraction from alluvial sediment or directly from their original host rock are known to have had their greatest impact on river systems since the start of the industrial revolution, *c*.1800. The fate of metal pollutants is similar to the natural sediment load; it may be stored within the channel, on the floodplains or it can be transferred out of the system via the estuaries and the ocean. It is the connectivity of river, estuarine and coastal transport systems coupled to the storage capacity of fluvial environments that determines the distribution of such sediment borne metals in a catchment. A basin-wide assessment of the fate and storage of mining related metal contaminants in the north-east of England (Macklin *et al*. 2000) indicated that only a small proportion has been flushed out into the Humber estuary with the remainder being stored in and along associated river systems.

Because of the protracted residence times of heavy metals within rivers and their floodplains (from 10^1 up to 10^4 years), metal-contaminated sediment may act as major sources of future contamination. Normal channel erosion and sedimentation processes are important in redistributing such contaminants across floodplains and for moving any inchannel material downstream. Fluvial sediments are spatially and temporarily complex, reflecting changes in the frequency and magnitude of river behaviour (erosion and sedimentation) to either extrinsic (e.g. climate or land use changes) or intrinsic (e.g. threshold adjustments) impacts. The distribution of metals in a river system is similarly complex. In more homogenous fluvial environments, it is often possible to determine pre-mining, mining era and post-mining era sediments through the examination of vertical overbank sequences. Here the deepest, oldest units may contain a geochemical imprint of the catchment prior to human disturbance. Moving up through the sediment profile one might encounter changes in sediment metal values that reflect the impact of human activity in a catchment. However, several studies (e.g. Macklin *et al*. 1994; Taylor 1996) have revealed that the imprint of mining activity may not be so simply distributed in floodplain sediments. Pre- and post-industrial anthropogenic deforestation may disrupt the predicted geochemical profile of alluvium either through dilution of the contaminant signal or through the erosion and transfer of sediments from a mineralized catchment that is naturally enriched in heavy metals. Floodplain geomorphology, such as terraces, levees, back channel environments, lakes and cut-offs may also assert a major control on the distribution of metals within floodplain environments such that overbank sediment profiles may show considerable variation in heavy metal concentrations both laterally and vertically.

The transfer of metals in rivers is dominated by four major mechanisms (Lewin and Macklin 1987): (1) hydraulic sorting according to individual particle size and mineral density; (2) chemical dispersal – solution, adsorption, the formation of Fe and Mn complexes and organic uptake (bioaccumulation in plants and animals); (3) dilution with clean uncontaminated sediments; (4) loss and exchange with floodplain sediment. Chemical remobilization and dispersal remains a significant problem particularly with respect to acid mine drainage where low pH values and changes in redox (i.e. reducing) conditions can liberate co-precipitated or adsorbed metals from relatively stable mine spoil into adjacent water and sediment bodies. Changes in water and sediment chemistry can increase the solubility, mobility and bioavailability of metal-contaminated sediment and thus make them more deleterious to the surrounding environment.

The transport of mining waste in river systems may occur in one of two modes: ‘passive dispersal’ or ‘active transformation’ (Lewin and Macklin 1987). In the ‘passive dispersal’ of riverborne metals, the river system remains in equilibrium and the waste is transported alongside the natural sediment load such that there is no significant alteration to the channel and floodplain’s morphology.

‘Active transformation’ is associated with a greatly increased sediment load such that it results in a major transformation of the types, rates and/or magnitudes of geomorphic processes that control the prevailing channel morphology (Miller 1997). Gilbert’s (1917) seminal paper described the effects of hydraulic gold mining in the Sierra Nevada (USA) on the tributaries of the Sacramento River between 1855 and 1884. Gilbert (1917) explained how mining waste resulted in rapid rises in the elevation of tributary channel beds during and following the cessation of mining. Aggradation was subsequently replaced by incision when the supply of mining debris declined and sediment was transferred downstream in wave-like form over time. However, the rate of incision (recovery) was markedly slower than that of aggradation (James 1993).

Other impacts of mining debris include the phytotoxic damage of riparian vegetation leading to bank destabilization, changes in sediment supply to the channel zone and ultimately planform metamorphosis from a meandering to a braided channel planform. In the Tasmanian Ringarooma River basin, tin mining during the nineteenth and early twentieth centuries (Knighton 1989, 1991) caused major changes to the width and planform of the Ringarooma River. The channel bed aggraded up to 10 m and channel width increased by up to 300 per cent. The response in the Ringarooma basin was very different to that on the Sacramento River because the source of the mining debris was much more diffuse, with mining sediments being distributed along the length of the river system as opposed to having a more geographically limited and discrete point source. Although the aggradation–incision cycle was evident in the Ringarooma and progressed downstream along with the mining debris sediment wave, the spatial pattern was highly variable due to the input of debris from tributaries all along the system.

The human imprint of land use change may manifest itself in many forms within a river basin. Those related to the direct impact of mining on rivers can result in the active transformation of a channel. This may cause major and highly visible disruption to the physical structure of fluvial environments through bed level adjustments, and cross-sectional and planform changes following the release of substantial volumes of toxic materials directly into the system. Less visible impacts such as the storage of heavy metal contaminants within a riverbed and on floodplains can result in the long-term storage of contaminants long after the primary pollution activity has dissipated. These may provide a latent but potentially insidious secondary source of pollution for future adjacent agricultural and urban land uses.

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SEE ALSO: alluvium; floodplain; fluvial erosion quantification; fluvial geomorphology; sedimentation; threshold, geomorphic

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MIRE

Definition of the term mire is not straightforward because mires may exist along a continuum from deep water aquatic systems through to terrestrial systems, and their boundaries are not easily defined and identified. Consequently, mires reflect different origins and patterns of development; are found in widely different geographical locations and under different climatic regimes; encompass different sets of controlling forces, and reflect different stages of successional development (Hofstetter 2000).

Mires are essentially peat-accumulating landscape features which, under natural conditions, are found where the water level is continuously near the soil surface resulting in a very narrow aerobic layer. These abiotic conditions favour specific plant and animal species, mosses or micro-organisms, which are adapted to wet and often nutrient poor conditions. The distribution of, and differences in, mire type, vegetation composition and soil type are caused primarily by geology, topography and climate but the formation, persistence, size and function of mires are controlled by *hydrological processes*. The source of the water, its quantity and quality and the mechanism by which it is delivered to the mire combine to influence mire development and character, giving rise to the wide spectrum of different mire types that occur in the landscape (e.g. Gore 1983; Heathwaite and Gottlich 1993; Moore 1984). Indeed, Mitsch and Gosselink (1993) went so far as to state that: 'Hydrology is probably the single most important determinant of the establishment and maintenance of specific types of mires and mire processes.'

There is no uniquely correct way of classifying mires because: (1) they are characterized by continuous gradation of properties with variable discontinuities, (2) their formation is affected by changing climatic, geomorphological and hydrological conditions, and (3) variations in mire types occur on a variety of scales. Thus it is not surprising that terms such as *bog, fen, mire* and *moor* are used widely but often imprecisely and occasionally interchangeably! The source of this imprecision lies with the range of criteria that have been used in attempts to define and classify mires. These criteria include floristic composition, site hydrology, site topography, water chemistry and nutritional status, and peat structure. Such criteria are sometimes used separately and sometimes in combination. The International Mire Conservation Group (www.imcg.net) suggest the following priority characteristics in differentiating mire types:

1 Source of water,
2 Prevailing hydrology-geomorphology,
3 Base content (saturation) or pH,
4 Nutrient availability and C : N, and
5 Prevailing plant communities.

Some of these classification characteristics are examined in more detail below.

Mire hydrology

The one common ingredient of all natural mires is an 'excess' of water, or at least a hydrological balance adequate to create conditions in which the surface is usually waterlogged for at least part of the year. The wet conditions characteristic of mires may result from impeded drainage, high rates of water supply or both. Water supply may consist of telluric water (i.e. water that has had some contact with mineral ground such as river water, surface runoff or groundwater discharge) or meteoric water (i.e. precipitation). Hydrological relationships play a key role in mire ecosystem processes, and in determining structure and growth. Thus different mires have a characteristic hydroperiod, or seasonal pattern of water levels, that defines the rise and fall of surface and subsurface water. An important geoindicator is the water budget of a mire, which links inputs from ground water, runoff, precipitation and physical forces (wind, tides) with outputs from drainage, recharge, evaporation and transpiration. Annual or seasonal changes in the range of water levels affect visible surface biota, decay processes, accumulation rates and gas emissions. Such changes can occur in response to a range of external factors, such as fluctuations in water source (river diversions, groundwater pumping), climate or land use (forest clearing). Waters flowing out of mires are chemically distinct from inflow waters, because a range of physical and chemical reactions take place as water passes through organic materials, such as peat, causing some elements (e.g. heavy metals) to be sequestered and others (e.g. dissolved organic carbon, humic acids) to be mobilized.

Because mire vegetation is largely responsive to primary environmental factors, such as hydrological and hydrochemical factors, hydrological classifications such as the one in Table 33 for British mires (see Heathwaite 1995) often offer the most direct explanation of mire types, because form and biotic characteristics are determined by these features.

Mire morphology

It has long been recognized that differences in topographic situation and water supply

Table 33 Hydrological classification for British mires

Source of water	Extent		
	Small (<50 ha)	Medium (50–1,000 ha)	Large (>1,000 ha)
Rainfall	Parts of basin mires	Raised mires	Blanket mires
Springs	Flushes, acid valley and basin mires	Fen basins, acid valley and basin mires	Fen massif, The Fens, Somerset Levels
Floods	Narrow floodplains	Valley floodplains	Floodplain massif

mechanism profoundly influence mire type. Such considerations formed the basis of the early but long-standing systems of mire classification. Von Post and Granlund (1926) subdivided mires into three types: ombrogenous mires, developed under the exclusive influence of precipitation; topogenous mires, irrigated by telluric water which naturally collects on flat ground and topographic hollows; and soligenous mires developed on slopes and kept wet by a supply of telluric water. Moore and Bellamy (1974) used this principle to subdivide British mires into broad groups based on their physiography which they justified on the basis that mire development is mainly determined by climatic, hydrological and geochemical conditions that control vegetation communities, leading to development of a particular mire type (Table 34).

Mire development

Viewed statically, mire types reflect a water table zonation from more or less open water through to conditions where the water table is rarely if ever above the substrate surface, though rarely far below base. Viewed dynamically, mires reflect successional or hydroseral changes. In the UK, it is difficult to demonstrate different successional stages because most hydroseres began developing at about the same time in the postglacial (Flandrian) period and are often, therefore, at similar successional stages. However, palaeoreconstruction based on the stratigraphy of accumulated peat deposits can be used to demonstrate the dynamic features of mire development over time.

The natural progression of the autogenic mire succession is in the direction of increasing acidity as the growing peat surface becomes progressively isolated from the nutritional effect of ground and soil water and more dependent on rainwater nutrition. Geomorphological criteria, principally climate, topography and substrate geology, are fundamental in directing changes to this autogenic succession. Thus the zonation of topogenous mires characteristically follows the water table gradient round enclosed basins or hollows which concentrate flow and allow the accumulation of a peat where lateral water movement is impeded. Soligenous mires develop where slow lateral gravitational seepage maintains waterlogged conditions at the ground surface. Topography is still important but it is the slow percolation of water through this mire type that distinguishes their type. Water flowing through soligenous mires is typically more oxygenated relative to the stagnant conditions of topogenous mires, and consequently the rate of organic matter decomposition is higher and the depth of peat accumulation lower. Ombrogenous mires develop where precipitation is high relative to evapotranspiration. Topography, whilst important, largely acts to retard runoff from the mire, rather than to concentrate runoff to it from other areas. The dependence on atmospheric inputs alone produces mire habitats that are characteristically of low base status, and where organic matter decomposition is low and the accumulation of unhumified acid peat high relative to other mire types. In the UK, ombrogenous mires are subdivided into raised mires and blanket mires or bogs. Blanket mires develop where the ground surface is

Table 34 Physiographical classification for British mires

Soligenous mires	Moving water/flushes/springs, slow peat formation (due to O_2)
Basin mires	Found in deep hollows, e.g. kettleholes; deep peat; vegetation surface may float; limited groundwater movement
Valley mires	Characterized by water flow along valley axis; broad range in pH, nutrients and vegetation communities
Floodplain mires	Develop on flood-prone alluvium; broad range in pH, nutrients and vegetation communities
Raised mires	Characterized by a peat surface that is isolated from the regional groundwater table; ombrotrophic; domed shape
Blanket mires	Usually develop on impermeable materials in regions with high precipitation and low temperature

permanently wet, initiating peat formation on flat and gently sloping ground. Raised mires occur over a wide range of climatic conditions and may represent a late stage in the autogenic succession of topogenous mires. In lowland Britain, raised mires are recorded in basins, floodplains and at the heads of estuaries, for example Thorne Moors National Nature Reserve which forms part of the Humberhead Peatlands. They are characterized by a raised central mire area where the peat has accumulated to the extent that it becomes isolated from water feeding the mire margins to become solely dependent on rainfall inputs. Here acidification ensues, the rate of decomposition falls, peat accumulation increases, and the mire type shifts from topogenous or soligenous mires to ombrogenous types. Raised mires form characteristic shallow domes of peat where the topography is typically convex, with a gently sloping rand away from its centre towards the surrounding moat-like drainage channel or lagg surrounding the bog.

Mire hydrochemistry

In addition to hydrological controls on mire development, the base and nutrient status of the mire water supply influence the mire type. The water chemistry of mires is primarily a result of geologic setting, water balance (relative proportions of inflow, outflow and storage), quality of inflowing water, type of soils and vegetation, and human activity within or near the mire. Mires dominated by surface-water inflow and outflow reflect the chemistry of the associated rivers or lakes. Mires that receive water primarily from precipitation and lose water by way of surface-water outflows and (or) seepage to ground water tend to have lower concentrations of chemicals. Thus ombrotrophic mires are rainwater-dominated and consequently base-deficient whereas minerotrophic mires are supplied with minerals and nutrients via the mire substrate which is in turn dependent on catchment geology and drainage water quality. Minerotrophic mires may range from oligotrophic through to mesotrophic or eutrophic types depending on the quality of their source water. Thus hydrochemical mire classifications focus largely on the source and quality of water to a mire giving a range of mire types from ombrotrophic raised mires, through transitional mires to minerotrophic mires or fens (Table 35).

On the basis of floristic variation in Swedish mires, Du Rietz (1949, 1954) suggested that mires could be divided into areas fed almost exclusively by precipitation and those in which water supply was supplemented by telluric water. These early concepts are important because they broadly correspond with major habitat differences still recognized today. The term fen is largely used as a synonym for minerotrophic mires and bog to refer to ombrotrophic examples (see Wheeler and Proctor 2000). These colloquial terms are still confused however, particularly as the vegetation of bogs and fens can be very similar. The distinction between these two habitats is based on their respective water sources. Bogs obtain their water from rainfall alone and this water is essentially stagnant, at least in the lower bog layers or catotelm. The water in a fen flows, although this may happen very slowly. Joosten (1998) used base conditions and trophic status to differentiate mire types, see Table 36.

The nutrient supply to bogs is characteristically low although nitrogen may be supplemented by

Table 35 Mean values of the concentration of major ions in waters from European mires

Mire hydrochemistry		Major ions pH	HCO_3	Cl	SO_4	Ca	Mg	Na	K	H	Total
Eutrophic ↓	1	7.5	3.9	0.4	0.8	4.0	0.6	0.5	0.05	0	10.25
	2	6.9	2.7	0.5	1.0	3.2	0.4	0.4	0.08	0	8.28
	3	6.2	1.0	0.5	0.7	1.2	0.4	0.5	0.02	0	4.32
	4	5.6	0.4	0.5	0.5	0.7	0.2	0.5	0.04	0.01	2.85
	5	4.8	0.1	0.3	0.5	0.3	0.1	0.3	0.07	0.03	1.70
	6	4.1	0	0.4	0.4	0.2	0.1	0.3	0.04	0.14	1.58
	7	3.8	0	0.3	0.3	0.1	0.1	0.2	0.04	0.16	1.20
Oligotrophic											

Source: After Moore and Bellamy (1974)

Table 36 Base conditions, trophic conditions and mire types

C/N ratio	> 33				< 20
pH		< 4.8		> 6.4	■
	Oligotrophic acid	Mesotrophic acid	Mesotrophic subneutral	Mesotrophic calcareous	Eutrophic
Lowland bog	■				
Mountain bog	■				
Kettlehole mire	■	■			■
Percolation mire		■	■	■	
Surface flow mire		■	■		■
Terrestrialization mire		■	■	■	■
Spring mire		■	■	■	■
Coastal floodplain mire					■
Fluvial floodplain mire					■

Source: After Joosten (1998)

atmospheric enrichment from industrial, urban and agricultural sources. Bog biodiversity is low. Typically the pH is <4.5 compared to fens where the pH range is 4.5–7.5.

Significance of mires in the landscape

In western Europe the majority of natural mires have been degraded through anthropogenic changes in hydrology, both at the regional and local scale, primarily for agricultural purposes. These measures have affected the biotic composition, the soil physical and chemical properties, the carbon and nutrient dynamic, as well as the landscape ecological functions of mires.

The regulatory function of hydrologically undisturbed mires compared to degraded mires has been neglected until recently. Natural mires act as ecotones between terrestrial and aquatic environments and are important owing to their transformation, buffer and sink qualities. For example, minerotrophic mires or fens are connected with their surrounding terrestrial areas via several hydrological pathways such as groundwater inflow, surface runoff, interflow or river water surplus. Nutrients transported with the inflowing water into such mires are transformed or accumulated by several biogeochemical processes. As a result, the nutrient concentration in the outflow can be reduced and water quality improved. Thus lowland mires are often areas of high biological productivity and diversity and mediate large and small-scale environmental processes by altering downstream catchments. For example, lowland mires can affect local hydrology by acting as a filter, sequestering and storing heavy metals and other pollutants, and serving as flood buffers and, in coastal zones, as storm defences and erosion controls. Upland mires can act as a carbon sink, storing organic carbon in waterlogged sediments. Even slowly growing peat may sequester carbon at between 0.5 and 0.7 tonnes ha^{-1} a^{-1}. Mires can also be a carbon source, when it is released via degassing during decay processes, or after drainage and cutting, as a result of oxidation or burning. Globally, upland mires have shifted over the past two centuries from sinks to sources of carbon, largely because of human exploitation.

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Key websites

RAMSAR: http://www.ramsar.org
Irish Peatland Conservation Council:
http://www.ipcc.ie
Society of Mire Scientists: http://www.sws.org
International Peat Society: http://www.peatsociety.fi
British Ecological Society Mires Research Group: http://www.britishecologicalsociety.org/groups/mires/index.php

LOUISE HEATHWAITE

MOBILE BED

A fluid, such as air or water, flowing over cohesionless sediment has the ability to entrain solid particles. The bed surface becomes mobile when the shear stress applied on the particles by the flow exceeds the critical shear stress of the sediment mixture. The initiation of particle motion is a stochastic phenomenon that depends on the average fluid motions and, because the dimensions of sediment particles usually are relatively small compared to the dimensions of the flow, on the magnitude of turbulent deviations from the average (Nelson *et al*. 2001). It also depends on the position of a particle on the bed, which determines its exposure to the flow (Kirchner *et al*. 1990; Buffington *et al*. 1992), and the relative proportion of each size fraction in a mixture (Wilcock 1993). The shear stress at which particle motion is initiated in heterogeneous sediment may be approximated by Shields's relation, if the median grain size is used to characterize the entire sediment mixture (Kuhnle 1993; Buffington and Montgomery 1997).

As the threshold condition for the initiation of particle motion is approached there is an abrupt increase in the rate of sediment movement. Particle movement is neither uniform nor continuous over the bed, because turbulent sweeps, the structures responsible for particle motion, move groups of particles intermittently at random locations on the bed (Drake *et al*. 1988; Williams *et al*. 1990). As the shear stress (and rate of sediment transport) increases the sweeps become more laterally stable and longitudinal streaks form on the bed, and in heterogeneous sediment a pattern of alternating coarse- and fine-grained stripes emerges (McLelland *et al*. 1999). At higher shear stresses the coarser sediment becomes more mobile and the stripes are replaced by flow-transverse BEDFORMS (Gyr and Müller 1996).

Sediment initially was thought to move in sliding layers, with the most rapidly moving layer positioned adjacent to the flow, but it was soon recognized that only the surficial grains move. In the absence of significant scour or bedform development, the depth of the active layer is of the order of 0.4 to $2D_{90}$ (where D_{90} is the size for which 90 per cent of the surficial bed material size distribution is finer). The sediment in transport is termed the bed material load. Bed material either can be swept up into the main part of the flow by turbulence, and transported in suspension, or it may move, by rolling/sliding or SALTATION, as BEDLOAD in a layer immediately above the bed. This layer is of the order of two to four grain diameters thick in water, and a few tens of centimetres thick in air. As the flow intensity increases above the critical value, particles first move by rolling. Saltation rapidly becomes the dominant type of motion as the flow intensity increases further, and at still higher flow intensities suspension begins to dominate. There is a clear physical difference between the two basic modes of transport (Abbott and Francis 1977). The weight of a saltating particle is supported by the bed, whereas the flow supports the weight of a suspended particle. However, the two modes of transport cannot easily be differentiated on the basis of particle size, and there is a continual exchange of particles between the bedload and SUSPENDED LOAD. There is also an important difference between the movement of particles by saltation, in air and in water. In air, once saltation commences subsequent movement is induced by the impact of particles hitting the bed, rather than by the hydrodynamic forces that act on static particles, as is the case in water. The difference arises because the submerged density of sediment particles is substantially greater than the density of air at atmospheric pressure, whereas it is less than twice the density of water.

Particles transported in suspension move at the velocity of the flow. Particles comprising the bedload continually move in and out of storage on the bed, and their pattern of motion can be characterized as a series of relatively short steps of random length, each of which is followed by a rest period of random duration (Habersack 2001). The sensitivity of travel distance to particle size decreases as size decreases below the median diameter of the substrate (Church and Hassan 1992), but the virtual velocity of particles in water is only of the order of metres per hour, compared to the flow velocity which may be of the order of metres per second (Haschenburger and Church 1998). This is because each particle spends a negligible time in motion compared to the time spent at rest. In the case of sediment that is deposited on the lee side of a bedform, the velocity at which the particles move is much slower and is determined by the rate of movement of the bedform (Grigg 1970; Tsoar 1974).

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BASIL GOMEZ

MODELS

'The sciences', wrote mathematician John von Neumann (1963), 'mainly make models.' Model building is as much a part of twenty-first-century geomorphology as it is in any science. A model, in its most general sense, is a simplified or idealized representation of an existing or potential reality. Examples of models range from architects' miniatures to quantum theory. In geomorphology, models serve as representations of Earth surface processes and landforms, and as such they embody the theory that underpins the science. All geomorphologists rely on models of one sort or another. However, the word 'model' is used in a variety of contexts in geomorphology, and it is useful to distinguish between three general forms: conceptual models, hardware (or experimental) models and mathematical models.

Conceptual models of landform origins must surely predate the word 'geomorphology'. As the formal scientific field of geomorphology began to take shape in the late nineteenth century, influential figures such as William Morris Davis, Walther Penck and G.K. Gilbert developed conceptual models of landscape systems that provided a guiding impetus for research. Davis's 'geographical cycle' is a classic example of a conceptual model in geomorphology. It provided an explanation for many observed landforms, made predictions about their course of evolution, provided guiding assumptions for the interpretation of particular landform elements (such as low-relief surfaces) and influenced the type of questions posed by researchers. Although many of Davis's ideas have not stood the test of time, the creation and progressive refinement of conceptual models are still fundamental parts of geomorphology. Unavoidably, our ideas about how geomorphic systems operate will always guide the type of questions we choose to ask and the kind of interpretations we make (Brown 1996).

Hardware models represent geomorphology's experimental side. A hardware model is a physical representation, often (but not always) scaled down, of a particular geomorphic system. G.K. Gilbert's (1914) flume experiments at Berkeley, which led to his classic paper 'The transportation of debris by running water', represent one of the first experimental studies in geomorphology. Gilbert's data are, in fact, still used today, and have been complemented by many other flume studies of sediment transport. The literature is replete with examples of experimental models of geomorphic systems. Phenomena that have been studied experimentally include drainage basin evolution, bedrock landsliding, soil creep, rock weathering, alluvial fans, and subaerial and subaqueous debris flows, to name a few. In some cases, laboratory experiments have coupled geomorphic processes with tectonic, eustatic, and/or depositional processes. In some instances, hardware models operate on the same spatial scale as the geomorphic system in question; the US Geological Survey experimental debris flow flume near Bellingham, Washington, USA is one such example (Major and Iverson 1999). More commonly, the physical system is scaled down, which can introduce problems in preserving basic scaling relationships between physical properties such as fluid viscosity and gravity. Nonetheless, hardware modelling continues to be an important source of information and insight into a wide range of geomorphic systems.

A mathematical model, like a conceptual model, acts as a simplified or idealized representation of reality that provides a framework for guiding and interpreting observations. Seen in this light, a mathematical model can be understood as a quantitative hypothesis or set of linked hypotheses. A mathematical model has an obvious advantage over purely conceptual models in its precision, lack of ambiguity and ability to satisfy basic constraints such as continuity of mass, momentum and energy. At the same time, mathematical models, like hardware models, allow for a degree of experimentation – in the sense of testing the behaviour of a system (one comprising a set of mathematical-logical postulates and assumptions, rather than a physical construct) that has been built as an analogy for a natural system. The use of ocean and atmospheric general circulation models to test palaeoclimate hypotheses (e.g. Cane and Molnar 2001) is a good example of this type of experimentation.

Although there is no generally agreed classification of types mathematical model, the categories suggested by Kirkby *et al.* (1992) provide a useful framework. They distinguish between black-box (statistical or empirical) models, process models, mass-balance models and stochastic models. There is significant overlap among these categories, and indeed many mathematical models combine elements of several of these.

Mathematical models in geomorphology began to emerge in the post-Second World War era. Many of these early models were descriptive or empirical (i.e. 'black box') in nature. R.E. Horton's drainage network laws (now known as HORTON'S LAWS), for example, provided a quantitative description of river network topology, while the hydraulic geometry equations of Leopold and Maddock (1953) provided a similar description of river channel changes through space and time. These and many other morphometric models are essentially statistical in nature.

Beginning in the 1960s, such statistical models were complemented by process models. Where a black-box model represents relationships in a purely empirical form, a process model attempts to describe the mechanisms involved in a system. For example, a black-box model of soil erosion would be based on regression equations obtained directly from data, whereas a process model would attempt to represent the mechanics of overland flow and particle detachment. Process models often overlap with mass-balance (or energy-balance) models, in the sense that equations for processes are used to model the transfer of mass or energy among different *stores*, where a store could represent anything from the water in a lake, the population of a species in an ecosystem, the energy stored as latent heat in an atmospheric column, the carbon mass in a tree, or the depth of soil at a point on a hillslope.

Process models have been widely used to study landform evolution. Usually phrased in terms of continuum mechanics, these landform evolution models provide a link between the physics and chemistry of geomorphic processes, and the shape of the resulting topography. Among the pioneers in landform 'process-response' modelling in the late 1960s and early 1970s were F. Ahnert and M.J. Kirkby. The latter showed, for example, that the convexo-concave form of hillslopes can be predicted from simple laws for sediment transport (Kirkby 1971).

Many process models are deterministic, meaning that for a given set of inputs they will predict a unique set of outputs. Often, however, the inputs to a particular geomorphic system are highly variable in time or space, and essentially unpredictable or unmeasurable. For example, we may know something about the frequency and magnitude characteristics of rainfall but cannot predict the sequence of rainfall events over time spans of more than a few days. Similarly, we may have a good estimate of the average hydraulic conductivity of an aquifer but little or no information about its heterogeneity. Likewise, a hallmark of many nonlinear systems (including some geomorphologic systems) is sensitivity to initial conditions: a small difference in the initial state of a system can lead to markedly different outcomes (see Gleick 1988). *Stochastic* models are designed to address such uncertainties by including an element of random variability. Such models typically use a random number generator to create a series of alternative inputs (e.g. rainstorms) or to trigger discrete events (e.g. landslides). Discussion and examples of stochastic models are given by Kirkby *et al.* (1992, Chapter 5).

At the heart of most geomorphic process models, both deterministic and stochastic, lies the continuity of mass equation

$$\frac{\partial \eta}{\partial t} = -\nabla q_s$$

where η is the height of the surface, t is time, q_s is the bulk volume rate of mass transport (rock, sediment or solute) per unit width and ∇ denotes a gradient in two dimensions. The continuity equation simply states mathematically that matter can neither be created nor destroyed (short of nuclear reactions). This particular form of the continuity equation is not universally applicable to all geomorphic problems; a slightly different form would be needed to describe horizontal (as opposed to vertical) retreat of a cliff face, the evolution of a fault block undergoing horizontal motion, or surface change due to changes in density rather than in mass, for example. Nonetheless, the continuity law in its various guises is arguably one of a handful of fundamental principles in geomorphological modelling. When combined with a suitable expression for q_s that represents a particular process or processes, the continuity equation can, subject to certain simplifying assumptions, be solved in order to predict landform features such as the shape of hillslope profiles, river profile geometry and soil-depth profiles.

An obvious advantage of mathematical process models over conceptual models (such as the influential concepts of G.K. Gilbert) is that they allow one to make statements not only about 'how' or 'why' but also 'how much' – for example, a mathematical slope model allows one to predict, based on processes, the degree to which the relief in a mountain drainage basin would ultimately change if the rate of uplift were to double (e.g. Snyder *et al.* 2000).

Central to geomorphic process models is the concept of 'geomorphic transport laws' (Dietrich *et al.* 2003). A geomorphic transport law is a mathematical statement about rates of mass transport averaged over a suitably long period of time. The definition of 'suitably long' depends on the process in question, but in general is much longer than the recurrence interval of discrete transport events such as floods, raindrop impacts, landslides, and so on. One of the current research frontiers in geomorphic process research lies in understanding the relationship between short-term transport events and long-term average transport rates.

Solving the continuity equation to predict landform shape requires assuming idealized conditions – for example, in the case of landforms with uniform soil or sediment properties, uniform climate, and height variations in one direction only. Modelling three-dimensional landforms generally requires approximating the solution to an appropriate form of the continuity equation, usually through the use of numerical techniques such as finite differencing, finite volume or finite element methods, cellular automata, or (in some cases) a combination of methods (e.g. Press 2002; Slingerland *et al*. 1994). Typically, these methods produce an approximate solution by dividing up space into discrete elements. Fluxes of mass are then computed within or between these elements. Beginning with a specified initial landform configuration, the evolution of landforms over time is computed by iteratively calculating the transport rates at each point, extrapolating these rates forward in time over a discrete time increment, and then adjusting the topography accordingly. This in turn affects the transport rates at the next time increment, so that the landform emerges as the result of an interaction between its shape and the processes acting upon it.

The development and use of numerical models of landform evolution has grown considerably since the 1980s. Examples include models of rill erosion, river basin evolution, glacial valley formation, and many other coupled tectonic–geomorphic–sedimentary systems. Numerical models of landform and landscape evolution generally operate on what Schumm and Lichty (1965) termed 'cyclic time' – that is, time spans on which landforms can change significantly and which, apart from rapid processes like rill erosion, are generally much longer than a human lifetime. Alongside these 'cyclic time' models are 'event time' numerical models aimed at understanding process dynamics. Here, event time refers to the timescales of individual process events such as floods. This is the timescale on which direct experimentation, observation and application of Newtonian mechanics are most feasible. For example, computational fluid dynamics models have been used to great effect to examine phenomena such as river flooding, coastal sediment transport, soil erosion and debris flows. Often, such models are founded on basic theory in fluid dynamics or material rheology, and combine well-established physical principles (e.g. the Navier–Stokes equations for fluid flow) with empirical laws obtained from laboratory experiments (for background and examples, see Middleton and Wilcock, 1994).

Both event-time and cyclic-time models have had a tremendous impact on geomorphologists' ability to understand the dynamics of processes, and to link these with the landforms that they shape. The applications of models range quite widely, and include both pragmatic forecasting and investigative analysis. Event-time models are often used in an applied context, to make predictions for purposes of planning, land management and insurance assessment. Models of soil erosion, for example, are typically used in this way. In an applied, predictive mode of application, a given model is generally taken 'as read', usually calibrated with existing data, and used to forecast the outcomes of different scenarios.

Numerical models in geomorphology serve other important roles as well. Both event-time and cyclic-time models have been, and continue to be, used in a heuristic mode; that is, they are used as theoretical tools for developing general insight and understanding, rather than for making precise predictions in a particular case study. One of the most valuable roles of mathematical models in geomorphology, in fact, is to make testable predictions about process and form connections. For example, numerous river basin evolution models have been used in 'what if' mode to predict the morphological consequences of statements such as 'the long-term average incision rate of a stream channel is proportional to the rate of energy dissipation per unit bed area' (e.g. Whipple and Tucker 1999). This exploratory process of *forward modelling* makes it possible to reject some models in favour of others, based on their ability to reproduce observed landform characteristics given a plausible set of initial and boundary conditions.

As in other sciences, models in geomorphology both drive and are driven by the results of observational and experimental work. In some cases, a model is developed for the express purpose of explaining a set of data. In others, one or more models are proposed before any relevant data exist, and they stimulate the search for new observations. One example of the latter concerns the relationship between the thickness of soil and the rate of lowering of the soil–bedrock contact. Several models were proposed in the 1960s and 1970s (see Cox 1980). Of these, some predicted an exponential decline in regolith production rate with increasing soil depth, with the maximum production rate occurring at or near the surface. Others predicted a 'humped curve' with a maximum production rate at some optimal soil thickness, due to the added efficiency of water retention. These models remained essentially untested for many years, until cosmogenic nuclide analysis made it possible to infer rates of regolith production. Research beginning in the 1990s has provided evidence for an inverse dependence of regolith production rate on regolith thickness, in some cases with a near-surface maximum (e.g. Heimsath *et al.* 1997), in others with a maximum at depth (e.g. Small *et al.* 1999), depending on process and environment.

The example of regolith production models serves as a caution against the common myth that a model is of no value until and unless it has been validated. In fact, untested mathematical models in geomorphology – like the regolith production models when they were first proposed – have served the field well in two ways: first, by forcing rigour into our hypotheses, and second, by spurring the development of new efforts, ideas and technologies to test the models (for discussion see Bras *et al.* 2003).

A common limitation of models in geomorphology is that different models predict similar outcomes. For example, a range of different river process models predict that graded river profiles should be concave-upward in form – thereby providing multiple, competing explanations for the same observation. This classic problem of EQUIFINALITY, which is common across the Earth sciences, reflects a paucity of data about geomorphic systems. This limitation is part and parcel of the deep-time problem in the Earth sciences (and in other fields such as astronomy and astrophysics). The systems that geomorphologists study are often too big or too slow to allow for direct experiments. Furthermore, most geomorphic systems are dissipative in nature (Huggett 1988). Dissipative systems, by the 2nd law of thermodynamics, lose information as they evolve (consider, for example, trying to reconstruct a snow crystal from a drop of water). Geomorphologists are therefore forced to rely on inference, analogy and indirect evidence. It is no surprise that the problems of equifinality and deep time limit mathematical modelling in the same way that they limit geomorphic knowledge more generally. In principle, the solution to both problems is to obtain as much information as possible about denudation rates, boundary conditions (such as tectonic, climatic or sea-level variations), and the nature of changes in topography over the geologic past. Developing techniques to obtain such data arguably constitutes one of the foremost challenges in geomorphology.

The deep-time problem highlights the fact that, in building models of landform genesis, geomorphologists are forced to 'scale up' contemporary processes over geologic time, and over spatial scales relevant to the landforms in question. This approach is of course a natural outgrowth of Hutton's (1795) ideas. The impossibility of direct experiments makes it especially important to develop accurate constitutive process laws, and to pay careful attention to the role of natural variability in driving forces (such as weather and climate) and in materials (such as soil properties). This represents a considerable scaling challenge, because the formative processes often occur on timescales that are vastly smaller than the timescale required for significant landform change. For example, floods may last for minutes to days while the river basins they sculpt may take shape over hundreds of thousands of years.

Despite their limitations, the future of mathematical models in geomorphology looks bright. Continuing advances in computing power will make solving the scaling problem easier by allowing modellers to link together a wider range of time and space scales. While the deep-time problem will never go away, geomorphologists' ability to explore and test multiple working hypotheses will continue to grow. Likewise, continuing improvements in data describing the Earth's surface topography and in technologies for dating and estimating rates of change will make it possible to test models with increasing degrees of precision.

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SEE ALSO: complexity in geomorphology; computational fluid dynamics; equifinality; laws, geomorphological; mathematics; mechanics of geological materials; non-linear dynamics

GREG TUCKER

MORAINE

A moraine is a glacial landform created by the deposition or deformation of sediment by glacier ice. Many different types of moraine exist, reflecting the many different processes by which glaciers deposit and deform sediment and the many locations and environments within the glacier system where deposition can occur. The material of which moraines are composed, which is generally referred to as till, is also highly variable, as its characteristics depend on the characteristics of the debris supplied by the glacier as well as on the processes and environment of GLACIAL DEPOSITION.

The term moraine has been used in a variety of different ways since it was originally introduced, and its definition remains controversial. Swiss naturalist Horace-Bénédict de Saussure originally introduced the term in 1779, and recognized that ancient moraines represented former extensions of existing glaciers. For the next two centuries the term was widely used to describe landforms created by glacial deposition, the sedimentary material of which the landforms were composed, and the debris in transport within, beneath or on the surface of glaciers. Although modern geomorphological definitions limit the term specifically to landforms, it is still sometimes applied more widely to glacial debris and glacially derived sediment. Many of the compound expressions that feature the term moraine, such as ground moraine and medial moraine, conflate elements of these different definitions, and so moraine continues to be used ambiguously in some geomorphological, glaciological and sedimentological literature. Dreimanis (1989) provides a useful review of the history of the term.

Moraines are classified both genetically according to the process by which they are created and geographically according to their position within the glacier system. There is a fundamental distinction between moraines that occur on the surface of the ice and moraines that occur on the ground surface beneath or at the margin of a glacier. Moraines on the ice surface, known as supraglacial moraines, are ephemeral features that move with the ice and are likely to be destroyed or redeposited on the ground surface when the ice beneath them ablates. They are not true landforms, and inherit the term moraine from now obsolete definitions that included debris in glacial transport.

Supraglacial moraines include lateral moraines (Small 1983), medial moraines (Vere and Benn 1989) and inner moraines (Weertman 1961). Lateral moraines occur at the edge of valley glaciers and comprise debris derived from the valley walls both above and below the ice. Medial moraines occur as longitudinal accumulations of debris downstream from junctions between confluent glaciers, and include debris derived from the lateral moraines of each tributary. Inner moraines are transverse accumulations of debris derived from the meltout of basal debris bands close to the glacier margin. All of these moraine types can develop into large ridges on the glacier surface as the debris cover protects the ice immediately beneath from melting while the surface of the surrounding, debris-free, ice is lowered by ablation. Thick and irregular accumulations of debris released onto the glacier surface by ablation have previously been referred to as ablation or disintegration moraine, forming part of a supraglacial land system, but these terms are increasingly being confined to terrestrial landforms that survive after ablation of the glacier. Supraglacial debris that does not form discrete topographic features on the glacier surface is not referred to as moraine.

Moraines can be formed on the ground surface subglacially or at the edge of the glacier, and by the lowering of supraglacial debris to the ground during deglaciation. They can be formed both by active (moving) ice and by stagnant ice. The principal processes by which moraines are created are the release of debris from ice by meltout and the deformation of proglacial or subglacial sediments by ice motion.

Moraines created by the lowering of supraglacial debris to the ground during deglaciation typically produce a chaotic topography and highly variable sedimentology as the landforms produced are strongly affected by resedimentation, water action and mass movement during their formation.

Subglacial moraines can occur parallel and perpendicular to ice flow or in irregular patterns. There is often a gradual transition between forms with different orientations such as Rogens (described below) and DRUMLINS. The origin of many of these features remains disputed. Areas of subglacial deposition without distinctive relief are sometimes referred to as ground moraine, but this term is falling into disuse and being replaced by non-topographic terms such as subglacial till.

Moraines parallel to ice flow include streamlined features within subglacial till, such as flutes and certain types of drumlins. The genesis of some of these features remains controversial. Whereas traditional analyses attribute them to subglacial deposition and the sculpting of subglacial deposits by moving ice, other interpretations based on subglacial meltwater processes (e.g. Shaw *et al.* 1989) imply that these features are not true moraines at all.

Transverse subglacial moraines include similar features in different locations that have been given various names and interpretations. The labels Rogen, De Geer, ribbed, washboard, corrugated, cyclic and cross-valley moraines have been applied to transverse features associated with subglacial processes. Rogen moraines are large ridges several tens of metres high, over 1km long and with crests several hundred metres apart, giving an irregular ribbed appearance to large areas of the landscape. They are often associated with flutes and drumlins, and are most commonly attributed either to thrusting of debris-rich basal ice into localized stacks beneath ice in compressive flow, or to deformation of subglacial sediment. The deformation hypothesis places Rogens at one end of a continuum of deformational forms that grades at the other end into longitudinal ridges such as flutes and drumlins. De Geer moraines are generally smaller in scale, and characterized by water-lain deposits within the moraine suggesting an origin beneath ice grounded in water.

Subglacial moraines lacking consistent orientation have been referred to as hummocky ground moraines. These are attributed either to the lowering of supraglacially released ablation moraines or to the release of debris beneath stagnant ice. The subglacial hypothesis places hummocky moraine at one end of a spectrum of forms that incorporates washboard moraines (weakly oriented hummocks) and drumlins (streamlined hummocks reflecting deposition beneath moving ice)

(Eyles *et al*. 1999). Some areas of hummocky moraine have recently been reinterpreted as complex assemblages of cross-cutting and discontinuous subglacial, supraglacial and ice-marginal moraine ridges.

Ice marginal moraines occur around the edges of glaciers and are defined by their position as either lateral or frontal moraines. Moraines marking the maximum extent of a glacial advance are referred to as terminal moraines. A terminal frontal moraine is called an end moraine. Moraines deposited at successive positions of the margin during a period of progressive retreat are referred to as recessional moraines. Moraines deposited at successive positions of the margin during periods of advance are usually destroyed by the advancing ice and are not preserved in the landscape, except for the terminal moraine.

Marginal moraines at existing glaciers are typically ridges of sediment resting partially on the edge of the glacier and partially on ice-free ground beyond the margin. Upon deglaciation, ice-cored moraines lose their ice support and therefore tend to shrink in size and may become structurally unstable (Bennett *et al*. 2000). Marginal moraines may be several tens of metres in height, tens or hundreds of metres across, and may stretch for hundreds of kilometres around the margins of large ice sheets. The main processes for the formation of marginal moraines are dumping of supraglacial, englacial or basal debris transported through the glacier, and pushing of sediments previously deposited in front of the glacier.

Small push moraines can be formed by seasonal bulldozing of proglacial sediment where an ice margin oscillates with seasonally varying ablation. Larger push moraines can form by the superposition of several seasonal moraines or by a substantial advance of the margin into deformable materials. Other glacitectonic features include moraines formed by the squeezing out of deformable sediment such as saturated till from beneath the ice margin.

Meltout or dump moraines occur where englacial or supraglacial sediment is transported to the margin and dumped where the glacier ends. Dump moraines grow in size for as long as an ice margin remains *in situ* to supply sediment, their rate of growth depending on the rates of sediment supply and ablation.

The morphology and sedimentology of moraines can be used to reconstruct the characteristics of former glaciers. The distribution of moraines reflects the geography of former glaciers and glacial process environments. Dated terminal and recessional moraines reveal the history of decay of a glacier, and process-controlled moraines reveal the locations of specific process. For example subglacial crevasse-fill ridges, which are formed by the squeezing of subglacial sediment into crevasses in the base of a glacier have been cited as indicators of glacier surging (Sharp 1985). Sediment characteristics reflect the source location of the debris: supraglacial debris is characteristically angular, while basally derived debris is typically basal faceted, subrounded and striated. Knight *et al*. (2000) showed how the distribution of clay-sized particles in a moraine reflected the distribution of a particular type of debris within the glacier that only occurred in certain process environments. Complex structures within moraines can reveal seasonal and long-term variations in processes of sedimentation. Small *et al*. (1984) showed how lateral moraine ridges derived aspects of their internal structure from seasonal variations in debris supply.

Moraines are one stage in the glacier sediment transfer system, providing long-term storage and a supply of debris to the proglacial zone. Sediment flux within glaciated basins is very sensitive to the position of glaciers relative to their moraines. When glaciers lie behind marginal moraines the bulk of sediment produced at the margin can go into storage in the moraine belt and not reach the proglacial region. When glaciers have no marginal moraines, sediment passes directly into the proglacial system. When glaciers re-advance over ancient moraines, large amounts of sediment from the moraine can be released from storage and transported into the proglacial landscape. Moraines can also focus meltwater discharge, localizing fluvial processes and causing meltwater from the glacier to be ponded up to form moraine-dammed lakes. These lakes are potentially unstable and pose a serious threat of catastrophic flooding.

Moraines are significant features within glaciated landscapes, useful indicators of past glacial activity and important components of the glacial sediment transfer system.

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SEE ALSO: glacial deposition

PETER G. KNIGHT

MORPHOGENETIC REGION

A morphogenetic region is an area where landforms are, or have been shaped, by the same or similar processes, mainly those controlled by climate. In climatic geomorphology there are two spatial categories: in morphoclimatic zones typical processes are considered, whereas in climato-morphogenetic regions the distinctive morphogenesis of an area is investigated. These definitions are more or less followed in continental Europe. In Anglo-American geomorphology, on the other hand, the term morphogenetic is used differently as 'the extent to which different climatic regimes are potentially capable of exerting direct and indirect influences on geomorphic processes, and thereby of generating different "morphogenetic" landform assemblages' (Chorley *et al*. 1984: 466). This nearly corresponds in German terminology to 'klimamorphologische Zonen' (morphoclimatic zones), and in French to 'les zones morphoclimatiques'. In the English literature these German and French terms are sometimes wrongly translated as 'climato-morphogenetic regions'.

The terms 'arid, humid, and nival' were introduced in 1909 by A. Penck as names for zones with distinct climate, hydrology and geomorphology. He had already recognized that these zones had shifted during the warm and cold periods of the Pleistocene and in 1913 introduced the term 'pluvial'. In 1926, in a symposion at Düsseldorf on the 'Morphologie der Klimazonen' (morphology of climatic zones), nine geomorphologists gave an overview of their research in certain areas ranging from the arctic to the humid tropics. Each one compared his findings with central Europe to stress the peculiarities. In 1948 Büdel introduced 'Das System der klimatischen Geomorphologie' (the system of climatic geomorphology). He gave a description of the typical processes in each morphoclimatic zone. The most important aspect was the interrelation of the processes in one zone, e.g. the work of a river is dependent on the relief of the area, which next to precipitation controls the amount and time of discharge. The load which has to be transported is generated from slopes and small creeks. By their interrelationships the relative strength or influence of the processes shaping the landforms should become clear. The relation between fluvial erosion and denudation was especially weighted. Thus there was some estimation of erosion rates, too. Not only were the most spectacular landforms looked for, but also the most widely distributed ones. Not only the catastrophic events but also the slowly working processes were investigated. For each morphoclimatic zone the processes were recorded as they were observed from recent occurances, from the observation of the REGOLITH, and from a check with the landforms, whose shape was interpolated and extrapolated with the processes, a feedback. The concept of morphoclimatic zones is a very open one and may be varied, e.g. according to rock resistance (petrovariance) or tectonics (tectovariance). This is a rather broad approach and of course uncertainty or even mistakes are possible. This does not spoil the concept though. Comparison of similar regions and results from different research has increased our knowledge of the different morphoclimatic zones, though no new complete version has been made since the handbook of Büdel (1977, 1982). However, many detailed studies are founded on this concept.

For the interrelation of forming processes the terms 'Prozessgefüge' (process fabric) or 'Formungsmechanismus' (relief forming mechanism) came into use. For one of the morphoclimatic zones, the humid mid-latitudes 'zone of Holocene retarded valley building', Büdel (1982: 14) named the following components, which make up or control the 'forming mechanisms for the highly complex phenomena and processes: solution, mechanical weathering, chemical weathering, plant cover, soil development, surface denudation, linear erosion, transport, and deposition'. They are connected on 'highly complex integration levels'. In quoting 'highly complex' twice and adding 'occuring only in nature, not reproducable', he wanted to stress that on this level field measurements and laboratory experiments should be combined with the 'predominant qualitative relief analysis'. The main methods are field observations in 'natural test sites', where the phenomena are typical and which have to be searched for. Then follows the comparison with similar areas, where e.g. the influence of different rocks can be observed. Thus by comparison the petrovariance and the tectovariance can be abstracted and the processes controlled by climate become clear.

It is easier to link relief forming mechanisms to ecological factors for which climate is an abbreviation, as these comply to a zonal order, than to build a system on lithology. Of course there are distinct landforms in limestones, sandstones and granites and excellent relevant handbooks, but there is no systematic arrangement of forms due to differences in rock hardness or structure. Thus a morphogenetic region according to one of these rock groups would more or less coincide with a geological map. That would not be a new insight. It is possible to outline morphotectonic domains, but the connection to geomorphological processes is only very slowly developing, as detailed knowledge of the influence of tectonic movements on processes, except landsliding, is very small so far, and for cratons almost unknown. In both cases palaeoforms are hard to incorporate systematically, but this is easy in climatic geomorphology.

A morphoclimatic zone defined by relief forming mechanisms is a framework for detailed studies. These may be of megaforms, mesoforms or microforms and it is possible to apply many different methods. For instance, if landform facets are linked to the thickness and texture of the regolith and/or sediments, their relative age and evolution is investigated. This can be verified by laboratory research of the material, and by absolute datings. If the extent of the landforms is mapped or their changes are derived from sequences or monitoring, there is an estimation of the volume of transport possible. As this holds mainly for several hundreds or thousands of years, this provides a long-term check for short-term measurements of material transport. Thus it is possible to discriminate between natural and human-induced erosion rates. The concept of morphoclimatic zones is helpful to provide working hypotheses with regard to the full breadth of processes possible, their interdependence and relative strength. Especially in extrapolation of measured properties, an appraisal of relief forming mechanisms should be incorporated.

The essence of the concept of morphoclimatic zones is the interrelation of processes and there are almost no attempts in climatic geomorphology, as understood in Europe, to link landforms to climatic data. As in a morphoclimatic zone the interrelation of weathering, denudation and fluvial erosion is described, and it is obvious that only a general combination with climate is possible. Büdel (1977, 1982) himself delineated ten morphoclimatic zones. Originally (1948) there were twelve, and in 1963 they were reduced to five with an emphasis on the tropical semi-humid zone of excessive planation and the subpolar zone of excessive valley cutting. The names were changed too, though only slightly. This may show that zonation was not Büdel's foremost interest. He never tried to link the boundaries of the zones to climatic data. He rather insisted on the complexity of relief analysis, covering as many ecological factors as possible. There is one inconsistency, too. The zone of excessive planation should be shifted to the perhumid tropics, as only there is weathering intense enough for the concept of double planation, which is still very valid. The term climatic geomorphology is a misnomer, but the attempt to change to dynamic geomorphology was not successful as the term was introduced for a long time in contrast to tectonic geomorphology.

The difference of the Anglo-American approach to morphogenetic regions is twofold: the relevance of climatic data at the start and the broadness of the approach. The first attempt at delineating morphogenetic regions in the USA was the diagram by Peltier (1950). It was much cited but had little influence on detailed studies. Even the more sophisticated diagram of Chorley *et al*. (1984) has not been filled by regional or areal studies. Thus climatic regions as a starting point and the deduction of possible processes does not seem to be very fruitful. Instead there are single features like drainage densities connected with climatic data, or gradients of rivers or slopes are linked to sediment transport and rainfall variables. Polygenetic landforms are quite often approached from the knowledge of palaeoclimates. On the other hand there are excellent books on tropical, desert, periglacial, glacial geomorphology and karst, which describe and explain landforms and processes. But there are few interrelations and almost no connection to climatic data, though these handbooks often contain a chapter on the climate of the zone.

An extension of the morphogenetic regions was done by Brunsden in creating tectono-climatic regions. He proposed linking geotectonic domaines with morphoclimatic zones. For example, he entered into a map of the present conditions of the Indo-Australian plate the recent morphoclimatic zones and second the environmental conditions of 18,000 BP. Comparison of these two pictures gives areas of tectono-climatic stability. These are interesting hypotheses, but here, too, the starting point is the concept from facts outside geomorphology. Only later shall it be filled with field observations. It is a way that proceeds from the top downward, not from the base upwards.

It is always possible to concentrate on a special process but this should not be done in an isolated way but in the realm of the relief forming mechanism. Thus it is tied up in an analysis of interrelations of larger to smaller forms, of single processes to the process fabric. Thus the extrapolation of single processes and the interpretation of landforms becomes more secure. An example might be river terraces in mid-latitudes. Are they of climatic or tectonic origin? Not only the material of the terraces and their gradient is indicative but the origin of the pebbles and the mode of transport from the source area on a slope (e.g. by solifluction into the rivers). Such features as periglacial ice wedges casts and covers like loess are studied in relation to former climatic conditions and age. Are similar terraces developed in neighbouring areas? Which forms are incised in the older terraces? This for instance led to a detailed history of incision for the middle Rhine valley. This part of the valley is antecedent and developed during slow uplift, but the forms in detail are climate controlled. This is an example for a morphogenetic region in German understanding. There are similar regional studies in the English literature. The methods are more detailed in CLIMATO-GENETIC GEOMORPHOLOGY.

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MORPHOMETRIC PROPERTIES

Morphometric properties of a DRAINAGE BASIN are quantitative attributes of the landscape that are derived from the terrain or elevation surface and drainage network within a drainage basin. GEOMORPHOMETRY is the measurement and analysis of morphometric properties. Traditionally morphometric properties were determined from topographic maps using manual methods, but with the advent of geographic information system (GIS) technology, many morphometric properties can be automatically computed.

Size properties

Size variables provide measures of scale that can be used to compare the magnitudes of two or more drainage basins. Size variables are derived from measurements of the basin outline as defined by the drainage divide or are obtained from the drainage network. Many size variables are strongly correlated with one another so can be used interchangeably.

Drainage area, the two-dimensional projection of area measured in the map plane, is the most important size measure and is specified as the area contained within the drainage divide. RUNOFF GENERATION and the frequency of FLOODS is directly correlated with drainage area in many environments.

Basin length indicates the distance from the basin outlet to a point on the drainage divide, but many different methods for measuring basin length have been devised. For example, the end-point of the length measure can be the highest point on the divide or the point on the divide that is equidistant from the outlet along the divide. Perimeter is a measure of distance around the drainage basin measured along the drainage divide.

Main channel length is the length from outlet to channel head along a subjectively defined main channel, or, more objectively, the length of the longest flow path to the drainage divide. Total channel length is the sum of lengths of all channels in a basin.

Stream order can also be used to indicate basin size (see STREAM ORDERING). The order of a basin is the order of its outlet stream. Stream magnitude is the number of FIRST-ORDER STREAMS in a basin. Magnitude is a more discerning measure of size than is order.

Surface properties

Surface properties are quantities depicted by fields comprising a value at each point within a domain (drainage basin). GIS technology provides the capability to derive surface properties from a DIGITAL ELEVATION MODEL (DEM) which is the numerical representation of an elevation surface. The elevation surface is the most fundamental surface property field, and quantifies the ground surface elevation at each point (neglecting cave and overhang special cases). DEM types include square or rectangular digital elevation grids, triangular irregular networks, sets of digital line graph contours or random points (Wilson and Gallant 2000).

The flow direction field is the direction that water flows over a surface under the action of gravity. This may be defined by the horizontal component of the surface normal. The flow direction field is represented numerically by a flow direction grid. The simplest flow direction grid is the D8 flow direction grid in which flow direction is represented by one of eight values. The value depends on which of the eight neighbouring cells (four on the main axes, four on the diagonals) is in the direction of steepest descent and thus receives its drainage. Other numerical flow direction fields can be derived using finite difference or local polynomial or surface fits to elevations of grid cells in the neighbourhood of each point (Tarboton 1997).

Terrain slope is a field giving the slope of the terrain in the direction of the flow direction field at each point. This is evaluated numerically by taking elevation differences from the elevation field over a short distance centred on each point.

Contributing area is a field representing drainage area upslope of each point. It is defined by tracing flow paths up slope from each point along the flow direction field to the drainage divide and measuring the area enclosed. Within a grid-based GIS, contributing area is evaluated by counting the number of grid cells draining to each grid cell. Contributing area is also referred to as catchment area or flow accumulation area.

Specific catchment area is a field representing contributing area per unit contour length. On a smooth surface, the contributing area to a point may be a line that has zero area. Specific catchment area is quantified using the measurable area contributing to a small length of contour (Moore *et al*. 1991: 12). Specific catchment area has units of length. On a planar surface with parallel flow, specific catchment area is equal to the upslope distance to the drainage divide.

Shape properties

Drainage basin shape is a difficult morphometric property to characterize simply, and there have been numerous attempts at defining shape variables. The simplest shape measures employ area, length, width or perimeter of the drainage basin or of a shape with area equivalent to that of the basin. More complex functions of drainage basin or drainage network shape are best portrayed using two-dimensional graphical plots.

The cumulative area distribution function is defined as the proportion of a drainage basin that has a drainage area greater than or equal to a specified area. It is typically represented by plotting cumulative area versus area on a log-log line chart.

The distance area diagram depicts the area of the basin as a function of distance along flow paths to the outlet. The channel network width function is the number of channels at a given distance from the drainage basin outlet, as measured along the drainage network, and is typically plotted as a line or bar chart. The distance area diagram and channel network width function both give an indication of basin hydrological response and are related to the instantaneous unit hydrograph.

Relief properties

RELIEF properties bring the dimension of height into morphometric analysis. Because many landscape processes are driven by gravity, relief properties are frequently used as indicators of EROSION potential and DENUDATION rates.

Total basin relief is the difference in height between the outlet and the highest point on the drainage divide. Relief ratio removes the size effect by dividing total relief by basin length. Sediment yield (see SEDIMENT LOAD AND YIELD) in small drainage basins has been shown to be exponentially related to relief ratio (Hadley and Schumm 1961: 172).

A more complex representation of basin relief is the area-elevation relationship or hypsometric curve. The hypsometric curve is a plot of the area of a basin (on the x-axis) above each elevation value (on the y-axis). The axes are commonly normalized to range between zero and one. The hypsometric curve is equivalent to one minus the cumulative distribution of elevation within a drainage basin. Davisian model evolutionary stage can be inferred from the shape of a basin's hypsometric curve.

Texture properties

Texture indicates the amount of landscape dissection by a channel network. The contours on a map of a highly textured landscape will have many small crenulations (wiggles) indicating the presence of numerous channels.

DRAINAGE DENSITY (Horton 1945: 283), the best-known texture indicator, is defined as the lengths of all stream channels in a drainage basin divided by drainage area and has units of 1/length. Drainage density ranges from less than $1\,\mathrm{km}^{-1}$ to over $800\,\mathrm{km}^{-1}$, attaining maximum values in semi-arid areas (Gregory 1976: 291). High drainage densities indicate highly textured landscapes, short hillslopes and domination by OVERLAND FLOW runoff typical of BADLANDS.

The area–slope relationship quantifies the area draining through a point versus the terrain slope at that point, typically plotted on a log-log scale graph. The scatter when all points or grid cells are used is removed by binning (e.g. using a moving average) to reveal a characteristic area–slope relationship with two distinct regions. For small areas, slope increases with drainage area and for large areas, slope decreases with area. The turnover point in the relationship has been interpreted as the drainage area at which diffusive hillslope processes (see HILLSLOPE, PROCESS) are overtaken by fluvial processes and channels are initiated (Tarboton *et al*. 1992: 73).

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SEE ALSO: Horton's Laws

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MORPHOTECTONICS

Morphotectonics is the term pertinent to links between geomorphology and tectonics, although individual authors apparently understand the exact nature of these links in slightly different ways. Most often, morphotectonics is considered synonymous with TECTONIC GEOMORPHOLOGY and defined simply as the study of the interaction of tectonics and geomorphology. Embleton (1987) lists four main lines of interest in morphotectonic research: (1) study of landforms indicative of contemporary or recent tectonic movement, (2) study of deformation of PLANATION SURFACES, (3) study of geomorphological effects of earthquakes (see SEISMOTECTONIC GEOMORPHOLOGY), (4) use of geomorphological evidence to predict earthquakes. It needs to be emphasized that in some countries morphotectonics is a term of very limited usage. For example, two recent American textbooks about tectonic geomorphology (Burbank and Anderson 2001; Keller and Pinter 2002) do not mention morphotectonics, although they evidently deal with this kind of phenomenon.

Fairbridge (1968) offers a different explanation and understands morphotectonics as a means to classify major landforms of the globe rather than any landforms related to tectonic processes. Accordingly, he distinguishes morphotectonic units of first and second order. In the first order these are continents and oceanic basins, in the second one there are shields, younger mountain belts, older mountain massifs, basin-and-range areas, rift zones and basins. This global context of morphotectonics is also evident in the study of great ESCARPMENTS (Ollier 1985).

In practice, the morphotectonic approach frequently means using landforms or any other surface features (e.g. drainage patterns) as a key to infer the existence of tectonic features, especially in relatively stable areas where seismicity and present-day rates of uplift and subsidence are negligible. They acquire the status of geomorphic markers of tectonics. Geomorphological maps, drainage pattern maps, digital elevation models and their various derivatives are analysed with the aim of locating anomalies in landform distribution, river courses, channel form, terrace profiles, local relief or specific landforms such as slope breaks. These anomalies in turn, if no other explanation for their occurrence is available, are considered to reflect the presence of tectonically active zones or areas. Detailed analysis of river patterns can be a particularly valuable tool in morphotectonic research in lowland areas, where hardly any other evidence is at hand.

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SEE ALSO: active and capable fault; active margin; fault and fault scarp; global geomorphology; neotectonics; passive margin

PIOTR MIGOŃ

MOULIN

Moulins, or glacier mills, are sink holes that owe their name to the roaring noise of water that engulfs itself in them. They form in the ablation zone of GLACIERS (Paterson 1994), where meltwater (see MELTWATER AND MELTWATER CHANNEL) manages to cut stream channels into the ice, generally parallel to glacier slope. These channels are eventually intercepted by crevasses, which form perpendicular or oblique to glacier slope in response to ICE flow related to bedrock surface irregularities. Moulins are the result of meltwater flowing into crevasses (Rothlisberger and Lang 1987).

Moulins are characterized by a vertical shaft up to 100 m tall, developing along the planes of single or cross-cutting crevasses and prolonging into a downflow dipping gallery that follows structures related to glacial flow. The gallery dips approximately 45° and forms a succession of pools on an irregular floor, but the gallery sometimes dips approximately parallel to glacier slope when the shaft is less than 50 m tall. Shafts are circular or elliptical in horizontal cross section and range between less than 1 m to over 20 m in their long axis, but detail of their morphology is controlled by the dynamics of the water that flows into it (Holmlund 1988).

The bottom of moulins is often submerged. Water level can vary within a few hours and from one season to the next in relation to meteorological conditions, glacial flow, ice plasticity and according to the facility with which water can flow along the base of the glacier.

During the summer, moulins provide the main inputs of glacial aquifers. During the winter, they are separated from the surface by a snow bridge. However, water level in the moulins usually increases during the winter and then decreases in a jerky fashion, the moulins functioning like surge tanks (Schroeder 1998). This implies that drainage in the glacier tends to clog in the front first during the beginning of the cold season, while the water column that then remains stocked within the moulins prevents glacial flow from closing it. With the onset of the warm season, this water that was stocked within the moulins helps in reinitiating subglacial drainage.

The life expectancy of moulins can reach several dozens of years. Moving along with the glacier, they eventually lose their connection to glacier surface drainage at the profit of new moulins forming upflow from them. In dead ice, moulins often reach down through the entire glacier. Megapotholes (diameter > 50 m) developed in rock bars of regions that were glaciated during the Quaternary are thought to be the result of extended water circulation at the base of moulins in stagnant ice.

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JACQUES SCHROEDER

MOUND SPRING

Small mounds formed preferentially along fault lines by artesian springs. Solutes and colloids are precipitated to form travertines or tufas (see TUFA AND TRAVERTINE) of calcium carbonate, together with various siliceous and ferruginous deposits. Wind-blown sand and accumulated plant debris, together with mud and sand carried up with the spring water, assist in their formation. Where springs display high rates of water flow there tends to be little or no mound formation – they are too erosive. However, springs with low discharge rates and laminar flow experience high rates of evaporation (especially in arid environments) and have a greater possibility of accumulating chemical precipitates.

Major examples of such features are known from the Great Artesian Basin of Central Australia (Ponder 1986) and from the depressions of the Western Desert in Egypt, where much accretion has occurred when vegetated fields, irrigated by the springs, have trapped aeolian sediment (Brookes 1989).

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A.S. GOUDIE

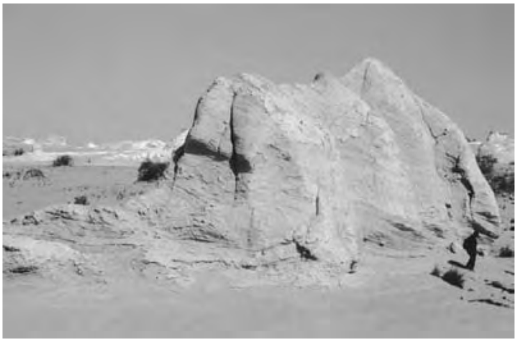

Plate 82 A silty mound deposit associated with spring activity in the Farafra Oasis of the Western Desert of Egypt

MOUNTAIN GEOMORPHOLOGY

Mountain geomorphology is a 'regional component within geomorphology' (Barsch and Caine 1984). The region in this case is the world's mountains defined by absolute elevation (>600 m above sea level), available relief ($>200\,\mathrm{m\,km^{-2}}$) and topographic slopes ($>10°$). There is no international standard definition, but other elements which are frequently incorporated are high spatial variability, presence of ice and snow and evidence of late Pleistocene glaciation. Carl Troll (1973) who was the modern creator of mountain geomorphology, defined mountain systems as those which encompass more than one vegetation belt, but do not reach alpine elevation by contrast with high mountain systems (hochgebirge) which extend above the timberline.

Fairbridge (1968) rehearses the classification of mountains by scale and continuity: (a) mountain is a singular, isolated feature or a feature outstanding within a mountain mass; (b) a mountain range is a linear topographic feature of high relief, usually in the form of a single ridge; (c) a mountain chain is a term applied to linear topographic features of high relief, but usually given to major features that persist for thousands of kilometres; (d) a mountain mass, massif, block or group is a term applied to irregular regions of mountain terrain, not characterized by simple linear trends; and (e) a mountain system is reserved for the greatest continent-spanning features.

A simple genetic system of mountain types, which was developed before global plate tectonics was understood, is nevertheless useful in local-scale understanding. There are two broad categories: (1) structural, tectonic or constructional forms and (2) denudational or destructional forms. Under the first category can be identified: (a) volcanic; (b) fold and nappe; (c) block; (d) dome; (e) erosional uplift or outlier; (f) structural outlier or klippe; (g) polycyclic tectonic; and (h) epigene mountains. Under the second category are defined: (a) differential erosion; (b) exhumed; (c) plutonic and metamorphic complex; and (d) polycyclic denudational mountains (Fairbridge 1968). In the structural mountain categories it is the tectonic process that has played a primary role; in the denudational categories it is the denudational processes that are primary. The lithology and the climatic history are both extremely important with respect to the detailed modification of these mountain types. Indeed, much of the science of geomorphology is centred on the discrimination of these second-order effects.

The simplest typology of mountain geomorphology makes use of the tripartite division into historical, functional and applied mountain geomorphology. Historical mountain geomorphology focuses on the evolution of mountains and mountain systems over both long and medium timescales. It is common, at the largest scale, to distinguish between young active mountain belts which have evolved throughout the Cenozoic and are still associated with active plate margins and mountains on passive continental margins. Nearly all the literature on mountain building in the past forty years has concentrated on active margins where collision and subduction may explain both mountains and the structures within them. Most exciting in recent years is the trend towards quantifying rates of uplift and denudation with the development of new geochemical, geochronological and geodetic methods. But, in reality, there are mountains on PASSIVE MARGINS too (Ollier and Pain 2000). The evolution of these older mountain belts is intrinsically more complex as they do not easily fit into the simple plate tectonic story of mountain building at collisional sites and include the history of the Earth since the breakup of Gondwanaland during the Mesozoic. A major difference of opinion has emerged between those who place greatest emphasis on the data from FISSION TRACK ANALYSIS and those who use whatever landform, stratigraphic and geological data that can be found to constrain the interpretation. Whereas geomorphic models of denudation history are difficult to validate, interpretation of fission track data in terms of denudation history is complex.

Functional geomorphology of mountains includes the assessment of processes, rates and spatial and temporal patterns of mountain belt erosion. The process framework should ideally involve a consideration of the coupling of uplift and erosion; many geomorphic models have failed to include realistic models of this coupling. In mountain belts, such a consideration is obligatory as both uplift rates and erosion rates achieve maximum values and the coupling of the processes is even more critical than in lowland regions. Improvements in understanding of fluvial bedrock incision processes, hillslope mass wasting, glacial valley lowering and SEDIMENT ROUTING are leading to the development of improved mountain landscape evolution models.

Feedbacks between tectonic, climatic and geomorphic processes have been explored in geodynamic models and solid, solute and organic fluxes from mountain belts have been constrained and considered within a global geochemical context. The topographic evolution of mountain belts can be modelled with increasing realism, but the issue of equilibrium conditions versus transience is still far from resolved.

APPLIED GEOMORPHOLOGY of mountains: mountain habitats create or magnify natural hazards in the form of dangerous geomorphic processes. The interaction of geomorphic processes with mountain societies, their land uses and their response capabilities determines risk. Recent social and environmental changes in the mountains has led to the modernization of the natural hazard problematique. As a result, planned responses, including mitigation strategies for specific hazards and mountain disasters, must be developed to reduce the vulnerability of mountain peoples. The applied mountain geomorphologist has a distinctive role to play.

There are three formal or semi-formal attempts to define the field in the literature: Hewitt (1972) Barsch and Caine (1984) and Slaymaker (1991).

Hewitt addressed two issues: the idea of a high energy condition and the relation of distinctive morphological features to clima-geomorphic conditions and denudation history. These he said express the distinctiveness of mountain geomorphology. Under the high energy condition, he dealt with regional rates of net erosion, magnitude and frequency of erosional events and energy in the mountain geomorphic system. Under distinctive morphological features, he picked out accordant erosion surfaces, valley asymmetry and threshold slopes for detailed treatment. The beauty of Hewitt's vigorous statement is that he foreshadows the increasingly heavy emphasis on the operation of geomorphic processes in mountain regions, but also warns of the danger of not relating these observations to the larger questions of mountain landscapes and neglecting to either solve them or to restate them in better terms.

Barsch and Caine (1984) divide mountain geomorphology into studies of mountain form and morphodynamics in mountains. These two categories they further subdivide into (a) morphometry and structure, (b) relief generation and history, (c) morphoclimatic models and (d) process dynamics and activity. Morphometry and structure depend heavily on plate tectonic setting of the mountains. There are four convergent plate settings in which some of the most rapidly evolving mountain systems of the world are located. These are: oceanic to oceanic plate convergence (e.g. Japanese Alps and the Aleutian Arc, Alaska); oceanic to continental plate convergence (e.g. South Island, New Zealand and Cascade Ranges, Pacific North-West); continental to continental plate convergence (e.g. Himalayas); and displaced terranes along accreted margins (e.g. British Columbia). Divergent plate settings include sites of oceanic spreading, such as Iceland and the Galapagos Islands, and intra-continental rifts, such as the Gulf of Aqaba and the Scottish Highlands. Transform plate settings are threefold: ridge past ridge (e.g. Coast Ranges of California); trench past trench (e.g. Anatolia, Turkey) and ridge past trench (e.g. Pakistan–Afghanistan). It is not difficult to understand why mountains are preferentially located in all of the above plate marginal locations. But mountains are also found in plate interior settings, such as the following: hot spots (e.g. Hawaii and Yellowstone National Park); continental flood basalts (e.g. Deccan, India and Columbia Plateau, Pacific North-West); shields (e.g. Ahaggar Mountains, Sahara); intracratonic uplift sites (e.g. the San Rafael Swell, Utah); post-tectonic magmatic intrusion sites (Air Mountains, Nigeria); and evaporite diapers (e.g. Zagros Mountains, Iran). They note that in most mountain regions, the balance between denudation and tectonic uplift is resolved in favour of the latter. They fail to differentiate between high mountains and mountain systems on the basis of morphometry alone, but they make the case that there are four distinctive 'relief types' within high mountain systems, namely the Alp type, the Rocky Mountain type, polar mountains and desert mountains. The Alp type is associated with an overriding impact of glacial ice and glacial erosion; the Rocky Mountain type has a less pervasive impact of glacial erosion and includes areas of low relief on flat summits and rounded interfluves; polar mountains give evidence of intense glaciation, but are frequently with a local relief of less than 1,000 m; and desert mountains are high mountains in the 'true' sense even though they do not reach timberline and were only lightly glaciated during the Pleistocene.

Relief development in high mountains revolves around questions of accordant surfaces and valley benches as indicators of mode of valley dissection. Attention is directed to (a) the alpine summit accordance or 'gipfelflur', which has been explained as a remnant of an old erosion surface; (b) the alpine crest and summit accordance, explained as the product of regular patterns of dissection which constrain summits to approximately the same elevation; (c) the timberline and alp slope accordance, explained as an inter-glacial alp slope associated with a higher timberline than the present one; and (d) benches along the sides of major valleys, variously explained as Tertiary, Pleistocene glacial and inter-glacial timberline effects. Ford *et al*. (1981) have suggested that the age of the present relief of the southern Canadian Rockies is Pliocene, considerably older than had previously been thought.

Building on Caine (1974), Barsch and Caine (1984) distinguish four mountain geomorphic process systems: (1) the glacial system; (2) the coarse debris system; (3) the fine clastic sediment system; and (4) the geochemical system. Of the four, the glacial and the coarse debris systems are most characteristic of high mountain terrain. The final section of their paper summarizes contemporary geomorphic activity in high mountain areas using calculations of sediment flux (in $\mathrm{J\,km^{-2}\,yr^{-1}}$) from Sweden, Switzerland and the United Sates (Rapp 1960; Jackli 1957; Caine 1976). Most interesting was the observation that talus shift, solifluction, soil creep and other processes of slow mass wasting accounted for no more than 15 per cent of the geomorphic work done in all three areas and their relative importance decreased with increasing size of basin. The authors suggest that there are two urgent needs for mountain geomorphology: (1) linking process and form in a meaningful way and (2) identifying anthropogenic influences and ways in which they may be propagated through the mountain system.

Slaymaker (1991) suggests that the meso and macro-scales are the only spatial scales at which a distinctive mountain geomorphology signal is likely to be apparent. He then adopts a slightly modified version of the Chorley and Kennedy (1971) open systems framework to identify five mountain systems: (1) morphological; (2) morphologic evolutionary; (3) cascading; (4) process-response; and (5) control systems. Each of these mountain systems is examined at meso- and macro-scales in search of characteristic mountain geomorphology forms and processes. He claims that this typology is useful in that different measurement programmes are appropriate within each of the ten mountain systems identified.

In fact, these ten mountain geomorphic systems serve to underline the huge variety of forms and processes that characterize mountain geomorphology and support the contention that mountain geomorphology is characterized by its extreme gradients, not only of topography, but also of energy and mass balances and ecological responses. High vertical and horizontal rates of change over space of landforms and processes and rapid rates of change over time distinguish mountain geomorphic systems from other regions. Hence the validation of mountain geomorphology as a regional component within geomorphology.

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SEE ALSO: plate tectonics

OLAV SLAYMAKER

MUD FLAT AND MUDDY COAST

The term mud is used to refer to sediments comprised chiefly of silts (size range 4 to 63 μm) and clays (finer than 4 μm). Such fine material is readily maintained in suspension and can be transported over long distances by coastal currents. Unlike coarser sands and gravels, muddy sediments tend to be cohesive. The electrochemical properties of clay mineral particles mean that these can bind together to form larger composite particles in a process known as flocculation. Flocculation is influenced by a variety of factors, notably salinity, fluid shear and suspended sediment concentration (Lick and Huang 1993). The effect of these processes may vary over quite short spatial and temporal scales, especially in estuaries, where mixing of freshwater and saltwater occurs and where marked variation in flow intensity occurs at tidal timescales. The cohesive nature of muddy sediments makes their behaviour far more complex than that of non-cohesive sands. Flocculated sediments typically settle from suspension far more rapidly than their constituent mineral particles, and the stability of natural muddy deposits is governed not only by physical processes but also by the activity of a rich and diverse biota including macroscopic and microscopic algae, invertebrates and bacteria (Paterson 1997).

Muddy coasts typically occur along low energy shorelines that are well supplied with silt and clay-sized sediments. They include many estuarine margins, delta shorelines, and areas of open coast subject to low wave energy. Such settings are usually dominated by tidal processes, and the characteristic landforms of muddy coasts – SALTMARSHes, MANGROVE SWAMPs and tidal flats – are often very well developed under macro-tidal conditions (Hayes 1975). Enormous quantities of muddy sediment are supplied by some of the world's major rivers, and their estuaries and deltas often feature extensive shore-attached mud banks. Open coast mud banks occur downdrift of major fluvial sediment sources, notably in the Gulf of Mexico (associated with the Mississippi River); more than 850 km of the Jiangsu coastline of China, supplied by the Huanghe and Changjiang Rivers (Ren 1987); and along the south-west coast of India. Both estuarine and open coast mud banks are highly dynamic landforms, which exhibit both seasonal and decadal style variability in response to variations in river flow and wave energy. Their deposits often have a high water content and include highly mobile 'fluid muds' that are highly effective in dissipating incident wave energy (Mehta and Kirby 2001). In other environmental settings, coastal and marine sediment sources are more important. In the North Sea, for example, erosion of unconsolidated Quaternary cliffs provides a major source of muddy sediments along the coast of eastern England (Ke *et al*. 1996).

The intertidal zone of muddy coasts typically comprises: a lower zone, characterized by sandy tidal flats; a middle zone of muddy tidal flats; and an upper intertidal of vegetated saltmarsh or mangrove. In some localities, the upper intertidal grades into a high supratidal plain or flat, inundated only by extreme water levels (e.g. during storm surges). The low topography is dominated by low gradient surfaces, crossed by shallow tidal channels (or 'creeks'). These channels vary in complexity from single 'rills' to intricate networks, and are generally best developed within the mud flat and saltmarsh sub-environments. Surface sediments generally decrease in grain size in a landward direction and the vertical stratigraphic sequence generally exhibits a fining upward sequence, usually attributable to transitions between tidal flat and saltmarsh as sedimentation proceeds.

The physical processes of mud flat sedimentation have been extensively studied, mainly from the perspective of sediment transport and deposition, with rather less emphasis being placed on processes of deposit consolidation and erosion (Amos 1995). A reduction in tidal current velocities in a landwards direction leads to the deposition of sediment suspended during the flood tide: the diminution in the competence of flows to transport material also explains the landward reduction in grain size. Although a portion of the newly deposited material is resuspended on the ebb tide, vertical and horizontal accretion of muddy intertidal sediments implies the dominance of flood-tide deposition (Evans 1965).

In the absence of any net (or 'residual') landward water transport, the accumulation of mud is further explained by reference to the concepts of 'settling lag' and 'scour lag'. Both these concepts were developed in the 1950s to account for tidal flat sedimentation in the Dutch Wadden Sea (see Amos 1995 for a recent review and evaluation of this work). Settling lag refers to the time elapsed between the slackening of tidal current intensity below the threshold of suspension for a given sediment and the deposition of the particle on the bottom. This means that particles are deposited some distance landwards of the point at which settling from suspension commences. Scour lag is a consequence of the higher flow intensity required to re-entrain a particle once it has been deposited. This is especially important for cohesive sediments, and means that ebb-directed transport occurs over a shorter duration than that of the flood tide. Both mechanisms tend to encourage the landward transport of mud and its accumulation in shallow intertidal areas.

Rates of mud flat sedimentation may be initially rapid (of the order of several centimetres a year), but diminish as the build-up of elevation reduces the frequency of inundation. Colonization by halophytic vegetation (and a transition to saltmarsh or mangrove) may be associated with a further increase in sedimentation rate owing to increased sediment retention under an energy-dissipative plant cover. However, this rapidly diminishes as vertical accretion further reduces inundation and wetland surface elevations tend towards a state of equilibrium between further sedimentation, the compaction of earlier deposits and sea level.

Mud flat topography arises from the dynamic interaction of tidal and wave-related hydrodynamics, sedimentation and morphology itself. Recent work has shown wave action to be more important than previously thought and has also drawn attention to the importance of biological processes in mediating sediment stability. Pethick (1996) draws an analogy between the morphological adjustment of mud flats to variations in wave energy and the morphodynamics of non-cohesive sandy beaches. The influence of waves differs between inner estuary sites, subject to small (fetch-limited) waves and outer estuary or open coastal sites, which experience a greater range of wave heights. At fetch-limited sites, waves may still exert oscillatory shear-stresses which exceed those generated by tidal currents and which are capable of resuspending mud flat sediments. A zone of resuspension migrates up and down the mud flat profile with the tidal variation in water level. Over time, the mud flat profile adjusts towards a form that is in equilibrium with wave induced stresses. The resulting profile is typically concave, a finding supported by numerical modelling experiments undertaken by Roberts *et al*. (2000). At more exposed sites, mud flats may undergo more episodic erosional adjustments in response to high wave energy conditions. In this case, the balance between individual erosion events and depositional recovery in the intervening periods determines longer term mud flat morphology.

Predominantly accretional mud flats tend to have a high and convex profile, whilst erosional mud flats are typified by a lower and concave profile. Mehta and Kirby (2001) attribute the contrasting stability of these mud flat morphologies to differences in their dissipative characteristics. In the case of high, convex mud flats, flexing of water-sediment mixture substantially dissipates tidal and wave-induced stresses, especially where thin surficial fluid mud layers are present. In low, concave mud flats, however, deposits are normally overconsolidated, such that hydrodynamic stresses are dissipated in overcoming interparticle cohesion, and in entraining sediment. Such systems are likely to be erosional.

The surficial sediments of mud flats support a variety of organisms, some of which act to stabilize the sediment and some of which act to increase the likelihood of erosion. Most mud flats support dense communities of benthic diatoms, which excrete large quantities of extracellular polymeric substances (EPS). EPS consist mainly of polysaccharides compounds and are a major component of surface films, which increase the stability of the sediment surface (Paterson 1997). Meso- and macro-fauna are active over a greater depth and may variously stabilize sediment (e.g. through the construction of EPS-coated tubular burrows) or reduce stability (e.g. by grazing on the micro-algae which helps bind sediment particles, or by reworking sediments through burrowing). Biological processes are extremely variable, both spatially and temporally, and are extremely important in determining the threshold stress at which erosion occurs. Once this threshold is exceeded, however, erosion may proceed more rapidly at a rate more closely controlled by bulk sediment properties.

Mud flats are increasingly valued as a habitat for large invertebrate populations which, in turn, provide a vital food source for wading birds. As landforms they are also of engineering significance as naturally dissipative systems which, allied to fixed defences, can provide an important component of integrated and more sustainable strategies for coastal protection. From both these perspectives, high and convex mud flats are preferable to low and concave ones (Kirby 2000). In the former case, waves are progressively attenuated as they approach the shore, a process which is further assisted by any saltmarsh fringe. Convex mud flats tend to have a larger invertebrate fauna, concentrated at a higher elevation within the tidal range, and capable of sustaining greater bird and fish populations. In contrast, erosional concave mud flats are less effective in dissipating wave energy and, in their upper portions, prone to rotational failure and slumping, with adverse consequences for the stability of sea defences.

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SEE ALSO: mangrove swamp; saltmarsh; tidal creek; tidal delta

J.R. FRENCH

MUD VOLCANO

Mud volcanoes are positive topographic features formed by periodic venting of fluid mud, water and hydrocarbons (Kopf 2002). Individual mud volcanoes are elliptical mounds up to 2,000 m in diameter and 100 m in height. Cones and pools are often concentrated near the summit, and active portions of mud volcanoes are hummocky, unvegetated and covered by mud flows. Although clay and silt dominate mud-volcano deposits, pebble- to boulder-size clasts are common.

Mud volcanoes are known from approximately thirty regions worldwide. Most examples occur in compressional tectonic settings such as convergent plate margins. However, mud volcanoes also occur along passive margins and continental interiors. Excellent examples of subaerial mud volcanoes are present in Azerbaijan, Burma, Colombia, Indonesia, Iran, Italy, Mexico, Pakistan, Panama, Trinidad and Venezuela (Higgins and Saunders 1974). Subaqueous mud volcanoes occur in the Gulf of Mexico and the Barbados Ridge accretionary prism.

Mud volcanoes are commonly underlain by thick sequences of organic and clay-rich sediments (Hedberg 1974). Rapid sedimentation combined with methane generation, clay mineral diagenesis and tectonic compression produces high pore-fluid pressures, which mobilize fluid mud. Mud, water and hydrocarbons, migrate upward along fractures and faults that are typically associated with mud diapir-cored anticlines. If pressures are sufficient, fluid mud erupts at the seafloor or on the land surface to form mud volcanoes. In some instances, violent eruptions are accompanied by gas flares.

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SEE ALSO: diapir; liquefaction; mudlump

ANDRES ASLAN

MUDLUMP

A diapiric structure composed of fine-textured material, especially clay, formed near the mouth of a delta's distributary. Mudlumps range in size from pinnacles to small, elongated islands. They are both subaqueous and subaerial with the subaerial forms often subject to rapid erosion by waves. The surface of mudlumps is usually irregular and most have gas (methane) and mud vents. Although several theories have been proposed for their formation, it is now generally accepted that they are the result of the intrusion of plastic, prodelta clay through overlying sand layers. They develop in sequence as distributary mouths advance seaward. Originally thought to have been unique to the distributaries of the Mississippi River, mudlumps are now known to exist in a few other deltaic areas.

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H. JESSE WALKER

N

NATURAL BRIDGE

Remnant arch-shaped formation developed through erosion of the surrounding bedrock. Natural bridges, or stone arches, are unusual features that predominantly develop in horizontally bedded sedimentary rocks such as sandstone and limestone, though they hardly ever occur in metamorphosed or igneous rocks. They may form in a variety of ways, though all are ephemeral and will eventually collapse. The most common mode of formation is by water erosion, forming in deep valleys with highly sinuous rivers. Eventually, the river will cut across the neck of the entrenched meander by eroding a route through the obstructing rock outcrop. Often this can be accomplished without the arch collapsing, thus forming the natural bridge. Natural Bridge, Virginia, USA, has an uncertain evolutionary history, though meander cutting by the James River is a strong possibility (Malott and Shrock 1930). The other possible mode of formation is by the collapse of an underground drainage tunnel, leaving a remnant of the tunnel ceiling. Natural Bridge spans 30 m across Ceder Creek and is one of the few remaining natural bridges that is used as a transport bridge, at 60 m high.

Natural bridges may also be formed by the near complete collapse of underground tunnels. Such formations are common on the Hawaiian Islands, where recent lava tunnels roofed by a solidified crust may collapse leaving all but a small arch-shaped portion. Other origins of natural bridges include those cut by the sea resulting in coastal wave-cut arches, while a more unusual natural bridge can be found in Petrified Forest National Park, Arizona, USA, where a silicified tree trunk, known as the Onyx Bridge, spans a canyon 15 m wide.

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SEE ALSO: arch, natural

STEVE WARD

NEBKHA

Nebkha, or nabkha, is an Arabic term given to mounds of wind-borne sediment (sand, silt of pelletized clay) that have accumulated to a height of some metres around shrubs or other types of vegetation. They are sometimes called shrub-coppice dunes. They may occur on bigger dunes, in interdune areas, on pan surfaces, near wadis and on or behind beaches. Morphometric data are provided by Tengberg and Chen (1998). The largest nebkhas (mega-nebkhas) accumulate around clumps of trees. In the Wahiba Sands of Oman these can be 10 m high and up to 1 km long (Warren 1988).

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A.S. GOUDIE

NEEDLE-ICE

Needle-ice (synonymous to 'pipkrake' or 'kammeis') is the accumulation of ice crystal growths in the direction of heat loss at, or directly beneath, the ground surface. Although needle-ice usually grows perpendicular to the ground surface, curved ice-filaments are sometimes observed owing to wind and gravity effects. Needle-ice may also connect normal to plant stalks that have drawn sufficient ground moisture. Needle-ice is common to areas of diurnal freeze–thaw, ranging from tropical alpine to subarctic environments.

Needle-ice best develops in moist, fine-textured sediment with at least 10 per cent clay/silt. Precise soil and near-surface thermal dynamics affecting needle-ice growth and decay are complex, thus making it difficult to predict annual frequency. Generally, needle-ice develops within the first hour of ground temperatures dropping below 0 °C. Further conditions necessary for needle-ice development include a relatively low soil water tension to enable ice segregation to take place and adequately rapid movement of unfrozen moisture to the freezing front, so that it corresponds with the rate of latent heat loss, and thus preventing the *in situ* freezing of pore water (Outcalt 1971). Very windy conditions may cause a rapid temperature drop through the soil pores, thus reducing the suction gradient and enhancing the development of pore ice rather than needle-ice. Typically, needle-ice phases will entail periods of growth, stagnation and ablation. The duration of freeze determines growth phases and consequently needle-ice length, which may vary from a few millimetres to several centimetres. Polycyclic or multilayered needle-ice, separated by thin veneers of sediment, occurs where there has been moisture stress. Alternatively, long-lasting growth phases over several days may produce multilayered needle-ice lengths exceeding 400 mm.

Needle-ice has been applied to studies examining SOIL CREEP, SOIL EROSION, the impacts on plant (particularly seedling) disruption and its function as a geomorphic process in miniature landform development. Needle-ice on stream banks or soil terraces commonly extrudes sediment, which is transferred by needle-ice induced direct particle fall, sliding and toppling failure and mini-mudflows. Geomorphological consequences of needle-ice as an erosion agent include notches and undercut fluvial banks, TURF EXFOLIATION and associated depositional microforms. Several soil structures including nubbin soils, gaps around stones and other varieties of PATTERNED GROUND, have been attributed to needle-ice. It is thought that soil stripes aligned parallel to the late morning sun may be a function of shadow and differential thaw effects during the ablation phase.

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SEE ALSO: freeze–thaw cycle; frost heave; ice

STEFAN GRAB

NEOCATASTROPHISM

Neocatastrophism, as defined by Schindewolf (1963), refers to the explanation of sudden extinctions in the palaeontological record. In geomorphology, George Dury (1975, 1980) expressed the view that high magnitude, low frequency events were more important in an absolute sense than low magnitude, high frequency events in moulding the Earth's landscapes. Dury's statement expresses the essence of the issue. Neocatastrophism is a response to one hundred years of geomorphic thinking in which the predominant role of low magnitude, high frequency events in landform evolution had become the prevailing paradigm. A side issue, expressed in an exchange between Brunsden (1996) and Yatsu (1996), is whether the word catastrophism should be excised from the geomorphic vocabulary, and hence, by implication, also the word neocatastrophism. I am not convinced that we need to fear this word; but there is a need for unambiguous definition. By contrast with catastrophism, which is an outmoded, pre-twentieth century mode of thought, neocatastrophism is thought to be an increasingly relevant way of viewing the geomorphic world.

Circumstances which have favoured the emergence of neocatastrophism include the following:

- improved precision in geochronology has demonstrated unexpectedly rapid past changes;
- the exploration of mass extinctions in the past has intensified;
- some geomorphological features, such as the Channeled Scablands of eastern Washington, are more amenable to explanation by low frequency, high magnitude events than by gradual, semi-continuous processes;
- space exploration has generated a strong interest in galactic scale events;
- interest in global environmental change has provided evidence of rapid past changes, such as found in the polar ice caps and the oceanic deep sediments;
- the rise of non-linear dynamics and chaos theory is beginning to provide ways of synthesizing gradualism and catastrophism.

Within geomorphology, it was the paper by Wolman and Miller (1960) which provoked a critical evaluation of the question of magnitude and frequency (see MAGNITUDE–FREQUENCY CONCEPT) of the operation of geomorphic processes. The authors directed attention to the importance of medium size and medium frequency events as having the greatest cumulative influence on the landscape. This was an important insight, but did not prevail after notable discussions by Wolman and Gerson (1978), Gould (1984), Gretener (1984) and Baker (1994).

Wolman and Gerson (1978), in following up their findings on magnitude and frequency, expanded on effectiveness of climate and relative scales of time such that they were forced to modify their view about the importance of the intermediate magnitude and intermediate frequency event in landform history. Introduction of the idea of the length of time over which a landform survived suggested that, in many cases, it was the extreme events which were most important. The influence of this paper on geomorphic thinking cannot be overestimated as it has emphasized the importance of combining measures of process magnitude and frequency with duration or lifetime of landforms.

Gould's discussion on punctuational change was an equally influential paper for the whole of Earth science (Gould and Eldredge 1977). The essential concept was a recognition that many important changes in Earth history have proceeded by relatively rapid flips between more stable conditions. Systems often absorb stress and resist change until the stresses accumulate past a breaking point. Systems then flip to a new stable state. This hypothesis, known as punctuated equilibrium, has gained widespread acceptance in the palaeontological community and it is viewed by other Earth scientists as a model for processes of inorganic change. Gretener (1984) advances the example of isostatic rebound to illustrate the relativity of gradualism. Isostatic rebound has been active during the last 10,000 years and is still in progress in such places as northern Canada and Scandinavia. The process covers all of humanity's conscious history and is generally perceived as a gradual phenomenon. However, if one considers that the Earth's skin can completely recover from the unloading of 1–2 km of ice within a period of 15–20,000 yrs, this process is effectively instantaneous from an Earth history perspective. This leads to a consideration of what constitutes an event? Gretener suggests that the duration of an event occupies no more than 1/100 of the total time span being considered. On this basis, geological processes may have durations as great as 10 Ma and still qualify as events. Indeed Earth history 'reveals long periods of tranquility interrupted by moments of action' (Gretener 1984: 86). The rare event in geology is a punctuation with such a low rate of occurrence that it has taken place, at most, a few times through all of Earth history. It is unscientific to call such events 'impossible'. Punctuationism would possibly be a better term than neocatastrophism. Nevertheless, either term is preferred to uniformitarianism, which fails to do justice to such extreme events. Brunsden (1996) illustrates Gretener's point well in his Figure 2.3.

Baker (1994) provides the most powerful geomorphic justification for neocatastrophism in his interpretation of the resistance of the geological community to Harlen Bretz's (1923) theory of the origin of the Channeled SCABLANDS of eastern Washington. He explains that the community was blinkered by its slavish adherence to gradualism and its suspicion of the mention of cataclysmic flood events. Nevertheless, Bretz's interpretation was finally vindicated in many of its neocatastrophist details, in part as a result of the identification of a source of this exceptional flooding (glacial Lake Missoula) which Bretz himself had not recognized, but also in part because of the recognition of the erosional and hydraulic concomitants of an extreme flood.

There remains an urgent need for geomorphologists to accommodate our thinking to the new diastrophic ideas associated with global plate tectonics. The focus on short timescales relevant to process studies has been partly responsible for a neglect of the longer timescales. Modes of vertical motion, the onset of Ice Ages and the appearance of volcanism all need to be reappraised in a neocatastrophist framework.

Thorn (1988) points out that there is an important intellectual issue associated with the rise of neocatastrophism. If greater significance is being attached to large events in a series, this only forces an adjustment of magnitude–frequency concepts. If the new perspective is one that identifies unique events as paramount in geomorphological records, then there can be no science of geomorphology because there is no science of uniqueness. Most of us are busily adjusting our magnitude–frequency concepts.

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SEE ALSO: catastrophism; magnitude–frequency concept

OLAV SLAYMAKER

NEOGLACIATION

Neoglaciation is a geological term, originating in North America, used to describe the period during the latter half of the Holocene when valley GLACIERS in many mountain areas readvanced to their maximum extent following Pleistocene DEGLACIATION. The term was first used by Moss (1951) and Nelson (1954) who attribute it to Matthes (though the term appears in none of his papers). Neoglaciation was formally defined by Porter and Denton (1967) as a 'cool geologic-climate unit . . . indicating a probable world wide synchrony of glacier fluctuations in response to climatic change'. Their classic paper established the standard division of the North American Holocene into a warmer and drier early Holocene (the Hypsithermal) followed by a cooler Neoglacial Interval characterized by several periods of glacier advance. The related term 'little ice-age' was first used by Matthes (1939) to define the period when glaciers re-established in the Sierra Nevada of California following the post-glacial climatic optimum. Matthes's 'little ice-age' was, in fact, what is now termed Neoglaciation. Subsequently, the term Little Ice Age (LIA) has been almost universally adopted to describe the latest and most severe part of the Neoglacial during the past few centuries when glaciers in many areas of the world reached their maximum Holocene extent (Grove 2003).

This terminology was established at a time when there were few detailed chronologies of Holocene glacier fluctuations with little absolute dating control (radiocarbon dates were just becoming generally available to Quaternary scientists). Holocene climates were interpreted on the basis of limited evidence from studies of glacier fluctuations and the zonation of pollen diagrams in Europe and North America. As the maximum Neoglacial (Little Ice Age) extent of glaciers at most northern hemisphere sites was between AD 1600 and 1850, almost all morphological evidence of earlier glacier events was destroyed. Stratigraphic evidence from lateral MORAINES and sections within the Little Ice Age limits was fragmentary, difficult to find and the dating often poorly constrained.

Over the past thirty to forty years new information has led to the modification of our knowledge and understanding of these glacial events. Significant glacier recession during the late twentieth century has exposed many new moraine sections and buried forest sites that yield detailed evidence of earlier glacier fluctuations. The advent of AMS and calendar-adjusted radiocarbon dates, plus dendrochronological dating of sub-fossil wood using millennial-length tree-ring reference chronologies, and the development of proximal varve sequences have improved the available record of dated Neoglacial sequences (Plate 83).

Most evidence of Neoglacial glacier fluctuations comes from western North America and western Europe where, generally, the LIA glacier advances were the most extensive. However, in the southern hemisphere several authors have identified deposits of an early Neoglacial advance *c.*4,400–4,600 yr BP, downvalley of the LIA limits. This evidence is critically reviewed by Porter (2000) who cautions that this conclusion should remain provisional until a larger population of better dated sites are available.

Early work in the northern hemisphere (mainly in Alaska and Scandinavia) identified three phases of Neoglaciation: early (*c.*6,000–4,000 BP), middle (2,500–3,500 BP) and late (last 1,000 years or LIA) with a minor event *c.* AD 700–900 (Denton and Karlen 1973). Evidence for the earliest events is fragmentary. Most investigations date initial Neoglacial advances between 4,000–5,000 ^{14}C yr BP and link them with other PALEOCLIMATE evidence of climate deterioration at this time. Although the preceding Hypsithermal was originally defined as a time stratigraphic unit (Porter and Denton 1967), dates for the Hypsithermal–Neoglacial transition are clearly time transgressive with evidence for some alpine glacier advances before 6,000 BP. Therefore the early Neoglacial is not well defined. There is widespread evidence for glacier advances between *c.*3,500–2,800 ^{14}C yr BP in the Canadian Rockies, Alaska, Switzerland, Patagonia and Scandinavia. There is also evidence from several areas of glacier advances *c.*1,300–1,500 ^{14}C yr BP and an 'early medieval advance' *c.*AD 600–800. However, the most detailed (and best dated) reconstructions of glacier fluctuations from the Alps (Holzhauser 1997) indicate that at least seven advances of the Aletsch Glacier occurred between 3,200 yr BP and AD 1000, plus three major LIA advances. It seems unlikely that the history of glacier fluctuations at less well-dated sites is any less complex than that shown by the Aletsch. Therefore the history of glacier fluctuations during the early and middle Neoglaciation remains incomplete but probably consists of multiple, relatively short-lived (50–200 years?) advances that appear to have been progressively more extensive over time and were separated by periods of glacier recession. In assessing the synchroneity of these events it is critically important to determine both the precision of the dating technique used and the precision of dating control (i.e. its stratigraphic or geographic relationships with the event being dated). In many cases the limiting dates are +/−50 years at best which is often inadequate to differentiate between synchronous and closely spaced events or determine whether the events are correlative and synchronous over large areas rather than simply locally significant records.

The beginning of LIA is traditionally placed at the end of the Medieval Warm Period. The MWP was initially defined from non-glacial evidence in Europe (Hughes and Diaz 1994) and encompasses the period AD 800–1200 when there is little evidence of extended glacier cover. The status of this period as a global interval of generally warmer conditions remains questionable until an adequate database of high resolution palaeoclimate records becomes available. Well-dated, early LIA glacier advances occurred in the twelfth to

Plate 83 Late Neoglacial (Little Ice Age) lateral and terminal moraines, Bennington Glacier, British Columbia, July 1990

fourteenth centuries in Patagonia, Canadian Rockies, Alaska, Switzerland and Scandinavia. These early advances were followed by an interval with little evidence of glacier fluctuations until the main LIA advances, dated between the sixteenth and nineteenth centuries. In many areas glaciers reoccupied positions at or very close to their maxima several times during the LIA, e.g. the early 1700s and mid-1800s in the Canadian Cordillera and coastal Alaska or the 1350s, 1650s and 1850s in the Alps. Most glaciers have receded rapidly during the twentieth century. The exposure of old buried forest sites and the alpine iceman suggests that this twentieth-century recession is the most rapid and severe during the Holocene. However, minor advances of glaciers occurred in many alpine areas during the 1960s–1970s and glaciers in western Norway advanced significantly during the 1980s and early 1990s as a result of increased winter precipitation due to changes in atmospheric circulation. In summary, these records indicate several intervals of glacier expansion over the past 5,000 years. Some, such as the nineteenth century, appear to be globally synchronous (at least at the centennial scale) whereas others may reflect local or more regional glacier histories.

The development of independent, high-resolution proxy climate records spanning the late Holocene (using tree rings, ice cores and other techniques) has greatly expanded our knowledge of climate variability and climate history. This work has shown that climate varies continuously at several spatial and temporal scales; that the relationships between glacier fluctuations and climate are complex; and that climate variability is rarely synchronous at the global scale. Recent palaeoclimate work also provides superior records of climate forcing. Some forcing factors have globally synchronous effects (e.g. variations in solar output, sunspot minima, etc.) whereas the effects of others may differ between hemispheres (orbital effects, volcanic eruptions) or between regions (e.g. circulation changes). Global climate variability reflects the interaction of all these factors as do climatically dependent fluctuations of glaciers. However, despite these strong links to climate, glacier behaviour and response times are also influenced by many factors unrelated to climate. The use of glacier-defined terms such as Neoglaciation and Little Ice Age to identify distinct, global, climate-geologic periods is inappropriate and misleading in the context of current knowledge of Holocene climates. Usage of these terms should be confined to describing the glacial advances of the late Holocene after *c.*5,000 BP and between *c.*AD 1000–1900, respectively.

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SEE ALSO: dating methods; dendrochronology; Holocene geomorphology; palaeoclimate

BRIAN LUCKMAN

NEOTECTONICS

Neotectonics concerns the study of horizontal and vertical crustal movements that have occurred in the geologically recent past and which may be ongoing today. While most crustal movements arise directly or indirectly from global plate motions (i.e. tectonic deformation), neotectonic studies themselves make no presumption about the mechanisms driving deformation. Consequently here 'movements' is a vague catch-all term that encompasses a myriad of competing deformation processes, such as the gradual pervasive creep of tectonic plates, discrete (seismic) displacements on individual faults and folds, and distributed tilting and warping through isostatic readjustment or volcanic upheaval. The phrase 'geologically recent past' is also intentionally vague. Early attempts to define the discipline by arbitrary time windows such as Late Cenozoic, Neogene or Quaternary have given ground to a more flexible notion that envisages neotectonism starting at different times in different regions. The onset of the neotectonic period, or the 'current tectonic regime', depends on when the contemporary stress field of a region was first imposed. For instance, in the Apennines of central Italy the 'current tectonic regime' began in the Middle Quaternary (~700,000 years ago) and it is even younger (< 500,000 years) in California; in contrast, in eastern North America the present-day stress regime has been in existence for at least the last 15 million years.

Typically then, neotectonic movements have been in operation in most regions for the last few million years or so. Over such prolonged intervals, neotectonic actions are revealed by the stratigraphical build-up of sediments in inland and marine basins, the burial or exhumation histories of rocks and the geomorphological development of landscapes. Geological studies of palaeobotany and palaeoclimate, numerical models of landscape evolution and techniques such as fission track analysis and cosmogenic dating are among the disparate tools unravelling this long-term tectonic activity. Over periods of many tens of, to several hundred thousand years, the actions of individual tectonic structures (faults and folds) can be determined, unmasked by their deformation of geomorphic markers, such as marine and fluvial terraces, and tracked with reference to the late Pleistocene glacial-eustatic time frame. The apparently smooth deformation rates discerned over intermediate timescales are revealed to be episodic and irregular when faults and folds are examined over Holocene (10,000 years) timescales. Over millennial timescales, secular variations in the activity of tectonic structures can be gleaned from a diverse set of *palaeoseismological* approaches, from interpreting the stratigraphy of beds that have been affected by faulting to detecting disturbances in the growth record of trees or coral atolls.

Although neotectonic movements continue up to the present day, the term *active tectonics* is typically used to describe those movements that have occurred over the timespan of human history. Active tectonics deals with the societal implications of neotectonic deformation (such as seismic-hazard assessment, future sea-level rise, etc.), since it focuses on crustal movements that can be expected to recur within a future interval of concern to society. Even contemporary crustal movements may reveal themselves in Earth surface processes and landforms, such as in the sensitivity of alluvial rivers to crustal tilting. In addition, geomorphological and geological studies are important in recording the surface expression of Earth movements such as earthquake ground ruptures which, due to their subtle, ephemeral or reversible nature, are unlikely to have been preserved in the geological record. However active tectonics also employs an array of high-tech investigative practices; prominent among these are the monitoring of ongoing Earth surface deformation using space-based or terrestrial geodetic methods (tectonic geodesy), radar imaging (interferometry) of ground deformation patterns produced by individual earthquakes and volcanic unrest, and the seismological detection and measurement of earthquakes (seismotectonics), both globally via the World-Wide Standardised Seismograph Network and regionally via local seismographic coverage. These modern snapshots of tectonism can be pushed back beyond the twentieth century through the analysis of historical accounts and maps to infer past land surface changes or deduce the parameters of past seismic events (historical seismology). In addition, earthquakes can leave their mark in the mythical practices and literary accounts of ancient peoples, the stratigraphy of their site histories, and the damage to their buildings (archaeoseismology). The time covered by such human records varies markedly, ranging from many thousands of years in the Mediterranean, Near East and Asia to a few centuries across much of North America. Generally they confirm that regions that are active today have been consistently active for millennia, thereby demonstrating the long-term nature of crustal deformation, but occasionally they reveal that some regions that appear remarkably quiet from the viewpoint of modern seismicity (such as the Jordan rift valley) are capable of generating large earthquakes.

In reality, the distinction between neotectonics and active tectonics is artificial; they simply describe different time slices of a continuum of crustal movement. This continuum is maintained by the persistence of the contemporary stress field, which means that inferences of past rates and directions of crustal movement from geological observations can be compared directly with those measured by modern geodetic and geophysical methods. Although the terms 'neotectonic' or 'active' are somewhat blurred and are often used interchangeably, societal demands (for instance, regulatory authorities for seismic hazard, nuclear safety, etc.) often require the incidence of tectonic movements to be strictly defined. For instance, the present definition in Californian law of an 'active fault' is one that has had surface-rupturing earthquakes in the last 11,000 years (established when the Holocene was considered to have begun at that time) (see ACTIVE AND CAPABLE FAULT). Other regulatory bodies recognize a sliding scale of fault activity: Holocene (moved in the last 10,000 years), Late Quaternary (moved in the last 130,000 years) and Quaternary (moved in the last 1.6 million years). Neotectonic faults, by comparison, are simply those that formed during the imposition of the current tectonic regime. 'Real' structures, of course, are unconstrained by such legislative concerns. Many modern earthquakes rupture along older (i.e. palaeotectonic) basement faults. Indeed, it is important to recognize that any fault that is favourably oriented with respect to the stress currently being imposed on it has the potential to be activated in the future, regardless of whether it has moved in the geologically recent past.

A more meaningful way to differentiate styles and degrees of neotectonic activity is in terms of tectonic strain rate, which is a measure of the velocity of regional crustal motions and, in turn, of the consequent tectonic strain build-up. Crustal movements are most vigorous, and therefore most readily discernible, where plate boundaries are narrow and discrete. In these domains of high tectonic strain, frequent earthquakes on fast-moving (>10 mm/yr) faults ensure that a century or two of historical earthquakes and a few years of precise geodetic measurements are sufficient to capture a consistent picture of the active tectonic behaviour. Intermediate tectonic strain rates characterize those regions where plate-boundary motion is distributed across a network of slower moving faults (0.1–10 mm/yr). Examples of such broad deforming belts are the Basin and Range Province of western USA or the Himalayan collision zone, where earthquake faults rupture every few hundred or thousand years, ensuring that the Holocene period is a reasonable time window over which to witness the typical crustal deformation cycle. In contrast, low-strain rates ensure that intraplate regions, often referred to as 'stable continental interiors', are low-seismicity areas with slow-moving (< 0.1 mm/yr) faults that rupture every few tens (or even hundreds) of thousands of years, making the snapshot of human history an unreliable guide to the future incidence of tectonic activity.

The global pattern of present-day crustal motions can be accounted for by PLATE TECTONICS theory, that elegant kinematic framework in which rigid plates variously collide, split apart and slide along their actively deforming boundaries. Closer inspection, however, reveals that the basic rules that govern global plate motions (i.e. rigid blocks separated by narrow deforming boundaries) break down at the regional and local scale. This is particularly so on the continents, where a patchwork of pre-existing geology and structure ensures that tectonic stresses are not applied in a uniform, straightforward fashion. Studies of how the contemporary stress field varies across the Earth's surface (Figure 108) distinguish between first- and second-order stress provinces. First-order provinces have stresses generally uniformly oriented across several thousands of kilometres. The largest of these are the midplate regions of North America and western Europe, where the stress fields are largely the far-field product of ridge push and continental collision. In contrast, first-order stress provinces in tectonically active areas are dominated by the downgoing pull of subducting slabs and the resistance to subduction. Second-order stress provinces are smaller, typically less than 1,000 km across, and are related to crustal flexure induced by thick sequences of sediments and postglacial rebound, and to deep-seated rheological contrasts. Although the bulk of the Earth's crust is in compression, significant regions of extension occur. In both the continents and oceans, these extensional domains are long and narrow and correspond to topographically high areas, though notable exceptions are the Basin and Range province and the Aegean region of the eastern Mediterranean. Most first-order stress provinces, and many second-order stress provinces coincide with distinct physiographic provinces.

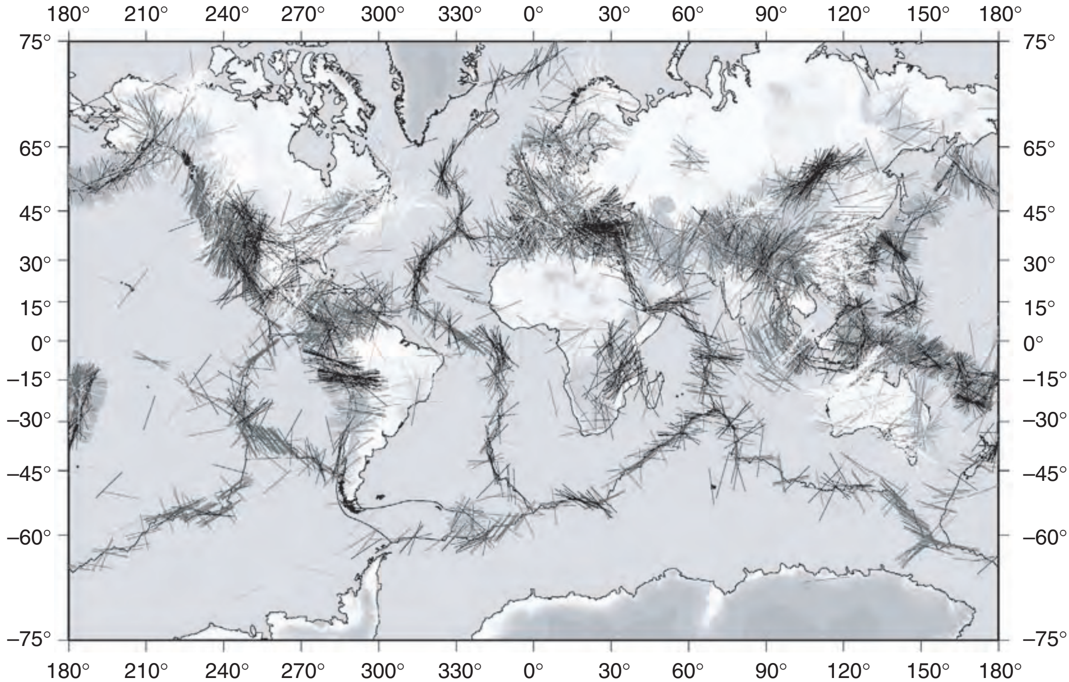

Figure 108 The World Stress Map with lines showing the directions of maximum horizontal compression. Black lines denote normal faulting (extension), dark grey lines denote strike-slip faulting, and light grey lines denote thrust faulting (compression); white lines show an uncertain tectonic regime. The longer the line length, the better the quality of the data. Around two-thirds of the stress data come from earthquakes and so highlights where the bulk of tectonic deformation is occurring; most of the remaining third comes from borehole stress measurements that are concentrated in petroleum-producing provinces. From Mueller *et al.* (2000)

Plate driving forces may exert the dominant control on the contemporary stress field, but another process contributes to crustal deformation at a global scale. That process is glacial isostatic adjustment (GIA), the physical response of the Earth's viscoelastic mantle to surface loads imposed and removed by the cycles of glaciation and deglaciation to which the planet has been subjected for the past 900,000 years (see GLACIAL ISOSTASY). Because large ice-mass fluctuations induce the sub-crustal flow of material, measurable crustal deformation extends for thousands of kilometres beyond the limits of the former ice margins (Figure 109); in short, the effects of GIA are felt globally. In addition, while the crust's elastic response to ice-sheet decay is geologically immediate, the delayed viscoelastic response of the mantle ensures that GIA persists long after the ice has gone. Although the effects of GIA can now be detected from space geodesy, its legacy is most clearly visible in the worldwide pattern of postglacial sea-level changes. Regions that were ice covered at the Last Glacial Maximum are uplifting (i.e. relative sea level is currently falling) as a consequence of postglacial rebound of the crust. Likewise the regions peripheral to the former ice sheets are subsiding (i.e. relative sea levels are rising) due to collapse of the 'glacial forebulge'. The effect of this subsidence outside the area of forebulge collapse is to draw in water from the central ocean basins, which is compensated by uplift in the ocean basin interiors in the far-field of the ice sheets. The final GIA component is the hydro-isostatic tilting of continental coastlines due to the weight applied to the Earth's surface by the returning meltwater load, which produces a 'halo' of weak crustal subsidence around the world's major land masses. For the most part, geological studies of Holocene relative sea-level changes are consistent with the uplift/subsidence

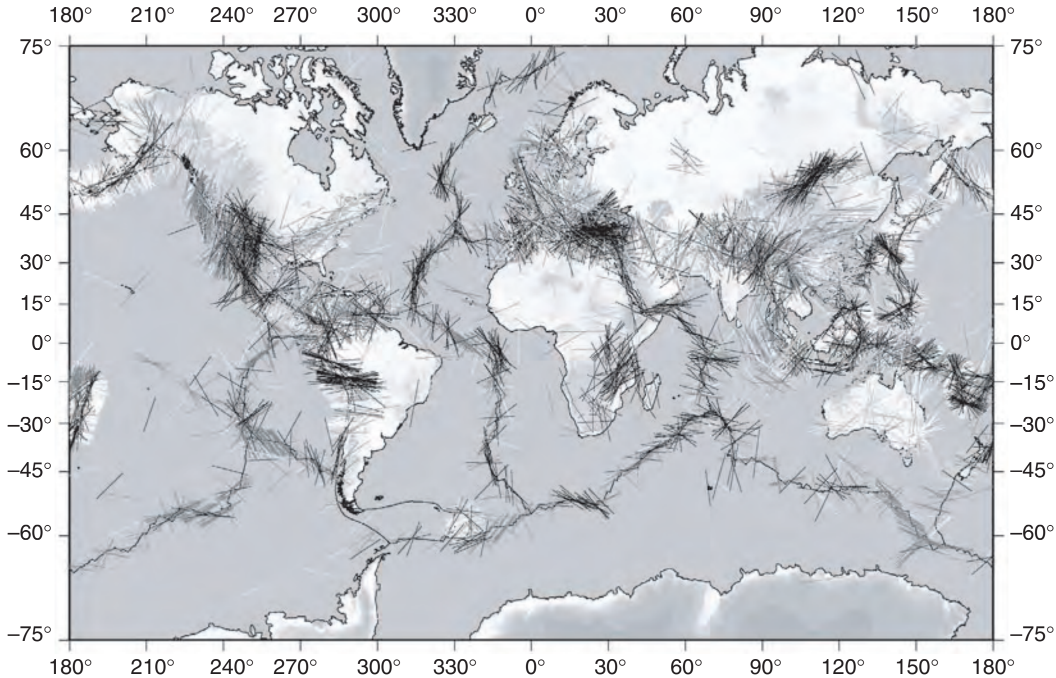

Figure 109 Map showing the outward radial motion of eastern North America (predicted by Peltier's (1999) postglacial rebound model) due to glacial isostatic adjustment to removal of the Laurentide ice sheet, and highlighting the concentration of contemporary seismicity along the former ice margins. From Stewart *et al.* (2000)

pattern predicted by global viscoelastic theory. The key areas of misfit are along plate boundary seaboards (especially subduction zones), where tectonic deformation dominates, and those areas 'contaminated' by local anthropogenic effects (groundwater extraction etc.).

The neotectonic implications of GIA are not confined to the coastline. Glacial rebound is now widely considered as an effective mechanism for exerting both vertical and horizontal stresses not only within the limits of the former ice sheets but for several hundred kilometres outside. Within the former glaciated parts of eastern North America and northern Europe both tectonic and rebound stresses are required to explain the distribution and style of both postglacial and contemporary seismotectonics. Outside in the ice-free forelands, predicted glacial strain rates are still likely to be one to three orders of magnitude higher than tectonic strain rates typical of continental interiors. Consequently, some workers argue that an apparent 'switching on' of Holocene earthquake activity in eastern USA and the occurrence of atypically large seismic events such as the great ($M > 8$) earthquakes that struck the Mississippi valley area of New Madrid in 1811–1812 may be associated with areas where glacial strains are particularly high. Glacial loading and unloading may also disturb the build-up of tectonic strain at glaciated plate boundaries, such as today in Alaska or previously when the Cordilleran ice sheet capped part of the Cascadia subduction zone. More recently, the isostatic component of glacier erosion in the mountain-building process is becoming appreciated.

In summary, the worldwide pattern of vertical and horizontal crustal movements arise from the global effects of plate motions and glacial isostatic adjustment. Regionally and locally, this is augmented by flexure from eustatic or sediment loading, volcanic deformation or anthropogenic change (dam impoundment). While many neotectonic investigations seek to disentangle movements arising from the imposition of tectonic strains from those augmented by non-tectonic processes, this is often a fruitless holy grail; because deformation of the Earth's crust typically induces compensatory flow underlying mantle, neotectonic movements are applied globally. Nevertheless, these disparate contributory mechanisms, coupled with the varying timescales over which their actions can de discerned, ensure that neotectonics encompasses a remarkable breadth of research disciplines. Few other fields easily blend topics as disparate as space science, seismology, Quaternary science, geochronology, structural geology, geomorphology, geodesy, archaeology and history. It is this interdisciplinary marriage that makes neotectonics particularly exciting and especially challenging.

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IAIN S. STEWART

NIVATION

Nivation is a morphogenetic term introduced by Matthes (1900) to describe and explain the processes associated with late-lying seasonal snow patches and landforms derived from them (nivation benches or terraces, and nivation hollows). The term became entrenched in periglacial geomorphology with little attention to process measurements until recently. One important vein of thinking envisages nivation hollows as precursors of glacial cirques.

While Matthes (1900) fails to produce a sharp definition of nivation, he exhibits a sophisticated grasp of snowpack accumulation dynamics. He invokes static snowpacks with intensified freeze–thaw around snowpatch peripheries, but assigns nivation only modest powers of landscape modification. Furthermore, while Matthes identifies a form continuum from nivation hollow to cirque, he distinguishes sharply between nivation and glacial effects and does not claim that nivation hollows enlarge into cirques. Nivation was soon adorned by others with bedrock freeze–thaw weathering, nivation hollows as precursors of cirques, the mobility of snowpacks, and solifluction as the primary mass wasting process. Thorn (1988) provides a comprehensive review of the development of nivation into the 1980s. The fundamental issue is to appreciate that nivation is a concept of weathering and transport intensification that invokes no unique processes.

Nivation benches or terraces are idealized as a gentle sloping flat or tread mantled in debris, unvegetated where snow is especially late-lying, with a steeper riser at the upslope end. Expansion is by headward incision promoted by the presence of late-lying snow at the inflection of slope. When suitably oriented such landforms, whatever their origin, are highly likely to become snow accumulation sites and it becomes tempting, if not irresistible, to move from correlation to causation, a step fraught with problems in the absence of process measurements. Nevertheless, available evidence does suggest that nivation is a likely mechanism for expansion in poorly consolidated materials (Thorn 1988; Berrisford 1992; Christiansen 1998). Where such forms occur in bedrock, headward incision becomes dependent upon more contentious weathering processes, rather than merely upon excavation by mass wasting.

While it is easy to envisage that the additional water supply associated with late-lying snow has considerable geomorphic potential, detailed field measurements were not undertaken until the 1970s (Thorn 1988). Important subsequent work includes Berrisford (1991, 1992) and Christiansen (1996, 1998). A reasonable summary of present knowledge of the mass wasting component of nivation is to state that late-lying snow does indeed accelerate or intensify periglacial mass wasting processes (e.g. solifluction, surface wash) by several factors, even by orders of magnitude, in comparison to nearby snow-free (or thinly snow-covered) surfaces.

However, the literature is not adequate to specify a consistent pattern of process or process rate intensification; indeed, considerable variability appears likely. On unconsolidated surfaces the elimination of vegetation cover by late-lying snow (not always the case) appears to represent an important process threshold. As rainfall inputs decrease and snowfall inputs increase proportionally, snowpatch meltwater influences emerge more starkly, especially in poorly consolidated materials (Christiansen 1998). In the presence of permafrost snowpatches may have important impacts on near-surface water flow (Ballantyne 1978), particularly by raising shallow subsurface flow to the surface where a snowpatch sustains a frozen subsurface.

While the role of nivation within the periglacial transport suite is generally non-problematic, and increasingly emphasizes meltwater impacts, the weathering role of nivation is problematic. For much of its history nivation was generally assigned no chemical weathering role. Intensification of chemical weathering processes beneath and around snowpatches is now documented (e.g. Thorn 1988), with a spatial pattern strongly dependent upon meltwater pathways that may even shift the impact downslope of the snowpatch itself. Knowledge of the role of nivation as a modifier of freeze–thaw weathering is largely constrained by the uncertainties associated with freeze–thaw weathering itself (Hall *et al*. 2002). Relevant ground climates (as opposed to largely irrelevant generalized air climates) are poorly known, laboratory studies do not necessarily mimic field conditions adequately, nor do they effectively isolate freeze–thaw from other possible mechanisms. Within the immediate context of nivation the critical issues rest with the interaction between snowpack insulation modifying, and perhaps eliminating, sufficient thermal regimes versus the obvious addition of abundant and necessary moisture through snowmelt. Berrisford (1991) found morphological evidence in the form of angular clasts beneath some portions of snowpatches to suggest enhanced mechanical weathering of coarse debris. However, he also emphasized the geomorphic importance of the annual temperature cycle, as opposed to shorter cycles, and views perennial snowpatches as protective.

Unlike CRYOPLANATION, of which it is a critical component, nivation research has been reinvigorated in recent years. While field data is increasingly available, definitional problems continue. Thorn (1988) suggested the term is so broad that it will always defy definition and should be abandoned, while Christiansen (1998) would like to expand it to embrace all snow-related processes making it equivalent to 'glaciation' in generality. Perhaps neither path is advisable, but the sharp contrast serves to highlight the problems presently associated with the term.

Snow-derived process will always be central to periglacial geomorphology and lead to some broad issues (Thorn 1978). Most nivation researchers appreciate that the wind-derived nature of a snowpatch means that there is potential for snow-bearing winter winds to orient landscape development through nivation. In fact, such orientation passes through a second filter, namely, available, suitable topographic traps because deep, late-lying snow cannot accumulate on a flat surface. Such concepts lead to ideas focused upon the landforms produced by snow-dominated regimes as opposed to those of full glaciation. Nelson (1989) goes so far as to suggest that not only is nivation central to cryoplanation, but that cryoplanation terraces are periglacial analogues of glacial cirques. His thesis invokes the presence of cryoplanation terraces where snowfall and temperature regimes are inadequate to generate cirque glaciation. Yet another view of nivation juxtaposes it with cold-based, that is non-erosive, glaciation. In such a context Rapp (1983) has suggested that interglacial nivation may represent the erosive, landforming regime and glaciation the quiescent, protective one.

The spatial extent of seasonal snowcover and lengthy interglacial periods, perhaps even more importantly relatively short pleniglacial periods, suggest that geomorphic processes derived from late-lying snow merit considerable attention. Clearly, nivation represents a core concept in such an appreciation, albeit not an exclusive one. Intensification of surficial mass wasting processes by nivation is now well established, but determination of systematic trends and rates must await generation of considerably more data. The weathering regime associated with nivation remains uncertain. Concentration of snowpatch meltwaters promotes chemical weathering; however, the location and degree of mechanical weathering with respect to a fluctuating seasonal snowpatch remains questionable, despite widespread willingness to invoke it. The extent to which nivation is able to shape a landscape is simply unknown.

Nivation cannot seemingly initiate topographic lows, but it can certainly modify them – but to what extent, in what fashion, and at what rates? Christiansen (1998) demonstrates headward expansion of nivation hollows in soft materials infused with permafrost, Berrisford (1992) suggests downslope expansion of nivation hollows. Thorn (1976) calculated nivation excavation rates in colluvium that would not produce a cirque from a nivation hollow in a feasible period of time. Quantitative research into such topics is urgently needed, but is immediately confronted by scale-linkage issues. In particular, periglacial process studies conducted on mesoscale phenomena must contend with the possibility that PARAGLACIAL conditions, rather than prevailing ones, hold the key.

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SEE ALSO: cryoplanation; freeze–thaw cycle

COLIN E. THORN

NIVEO-AEOLIAN ACTIVITY

Niveo-aeolian processes involve the entrainment, transport and deposition of fine (mainly sand-sized) particles by wind in seasonally snow-covered areas, and modification of such sediments during snowmelt. The defining characteristic of niveo-aeolian activity is the deposition of wind-blown sediment on snowcover. This involves either simultaneous deposition of mixed sediment and drifting snow, or deposition of windblown sediment alone over earlier snowcover. Sediments deposited by wind on snowcover are referred to as niveo-aeolian deposits. The term *denivation* is used to refer to the processes, microforms and sedimentary structures associated with melting of the underlying snowpack.

Niveo-aeolian activity occurs in polar deserts, in subarctic environments, on alpine plateaux (Ahlbrandt and Andrews 1978) and on maritime mountains (Ballantyne and Whittington 1987). Most niveo-aeolian deposits are annual (associated with complete melting of snow in summer), though in exceptionally cold arid areas of Antarctica perennial deposits consisting of inter-statified sediment and snow occur. Elsewhere, buried snow may persist under niveo-aeolian deposits for one or more summers (Bélanger and Filion 1991).

Sources of niveo-aeolian sediments include unvegetated or partly vegetated floodplains or outwash plains, raised beaches and deltas, aeolian sandsheets and dunes, glacial deposits, and sandy regolith. Sediments are entrained by wind in autumn or spring when snowcover is incomplete, or during winter when strong winds strip snow from the crests of ridges or hummocks. Sublimation of pore ice and abrasion by blowing sand are important in releasing sand particles from frozen surfaces. Most sand-sized particles travel by saltation or creep over ice or crusted snow, with fine sand and silt particles travelling in suspension. During violent storms coarse granules may travel up to 4 m above the surface and pebbles may be blown over ice (McKenna Neuman 1990).

Niveo-aeolian deposits have poor to moderate sorting and a wide range of modal grain sizes, but most are dominated by medium and coarse sand (0.2–2.0 mm). Fresh deposits often reveal concentrations of sediment at the top and base of the snowpack, layers of mixed snow and sediment, and sediment-rich layers separated by clean snow. During melt, sediment becomes concentrated at the snow surface. Such supranival deposits tend to be thickest on the lee of obstacles, notably on the slip faces of dunes. Melt of the underlying snow produces a range of denivation features including dimpled surfaces, snow hummocks, contorted bedding, sinkholes, cavities, tension cracks and faulted slipface strata (Koster and Dijkmans 1988; Dijkmans 1990). Meltwater fans may accumulate at the lower edge of niveo-aeolian beds (Lewkowicz and Young 1991).

Unless rapidly buried under prograding slipface beds, most denivation structures dry out and are destroyed by summer winds. Niveo-aeolian deposition therefore rarely leaves any distinctive sedimentological or structural signature in cold-climate aeolian sequences. For this reason the role of niveo-aeolian activity in the formation of the extensive Late Pleistocene COVERSANDS and dunefields (see SAND SEA AND DUNEFIELD) of Europe and North America remains contentious.

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SEE ALSO: aeolian processes; nivation; periglacial geomorphology

COLIN K. BALLANTYNE

NON-LINEAR DYNAMICS

Geomorphologists have long recognized that landforms are the result of a complex set of interactions operating over different scales of space and time (Schumm and Lichty 1965; Schumm 1979; Brunsden and Thornes 1979). More recently, these ideas have been variously strengthened, confronted and extended by incorporating findings and analytical tools from the subject of non-linear system dynamics (NSD) developed in the mathematical and physical sciences. The term 'non-linear' expresses an unequal relationship between the driving force or the stress and the geomorphic response, most simply described as where *outputs are disproportionate to inputs*. A good example is the classic Hjulström curve of velocity of water plotted against the size of entrained sediment particles. It is a non-linear curve showing that the output (sediment size entrained) is not proportional to the input (water velocity); factors such as particle density and inter-particle cohesion are also important. The geomorphology of a landscape is governed by the interaction of a vast array of such processes operating in different parts of a landscape and over different timescales. Hence, the term 'non-linear dynamics' is used to describe the behaviour of the system rather than the behaviour of discrete process interactions. In the example of an unstable slope system, this means that the relationship between the intensity of precipitation falling on a slope and the size and timing of landslides may be complex and non-linear, rather than a simple cause and effect. It is widely believed that all complex systems consisting of interacting components behave non-linearly. The attractions of NSD lie with the possibilities of finding generic insights about geomorphic system behaviour and mapping well-studied model behaviours onto real systems that would normally be non-observable by conventional field methods. In practice, the demonstrable existence of model phenomena in real systems and hence the usefulness of NSD ideas are contested.

NSD may provide useful insights about (1) the predictability and unpredictability of geomorphic phenomena; (2) distinguishing between spontaneous and forced geomorphic change; (3) the sensitivity and resilience of landscapes to impact; and (4) the use of appropriate conceptual and modelling scales and methodologies.

A useful division in the discussion of non-linear system behaviour is to identify intrinsic or extrinsic changes. Intrinsic changes are those that spontaneously occur through self-organization as part of the system's own dynamic without any direct and proportional external forcing, analogous to the idea of biological evolution. One explanation is provided by the so-called 'arrow of time' implicit in the second law of thermodynamics, which predicts that a system will develop towards an equilibrium point where the free energy is minimized and thermodynamic entropy (disorder) is maximized. Thus, free energy in the form of flowing water drives the organization of a river network, especially its drainage density, so that the total potential of the flowing water to erode and carry sediment is minimized (or diffused) across the landscape. As in the classic Davisian cycle of landform evolution, thermodynamic equilibrium is reached when the relief is progressively lowered to a peneplain. However, since the Earth's surface is not dominated by peneplains but rather by an array of other geomorphic forms, it is clear that most landscapes have had their 'arrow of time' development arrested. Therefore, embedded within the idea of 'arrow of time' are other system behaviours that explain geomorphic systems set at some point far from equilibrium, known as dissipative systems (Prigogine 1996).

One of these, emergent complexity, describes the conditions found in open, and often partitioned systems that show ordered states maintained by flows of energy across the system boundary. Within human timescales, they may appear unchanging in their underlying forms but over longer timescales these systems evolve or emerge to become increasingly ordered and complex, as in the example of progressive weathering leading to soil horizons supporting a complex terrestrial ecosystem. Another, chaotic behaviour, may explain the local variability of many geomorphic systems. Chaos is most easily seen in mathematical equations that describe processes such as turbulent water flow, but the implications of chaos theory for geomorphology are large. Chaotic behaviour means that the exact pathway of a set of interacting processes over time is crucially sensitive to initial conditions and to even the smallest external perturbations. Over a long period of time, an observer of a chaotic landscape would see small initial differences in relief, drainage and soils amplified over time: divergence, rather than convergence. One result might be a mosaic of soil types overlying a fairly uniform parent material that initially differed from place to place only in small differences in texture. The variability would be amplified by subsequent vegetation succession and positive feedback controls. In one sense this means that a chaotic system, as we know for weather forecasts, is an unstable system becoming progressively unpredictable as the system moves from its starting point. However, model chaotic systems also evolve to lie within well-defined ranges, known as attractors, which translates to the natural environment in terms of the degree of variability at a higher scale being constrained and therefore predictable in probabilistic terms. Thus the overall range of soil types encountered under the emergent complexity displayed in a temperate woodland is fairly easy to predict from knowledge of the tree species, climate and geology but the local scale variability of soil properties, driven by chaos, may appear almost random.

Embedded system behaviours may also explain the remarkable fact that many emergent geomorphic patterns are the same, irrespective of the spatial scale. Aerial photographs of ripple beds and dunefields may be indistinguishable without an absolute distance scale. Stream networks, as measured by a stream-ordering parameter, such as Horton's bifurcation ratio, are often statistically similar whether viewed at the scale of the whole drainage basin or a small sub-basin. Many other geomorphic features show this so-called scale-invariance or fractal geometry (Figure 110a), which is easily identified by a power law (straight-line) relationship in a log–log plot of the variable and spatial scale. The behaviour of scale-invariant phenomena has been studied under the heading of self-organized criticality (SOC). This term was used to explain the response of a model sand pile to continuous additions of sand from above. The pile maintains a characteristic form (at the macrolevel) through losses of sand by avalanches (negative feedback at the microlevel), whose distribution over time (either size or frequency) conforms to a power law. The power law distribution means that it is possible to predict the overall

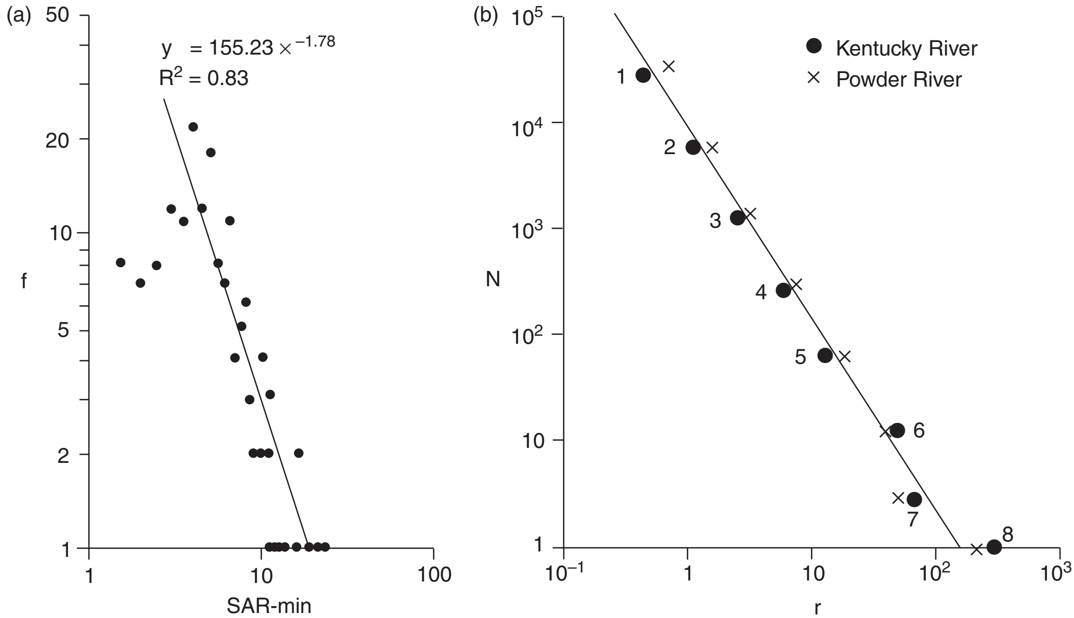

Figure 110 Scale-invariant phenomena in geomorphic patterns and processes. (a) Dependence of the number of streams N of various orders 1–8 on their mean length r for the Kentucky River basin in Kentucky and the Powder River basin in Wyoming (Turcotte 1997); (b) inverse power law relationship between the minerogenic sediment accumulation rate SAR-min and its frequency f for a mid-Holocene lake sediment record of catchment erosion at Holzmaar, Germany, indicative of self-organized criticality (Dearing and Zolitschka 1999)

properties of the system, such that there will be fewer large avalanches than small ones, but the details of timing and size of the next avalanche are unpredictable. A theory has developed that argues for systems of all kinds to evolve to critical states where they may similarly respond to small perturbations in a disproportionate manner, but importantly will evolve back to the original state (Bak 1996). Evidence for SOC now exists for many real spatial patterns, such as river networks and forest fires, and also real time series (Figure 110b), such as earthquakes, landslides and river sediment transport (e.g. Dearing and Zolitschka 1999). Thus, many geomorphic landscapes and landforms may be viewed at one scale as complex and ordered non-linear systems, emerging from the evolution of chaotic processes at another. Chaotic processes may produce highly variable or apparently random forms at a small scale, but through constraints imposed by the nature of the environment (e.g. geology, climate) may produce identifiable and self-similar patterns at large spatial scales, and these may exist in critical states.

Geomorphic systems also respond to external forces, like climate and human activities. The nature of the response depends on the force and the condition of the system and these may vary from direct and reversible to time-lagged and irreversible (Figure 111). Knowing the NSD dynamics that underlie a system may help to define the resilience or sensitivity of a geomorphic system. For example, how close a non-linear landscape system lies to a bifurcation point, as defined by the mathematical representation of that system, helps define its sensitivity to external impacts: the closer it is, the more likely is the system to be driven along a new irreversible trajectory towards an alternative steady state (Thornes 1983).

The implications of NSD in geomorphology are profound. We may have to accept that some geomorphic outcomes are unpredictable and strongly contingent on a landscape's history. It may be that reductionism as a methodology is unlikely to afford extrapolation of mathematical rules to large scales (Lane and Richards 1997). There may be a strong case for adopting some form of 'scientific

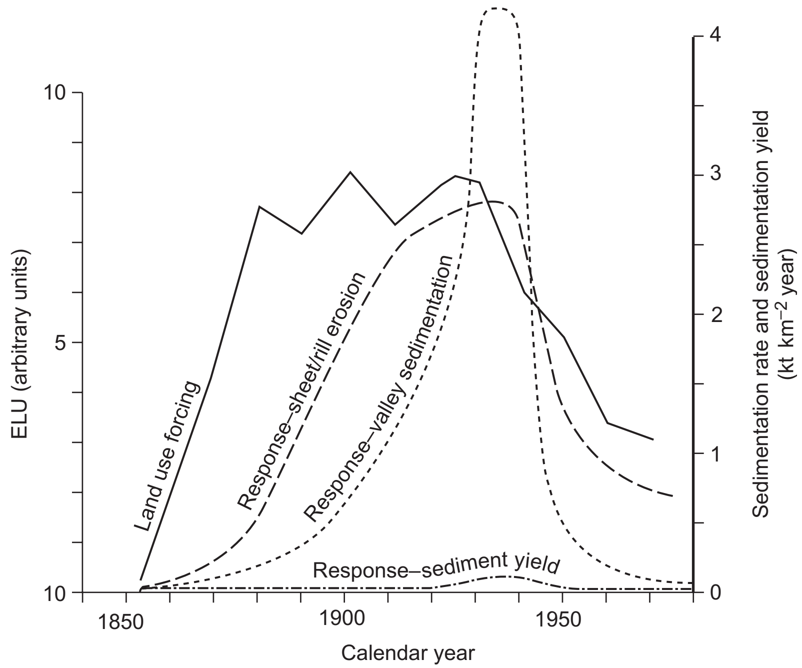

Figure 111 Non-linear relationships between forcings and geomorphic responses in space and time. Erosive land use change ELU in the Coon Creek basin, Wyoming, forces lagged temporal responses in hillslope sheet and rill erosion, main valley sedimentation, and catchment sediment yield (Trimble and Lund 1982; Wasson and Sidorchuk 2000)

realism' as the correct methodology, where we accept the existence of non-observable phenomena, structured and stratified systems with emergent properties, contingent relationships and prediction based on probabilities, while rejecting a belief in direct and enduring relationships between cause and effect (Richards 1990). Paradoxically, overcoming the difficulties of mapping NSD concepts and ideas onto real landscapes may be achieved through the use of simple computational models with simple rules. This 'new kind of science' (Wolfram 2002) uses cellular automata, where grids of interacting cells each containing a set of simple rules are updated at subsequent time steps in order to simulate the evolution of the system. Landscape models based on cellular grids and simple equations for sediment and water flows (e.g. Coulthard *et al*. 2000) show much promise in being able to simulate the spatial and temporal development of complex geomorphic forms with realistic non-linear behaviour. These may prove to be the most efficient means for simulating how non-linear landscapes will really respond to future combinations of climate and human activities.

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JOHN DEARING

NOTCH, COASTAL

Notches can be cut at the cliff foot by waves, but they are generally poorly defined in fairly homogeneous rocks, and locally restricted to geologically favourable locations in more variable rocks. Notches, typically between 1 and 5 m in depth, are most common and best developed on tropical limestone coasts, where low tidal range concentrates the erosional processes. The formation of notches throughout the tropics is generally attributed to chemical or biochemical CORROSION, or to biological grazing and boring (see BORING ORGANISM), especially in sheltered locations. Nevertheless, ABRASION and other forms of mechanical wave erosion contribute to their formation in some areas. As there is little agreement over the level, or levels, that tropical notches are developing today, the occurrence of double or multiple notches in some places has been ascribed to changes in relative sea level, intermittent tectonic activity, variable rock structure and lithology, and the effect of organisms and other notch-forming mechanisms operating most efficiently at different elevations.

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ALAN TRENHAILE

NUÉE ARDENTE

A French term most frequently translated into English as a 'glowing cloud', to refer to a pyroclastic flow (see PYROCLASTIC FLOW DEPOSIT) or surge from a volcano. First used by Lacroix (1904) to describe the pyroclastic clouds erupted from Mt Pelée, Martinique, on 8 May 1902 and subsequently. They destroyed the town of St Pierre killing most of its 27–28,000 inhabitants in the worst volcanic disaster of the twentieth century, as measured by loss of life. The term 'ardent' was originally used strictly to refer to 'very hot, burning or scorching' rather than 'glowing', and specified that such clouds were not incandescent at night, except close to the crater (Tanguy 1994).

Nuée ardente has come to be used as a general term for a hot, turbulent, self-expanding gaseous cloud with its entrained particles of rocks and ash, which may be incandescent, that descends the flank of a volcano at high or exceptionally high velocities. In the high velocity flows the denser, lower part hugs the ground surface and becomes strongly controlled by the pre-existing topography. This portion usually forms the bulk of a resultant deposit. Such deposits show massive coarse bouldery facies in channels or the axis of flow, and usually grade to sandy deposits both laterally and upwards. There is evidence of searing heat for only a few brief minutes, but if objects are buried in the deposit they show signs of being cooked. The lighter, upper part of the flow, comprising hot gases and ash particles, rapidly expands upwards as a dark, towering cloud to many kilometres in height. This cloud may spread over a much wider area than the dense lower flow, resulting in a widespread coeval ash.

In exceptionally high velocity surges, a blast of gases and entrained particles travels outward from the volcano, usually independent of the topography, removing most structures and trees in its path. Vertical surfaces become 'sandblasted', the bark of trees may be stripped, trees may be snapped and removed laterally for large distances, rocks may become embedded within materials they impact against, buildings may be razed with their contents twisted or carried from the scene. Resultant deposits show a high degree of fluidity (a low concentration fluid) of the gaseous mixture with cross-stratification, antidunes, highly irregular erosion breaks, considerable lateral variability in lithology and thickness, and frequently charcoal.

Nuée ardente has been used in a broad sense to encompass all pyroclastic flows and surges and in a restricted sense to refer specifically to small volume monolithologic block-and-ash flows generated by the collapse of actively growing lava domes or lava flows on steep terrain (Pelean eruptions).

Nuées ardentes are one of the most feared volcanic hazards. It is not simply the destructive energy of the cloud but the searing heat of gases and ash particles that make them so lethal to life and property.

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VINCENT E. NEALL

NUNATAK

Nunatak is an Inuit word referring to a mountain top that protrudes above the surface of a GLACIER or ice sheet. Such summits are subject to intense frost weathering but escape glacial erosion. On relatively flat surfaces this often results in the formation of autochthonous blockfields (see BLOCKFIELD AND BLOCKSTREAM), whilst sharper peaks and ridges can become broken along joints to produce sharp ARÊTEs and pinnacles. In areas that were glaciated in the past, the difference in weathering and erosion can be identified as a 'periglacial trimline' that separates glacially eroded terrain from higher ground that retains evidence of prolonged weathering.

A range of simple measures have been used to quantify the difference in degree of rock surface weathering above and below proposed trimlines, including rock surface roughness, joint depth and surface hardness as recorded using a SCHMIDT HAMMER (McCarroll *et al*. 1995). One of the most reliable indicators of past nunataks is the presence of the clay mineral gibbsite at the base of the soils. Gibbsite, an aluminium oxide, is an end product of the weathering of silicate minerals and in mountainous environments in the extra-tropics is thought to represent a long period of *in situ* weathering. Cosmogenic isotope exposure age dating (see COSMOGENIC DATING) can also be used to test the hypothesis that summits escaped glacial erosion and to date the exposure of glaciated bedrock and ERRATICs (Stone *et al*. 1998).

The identification of 'palaeonunataks' using geomorphological and dating evidence provides field evidence with which to test and improve models of past ice sheets in formerly glaciated areas such as the British Isles, northern Europe and North America (Ballantyne *et al*. 1998; McCarroll and Ballantyne 2000). It has also been argued that nunataks in these areas may have provided refugia for plant species to survive glaciations and then re-populate when the glaciers receded. In some cases this would help to explain the rather patchy occurrence of some species, though conditions would have been extremely harsh (Birks 1993).

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DANNY McCARROLL

O

ORGANIC WEATHERING

Organisms and organic compounds play a wide range of roles in both enhancing and retarding rock and mineral weathering processes in most environments. Indeed, as Reiche (1950: 5) recognized, weathering involves 'the response of materials which were at equilibrium within the lithosphere to conditions at or near its contact with the atmosphere, the hydrosphere, and perhaps still more importantly, the biosphere'. Thus, by definition, weathering in most places operates in a non-sterile environment and biological processes and influences need to be taken into account if we are to understand weathering fully. The term organic weathering is not widely used by geomorphologists, with biological weathering more commonly found in the literature, but both have a similar broad definition as the suite of biological weathering processes and indirect biological influences on weathering. Bioerosion is an allied term, referring to the erosive activity of organisms especially on bare rock surfaces, and there is a spectrum of such organic influences on weathering and erosion. Many books refer to a three-fold classification of weathering into physical, chemical and biological processes, but it is perhaps simpler to keep the classic division into physical and chemical weathering process groups, acknowledging that many processes within these groups can be seen to be biophysical or biochemical in nature.

Although there was a flourishing of interest in biological weathering in the late twentieth century, as analytical and microscopical techniques became available which permitted closer study of the rock:organism interface, interest in the possible roles of organisms and organic compounds in weathering is nothing new. Several late nineteenth-century workers carried out pioneering experiments on the role of plant seedlings (Sachs 1865), lichens (Sollas 1880) and organic acids (Julien 1879) in weathering common rocks and rock-forming minerals. However, there was little general assessment of the overall nature, rate and importance of biological weathering.

Research into organic weathering has tended to focus on a number of key types of organism or organic influence. First, there has been a huge concentration of effort into understanding the role played by organic acids and other organic compounds within soils on the weathering of minerals (Drever and Vance 1994). Second, there have been a number of studies on the role of plant roots in weathering both rocks and minerals, illustrating the important uptake of many elements by such processes (Kelly *et al*. 1998; Hinsinger *et al*. 2001). Third, a few studies have been made of the role of animals in weathering, especially their contribution to the decomposition of organic material leading to the production of organic acids. Fourth, there have been many studies on the role of lichens in the weathering of rock surfaces. These studies illustrate the complexity of roles played by organisms in weathering, with biophysical and biochemical lichen activity recognized, as well as a bioprotective role in some cases where they retard the action of other weathering processes and bind the rock surface together. Different lichen species play different roles, with some species being highly biodeteriorative and others not. For example, a recent study by Robinson and Williams (2000) in the Moroccan High Atlas show accelerated weathering of sandstone by *Aspicilia caesiocinerea* agg. producing notable scars on the rock surface. Finally, there have been many studies of the roles played by micro-organisms (notably algae, fungi and bacteria) in weathering. In most environments, combinations of different processes and organisms are involved producing a complex biological imprint on weathering. Thus, for example, the large percentage of the terrestrial landsurface covered by soils will experience weathering conditioned by soil organic acids, plant roots, animal decomposition and the many micro-organisms that inhabit soils (e.g. symbiotic mycorrhizal fungi have been shown by Jongmans *et al.* 1997 to bore into feldspars in soils, thereby allowing uptake of Ca and Mg by plant roots). Similarly, most bare rock surfaces are not actually bare at all but are coated with a diverse community of micro-organisms and lower plants which play a range of roles in weathering.

Biophysical weathering often involves the creation of stresses by the expansion and contraction of organisms or parts of organisms. Lichens, for example, can absorb a vast amount of water leading to huge expansion on wetting and a concomitant shrinkage on drying. Crustose lichens, which grow very closely attached to rock surfaces, can cause much weathering to the underlying rock. Fry (1927) and Moses and Smith (1993) provide experimental evidence of the physical weathering caused by such wetting and drying on a range of rock types. Natural processes of growth and decay of rock-dwelling organisms or parts of organisms can also cause biophysical weathering. Some crustose lichen thalli, for example, start to peel away from the rock surface at the centre as they senesce, producing patches of intensive weathering. Plant roots may also force their way into cracks and joints, with large pressures exerted during growth sufficient to induce weathering. Grazing and mechanical burrowing of animals can also produce physical weathering, although this is usually categorized as a form of bioerosion.

Biochemical weathering includes a host of individual processes, with the production of carbon dioxide, organic acids and chelation (often involving organic acids) being particularly important. Within soils, organic by-products can provide a key control on weathering processes, especially within the rhizosphere. Many organic acids, such as humic, fulvic, citric, malic and gluconic, have been shown to be capable of weathering a range of substrates. Several types of micro-organism have been found to be capable of boring into suitable substrates (such as rock, corals and mineral grains) through chemical means. The exact mechanism by which different micro-organisms bore has been much debated, but extracellular acids and other substances probably promote chemical decomposition of minerals. The production of a network of near-surface boreholes can encourage further weathering by other (often inorganic) processes as they weaken the surface and increase the reactive surface area. Biochemical weathering by lichens involves the production of carbon dioxide, oxalic acid and the complexing action of a range of sparingly soluble lichen substances. Lichens such as the highly biodeteriorating *Dirina massiliensis* forma *sorediata* produce oxalic acid which reacts with calcium carbonate within rocks to produce calcium oxalate (Seaward 1997).

Another important aspect of organic influences on weathering is bioprotection. In this case, organisms do not play an active role in encouraging weathering, but rather play a passive role. The very presence of a cover of lichens or biofilm, for example, protects the underlying rock surface from extremes of temperature (thus reducing the potential for damaging freeze–thaw weathering for example). Furthermore, the lichen or biofilm can also act as a sponge soaking up incident rainwater and providing a chemical buffer for the underlying surface, thus reducing the potential for chemical weathering.

Identification of the various organic weathering processes and influences is a challenge; an even greater challenge revolves around trying to quantify their effect and identify their contribution to overall weathering rates. Some measurements of biological weathering rates are presented in Goudie (1995) derived from detailed empirical studies in the field as well as experimental studies. Many biological processes have proved very difficult to measure in the field with currently available techniques (e.g. physical weathering by plant roots) and experimentation is also challenging where growing organisms are involved. Although some organic weathering processes operate at a fast rate and are quite dramatic, their action is often highly localized in time and space. Thus, for example, biodeteriorating lichens can act at a very fast rate, but they are often patchily distributed over a rock surface, so their net effect may be reduced. Furthermore, over time community dynamics will influence the species present on any surface, limiting the net contribution of biodeteriorating species.

Despite the wide range of studies on a variety of organic weathering processes and influences, there is still uncertainty about their general importance and contribution to weathering in different environments. Viles (1995) hypothesized that on bare rock surfaces biological weathering would increase in 'hostile' environments, and that less bioprotection would take place. The reasoning behind this assertion is that in hostile environments organisms extract more nutrients, water and shelter from the rock itself, producing a weathering effect as they do so. In more benign environments organisms tend to have a less close contact with the surface, and their net role will be a protective one. Thus, hot desert, cold tundra and coastal environments should be characterized by high rates of biological weathering, in comparison with humid temperate and tropical ones. However, many harsh environments are also characterized by slower biological growth rates which may complicate this pattern.

Even if biological weathering on bare rock surfaces can be seen to be more intense in some environments rather than others, this does not necessarily imply that it will be the dominant series of processes there. In hot arid areas, for example, although lichen weathering can be spectacular, heating and cooling and salt weathering can also be highly effective and operate at a much faster rate than can slow-growing lichens. Considering the soil-covered landscape it is probable that the reverse applies, with higher rainfall and temperatures encouraging organic growths within the whole ecosystem, thus enhancing the production of organic acids and increasing both the overall rate of weathering and the contribution of organisms to weathering.

In some instances, recognizable small-scale landforms can be produced by biological weathering processes. For example, lichen fruiting bodies create small pinhead-sized pits in rock surfaces and various authors have identified larger pitting and grooving as being organically produced (e.g. Danin and Garty 1983; Robinson and Williams 2000). On limestone surfaces such features may be called BIOKARST. It has proved difficult to establish convincing process-form links in many cases, as there are a multiplicity of ways in which similar landforms at this scale can be produced.

Future challenges for geomorphologists in understanding organic weathering include the need to provide quantitative comparisons of sterile vs. organically mediated weathering rates in different environments, and the need to provide a broader assessment of the overall role of organisms and organic by-products in weathering and sedimentation. Finally, scientists such as Robert Berner have suggested that biotically enhanced weathering has played a major role in the global carbon cycle over long time spans (Berner 1992), and geomorphologists can play a key role in providing reliable empirical data on biological weathering rates and their spatial and temporal variability in order to test such models.

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HEATHER A. VILES

ORIENTED LAKE

Oriented lakes are subparallel, elongated lakes, which commonly occur in extensive clusters. Many of these clusters, especially in bedrock, are the result of processes antecedent to lake formation. For instance, orientation on the Canadian Shield is commonly associated with glacial fluting. However, there are belts of oriented lakes covering thousands of km^2 in unglaciated areas of the sandy Arctic coastal plains of Russia, northern Alaska and north-west Canada. They also occur in non-permafrost areas, such as the Atlantic coastlands of Maryland, the Carolinas and Georgia, USA. Two forms of oriented lake have been recognized: the most common are elliptical, but rectangular shapes occur in the Beni Basin of northeastern Bolivia, and the Old Crow and Bluefish Basins of northern Yukon Territory, Canada.

In North America, the elliptical lakes occur outside the glacial limits, at sites where there is no evidence of glacial deposition. Some occur in postglacial marine terraces. Well-known examples are the Carolina Bays of the southern Atlantic coast, and the lakes near Liverpool Bay, NWT, and Point Barrow, Alaska, on the western Arctic coast. In permafrost environments, the lakes are in depressions formed by thermokarst processes, which have been elongated during their development. The lakes range in size over several orders of magnitude, from ponds with long axes of less than 30 m, to water bodies of over 1,500 ha. The lake-size distributions are skewed, with the mean being less than 250 ha. Along the western Arctic coast, the mean length to width ratio of the lakes is about 1:7. The major axes of the lakes are aligned perpendicular to the prevailing winds, and, in the case of the lakes near Liverpool Bay, the standard error of mean orientation is less than 3°.

Several theories have been advanced for the causes of lake orientation. Many of these, including bombardment by meteorite showers, effects from upwelling of artesian springs or the action of fishes hovering while spawning, have been discredited. However, consideration of the effects of wind action perpendicular to the long axes of the lakes has been supported by field data and laboratory experiment.

The hydrodynamic theory proposes that winds blowing across a lake establish a two-cell current circulation within the water body, with water returning to the windward shore around the ends of the lake (Rex 1961). The maximum littoral drift, and associated erosion, occurs at the ends of the lakes, where the angle between waves propagating in deep water and a line perpendicular to the shore is 50°. A similar circulation has been measured in large lakes of the Alaskan coastal plain (Carson and Hussey 1962), and has been reproduced at laboratory scale by blowing air across a 2 m square box containing a scale model pond. The model pond, initially circular, became elongated in the process. The equilibrium form of a finite shoreline in readily transported, unconsolidated material is cycloidal, which corresponds to the approximate shape of the oriented lakes along the western Arctic coast. The orientation of the lakes in the western Arctic is consistent with the hydrodynamic theory, when applied using the prevailing wind regime of the region (Carson and Hussey 1962; Mackay 1963).

The rectangular form of the oriented lakes in the Old Crow and Beni basins has been associated with aspects of bedrock structure propagating through the overlying sediments (Allenby 1989), but the explanation has not been verified. In northern Yukon, the lakes are THERMOKARST features that have developed in sediments up to at least 150 m thick, in which permafrost is present in the upper 60 m. The permafrost impounds the lakes. In Bolivia, the sediments are similarly thick, but lake levels are maintained by a high water table close to the surface.

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C.R. BURN

OROGENESIS

Orogenesis is the building of mountains by the forces of PLATE TECTONICS. Driven by the internal heat of the Earth, motions of lithospheric plates produce changes in crustal thickness structure that result in vertical motion of the topographic surface. It is this motion that is responsible for creating impressive mountain landscapes that have been the inspiration of so much geomorphology.

Mountains are built slowly over geologic time as an accumulation of CRUSTAL DEFORMATION. Rates of mountain building are quantified as the rates of vertical motion of rock with respect to the geoid (rock uplift), the surface with respect to the geoid (surface uplift), or rock with respect to the surface (exhumation, see also DENUDATION) (England and Molnar 1990). The relationship, rock uplift = surface uplift + exhumation, is widely adopted. Surface uplift describes topographic growth and creation of the positive landforms of mountains. Long-term surface uplift is difficult to measure, but is approximated using the altitude dependence of fossil organisms or displacement of features with respect to eustatic sea level (Abbott *et al.* 1997). Short-term surface uplift can be measured using geodetic techniques (e.g. GPS and synthetic radar interferometry). Relative surface uplift can be constrained using geomorphic markers, such as river terraces (see TERRACE, RIVER) or erosion surfaces. Exhumation relates to erosion, which is critical for accommodating plate motion by transferring mass from thickened mountain belts to adjacent basins. It is generally inferred from rock cooling histories (see DENUDATION CHRONOLOGY and FISSION TRACK ANALYSIS), basin sedimentation records or COSMOGENIC DATING.

Active orogens may form at rates of rock uplift of 0.01–10 mm yr^{-1}, and a rate of ~1 mm yr^{-1} is representative of many actively growing mountains around the world. Mountain building tends to be stable over millions to tens of millions of years, such that major episodes of orogenesis are commonly given formal names (e.g. the Alpine Orogeny). The timescale of mountain building is short enough, however, that changes in plate motion and global climate throughout the latest Tertiary and Quaternary have had noticeable effects on most orogens.

Orogenic belts are most common along ACTIVE MARGINs, such as the arcuate mountain belts of the 'Ring of Fire' along the continental rim that surrounds the Pacific plate. The characteristics of these mountain ranges depend on the type of plate boundary. The majority of mountain uplift is produced by convergent tectonic motion, where two or more plates collide and increase crustal thickness. One setting in which this occurs is Cordilleran-type orogens, well represented by the Cascades of northwestern North America or the Andes. These mountains stretch above subduction zones and are produced by permanent convergent deformation of the overriding continental plate and thermally driven buoyancy of a magmatic arc. Their anatomy includes coastal ranges near the accretionary complex of the subduction zone, lines of volcanoes that are separated from the coast by elongate valleys, and highlands that rise above foreland fold and thrust belts that verge towards the continent interior. Width of these highlands is largely dependent on the dip of the subducting oceanic lithosphere.

A second setting of convergent mountain building is continental collision, typified by the active collision of India and Asia. India has underthrust beneath Asia over the past 50 Ma, resulting in uplift of the High Himalaya, Tibetan Plateau and associated mountains that penetrate thousands of kilometres into the interior of Asia (Hodges 2000). Several thousand kilometres of plate convergence has been accommodated by a combination of orogenic crustal thickening and lateral escape of microplates via strike-slip faults. That continental collision is the most effective mechanism of mountain building is evident. Half of all mountain peaks worldwide that rise above 7.5 km elevation occur in the Himalaya, while all of the remaining half are associated directly with the India–Asia collision. The crystalline core of the Himalaya has also experienced ~10–20 km of DENUDATION in the Neogene at rates locally as high as 1 cm yr^{-1}, leading to rapid deposition in the Bengal and Indus fans (Searle 1996). In Tibet, arid conditions and internal drainage basin geometry have hindered erosional exhumation, leading to formation of an orogenic plateau above the under-thrust Indian plate. Geodynamic processes limit the plateau's elevation to ~5 km, via lower crustal flow where strength is exceeded by gravitational load (Royden *et al.* 1997).

Mountains are also built at divergent and transform plate boundaries and within continental interiors. Crustal extension via normal faulting leads to tilting and uplift of large hanging-wall blocks and results in characteristic basin and range topography (e.g. the western USA). Strike-slip plate boundaries may produce narrow zones of orogenesis where plate motions are somewhat oblique in a convergent (transpressional) or divergent (transtensional) sense. The components of non-strike-slip motion along such boundaries may be accommodated by reverse or normal faulting, so that similar but laterally confined mountain systems result (e.g. the Transverse Ranges along the San Andreas fault). Mountains within continental interiors may represent earlier stages of continental deformation or the effect of geodynamic processes. Dynamic topography on cratons can occur where compositionally or thermally induced density contrasts occur in the sub-lithospheric mantle.

Although a mountain may owe its origin to tectonic construction, mountain landscapes are dominated by erosional processes. Mountain topography consists of valleys and hillslopes shaped by erosional agents (e.g. mass wasting, glacial erosion, fluvial erosion). Agents of erosion are poised to reduce the terrestrial surface to low RELIEF, such that tectonic orogenesis is critical for maintaining topographic variation above the mean elevation of the continents. The act of mountain building hence has important effects on the dynamic processes of erosion itself, both directly (e.g. changes in BASE LEVEL) and indirectly (e.g. the effect of mountains on local climate). Because of this erosional character of mountains, topographic character does not always discriminate orogens that are actively forming from those formed by prior plate motion.

Mountainous topography may linger for more than one hundred million years after cessation of active tectonic deformation (e.g. the Appalachian Mountains). Rock uplift and erosion continue as long as an orogen contains a thickened crustal root, that must be removed by DENUDATION. Topography in ancient mountains is maintained, even after erosion has removed many kilometres of rock, because of ISOSTASY. The ductile nature of the upper mantle permits adjustment to gravitational loads over timescales of ~10^3 yr. The increased thickness of crust beneath orogens is more buoyant than the mantle rocks that lie beneath adjacent crust, such that the topographic surface is higher than the surrounding region. As erosion removes mass from the mountains, the crust rebounds to remain in gravitational equilibrium. The magnitude of rebound is proportional to the ratio of crust and mantle densities, such that mean elevation decreases much more slowly than the rate of regional denudation. Isostatic rebound can even produce the uplift of peaks where mean elevation is decreasing, where valleys are incised faster than interfluves erode.

Although inactive mountain ranges are still referred to as 'orogens', they experienced their tectonic orogenesis in the geologic past and are thus distinct from mountains presently rising along active plate margins. Erosion rates of ancient mountain belts may become quite slow, such as average rates in the Appalachian Mountains of 0.02–0.04 mm yr^{-1} (Mills *et al.* 1987). Denudation this slow is essentially weathering-rate limited, yet it is enough to reduce crustal thickness over long periods. Variations in the geomorphic system, such as climate change, may impose upon the stagnant erosional setting of ancient mountains and force readjustment of erosional processes. This often leads to incision and REJUVENATION of topography.

The character of topography itself has traditionally been used to interpret the surface uplift and exhumation history of mountain belts. This is one goal of landscape evolution studies, which seek to define cyclic changes in landforms or topographic 'maturity' through stages of orogenic and post-orogenic development (see CYCLE OF EROSION). However, this approach requires tenuous assumptions about the relationships between topographic parameters and uplift and exhumation. Studies have shown that many aspects of topography, such as RUGGEDNESS, DRAINAGE DENSITY and hypsometry (see HYPSOMETRIC ANALYSIS), are dependent on the erosional resistance of rocks, the nature of local climate and the individual processes of the dominant erosional agent (e.g. Willgoose and Hancock 1998), such that direct interpretation of orogenic history from topographic parameterization is dubious. One parameter that has been demonstrated to correlate with exhumation rate is slope or short-wavelength relief. Slope is almost synonymous with erosion rate, as rates of all erosional processes increase with slope and slope must increase in areas of rapid surface uplift due to ever-growing gravitational instability. Nonetheless, the clues to orogenic evolution do not lie entirely in topographic parameterization using DIGITAL ELEVATION MODELS.

Because the geomorphic and geodynamic processes that shape orogenic belts are complex and occur over very long scales of time and space, the topographic evolution of mountain landscapes is difficult to study comprehensively with physical experimentation or direct observation. For this reason, the approach of numerical modelling has become important (see MODELS). Models can quantitatively test how elements of landscapes should evolve under a given set of conditions, based on the use of erosion laws derived from previous research. Models can be constructed as grids of cells with set boundary conditions, such as the distribution of tectonic uplift or base-level change, frequency and magnitude of precipitation events, and rheology of eroding materials. Erosion laws, such as stream power bedrock incision or diffusive hillslope transport, can be run under different conditions. The result is a representation of the long-term landscape evolution of a hypothetical setting, and can be instructive with respect to the role of boundary conditions or specific processes (Burbank and Anderson 2001). However, these models are limited by present degree of understanding of individual erosional processes and the difficulty of capturing real-world complexity.

The significance of erosion goes beyond shaping the landscape of orogens. Erosion can itself influence tectonic processes. For example, erosion modulates surface slope in deforming thrust wedges that can in turn affect deformation (e.g. critical taper wedges; Dahlen *et al.* 1984). Concentrated erosion in parts of an orogen also control trajectories of crustal motion, and thus influence deformation partitioning (e.g. the Southern Alps, New Zealand; Koons 1989). This in turn causes faster tectonic uplift and increases gravitational potential energy, leading to more rapid erosion (i.e. positive feedback loop). Eventually, erosion and tectonic rock uplift rates may become balanced in a steady state of mass flux. In the north-west Himalaya, for example, a steady state has been achieved where topography is sufficiently rugged as to result in bedrock landsliding that keeps pace with base-level induced incision driven by tectonic uplift (Burbank *et al.* 1996).

The most important external influence on erosion landscape evolution and mountain building may be climate. Rates of erosion tend to correlate with precipitation. Spatial concentrations of precipitation due to the orographic effect of topography can lead to concentrated erosion (e.g. Irian Jaya fold belt; Weiland and Cloos 1996). This is a core element of models of coupled erosion and tectonic evolution of orogens (Willett 1999). Latitudinal, altitudinal and temporal variations in climate also determine where glacial and fluvial erosion dominate. Glacial erosion is thought to be more effective, and can force relief production via valley incision and intense erosion at equilibrium line (see EQUILIBRIUM LINE OF GLACIERS) altitudes that can 'buzz-saw' mountain landscapes (Brozovic *et al.* 1997). The effects of climate may be so important that increases in erosion and sediment production in many mountain systems worldwide may have been caused by the onset of glacial climate ~4 Myr ago (Molnar and England 1990).

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JAMES A. SPOTILA

OUTBURST FLOOD

Outburst floods or jökulhlaups, are high magnitude, low frequency catastrophic (see CATASTROPHISM) events involving the sudden release of glacial meltwater (see MELTWATER AND MELTWATER CHANNEL) stored in subglacial (see SUBGLACIAL GEOMORPHOLOGY) reservoirs or ice-dammed lakes. The volume of water discharged is usually orders of magnitude greater than normal flow, with modern outburst floods estimated up to 2,000,000 $m^3 s^{-1}$ and Pleistocene-aged flood peaks estimated at 21,000,000 $m^3 s^{-1}$. However, most discharges usually measure hundreds to thousands $m^3 s^{-1}$. Importantly, outbursts leave characteristic erosional and sedimentary signatures, which enables palaeo-outburst flood reconstructions.

Hydrographs with a slowly rising limb followed by a rapidly falling limb characterize outburst floods through glaciers. The steadily increasing discharge reflects a greater efficiency in subglacial water routing, aided by a positive feedback interaction of channel enlargement caused by melting and abrasion. The increasing discharge may overwhelm the subglacial drainage network, and it is not uncommon for water to burst from SUPRAGLACIAL positions at the ice margin. The rapid decrease in discharge is a reflection of either a drained reservoir, or rapid tunnel closure caused by ice deformation or collapse.

The water source is variable, dependent upon the location of the glacier and the reservoir. In volcanic regions, such as Iceland, high geothermal gradients and subglacial volcanic eruptions lead to the rapid production of meltwater and cyclic outburst floods. For example, Grimsvötn, beneath the Vatnajökull ice cap in Iceland, drains approximately every six years with discharges up to 50,000 $m^3 s^{-1}$. In non-volcanic areas, the collection and storage of water often takes much longer. In these areas, water production is a by-product of lower geothermal gradients, precipitation, insolation (see INSOLATION WEATHERING) and frictional heat from sliding and deforming ice. Water may be stored in supraglacial, englacial, subglacial or ice marginal positions. Supraglacial drainage is dependent upon connections with englacial or subglacial conduits such as crevasses or MOULINs. The largest known possible reservoirs are subglacial. About seventy subglacial lakes have been identified with radio-echo sounding beneath the Antarctic Ice Sheets. The lakes vary in size from a few kms^2 to 14,000 km^2 with between 4,000 and 12,000 km^3 of stored water.

Outburst floods also develop where proglacial (see PROGLACIAL LANDFORM) dams fail. Commonly in mountainous terrain and during glacial recession, proglacial lakes develop behind moraines or in ice-dammed valleys. Dam failure may be initiated by sudden inputs of water or iceberg calving, and is usually the result of rapid fluvial incision initiated by overflow, internal loss of support, or sapping processes, especially with dams composed of sediment or ice. These hydrographs show a rapid increase in discharge with a slowly falling limb.

Outburst floods may deeply cut canyons into bedrock or sediment, and form extensive outwash plains (sandurs) or discrete gravel bars. Deposits may consist of clast-supported, boulder-gravel sequences greater than 10 m thick, which coarsen upward, and are capped with a fining-upward sequence of gravels, sand and silt. However, boulder-gravel deposits described from Pleistocene-aged subglacial outburst floods tend not to show this fining-upward sequence. In backwater areas, rhythmically deposited couplets up to 15 m thick of fine gravel and sand with rip-up clasts and boulders (also called eddy bars), indicate pulsed flow with high sediment concentrations. HYPERCONCENTRATED FLOWs and DEBRIS FLOWs are commonly associated with outburst floods. Giant current ripples, deposited by the Pleistocene-age Lake Missoula floods that scoured out the Channeled SCABLANDs in central Washington, USA, have wavelengths up to 125 m and are up to 7 m high. These BEDFORMs were also instrumental in the development of expansion and pendant bars composed of foreset-bedded gravel. Such bars are also described from the Interior Plains of North America where outburst floods scoured channels across the prairie surface with discharges estimated at 10^5 $m^3 s^{-1}$. The spillway geometry consists of an inner channel 25–100 m deep and 1–3 km wide, and an upper-scoured zone as wide as 10 km. On the upper scoured zone, channels have an anastomosing (see ANABRANCHING AND ANASTOMOSING RIVER) pattern with residual streamlined hills that resemble DRUMLINS, and boulder lags are common. Within the inner channel, streamlined hills are rare, gravel bars may be found in landslide alcoves or as point bars, and large fans (expansion bars) are found where the outburst flood entered a basin.

Outburst floods have also been used to explain Pleistocene-age subglacial bedforms. In this hypothesis, large subglacial reservoirs are released as sheet flows scouring the subglacial landscape, leaving erosional remnants such as drumlins, fluting, scabland and hummocky terrain up to 100 km wide. Some drumlins and Rogen moraines are explained as sediment moulds from cavities cut into the ice by the outburst floods. Tunnel channels, that often closely resemble spillway channels, are associated with the subglacial bedforms, and develop as the sheet flow collapsed into discrete channels.

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SEE ALSO: geomorphologial hazard; glacier; glacifluvial; palaeoflood; palaeohydrology

TIMOTHY G. FISHER

OVERCONSOLIDATED CLAY

Overconsolidated clays are those that have been highly compressed by burial. The burial may be by superincumbent sedimentation or by short-term loading, but glacial ice is the usual agent. Water is expelled from the clay as it assumes a denser packing. If the clay is subsequently unloaded, perhaps by the ice melting, the clay may become fissured and jointed. An example can be given from the glacial sediments of middle England.

Fissure studies in glacial lake clay and tills at Happisburgh and Cromer show well-defined fissure patterns which can be related to the glacial history of the area and the subsequent erosion history. The presence of fissures influences the strength, consolidation, and permeability characteristics of the clay; the strength along a striated fissure plane is almost reduced to its residual value. The coefficients of consolidation and permeability are significantly increased in the presence of fissures. Attention is drawn to the well-developed fissure systems in tills which have commonly been regarded as non-fissured materials.

Such fissured clay presents a lowered strength as a total mass than would be observed by a triaxial test on a small, unfissured sample. Such materials should be analysed in a manner similar to those used in rock mechanics, which consider the state of discontinuities in lowering the bulk strength. Another consideration in overconsolidated clays is that they have generally been sheared past their peak strength by an earlier loading phase. The subsequent performance of the material will thus be dictated by the residual strength, even though laboratory tests might indicate higher peak values.

During the formation of a sedimentary soil the total stress at any given elevation continues to build up as the height of the soil over that point increases.

The removal of soil overburden, perhaps by erosion (perhaps by bulldozer) results in a reduction of stress. A soil element that is at equilibrium under the maximum stress it has ever experienced is normally consolidated, whereas a soil at equilibrium under a stress less than that to which it was once consolidated is overconsolidated. This means that a clay soil whose *in situ* stress is less than the preconsolidation pressure is, regardless of the cause, called overconsolidated. Various geological and landscape factors responsible for causing preconsolidation stress have been recognized. Mechanisms that cause a preconsolidation pressure:

- Changes in total stress due to: (1) removal of overburden, (2) past structures, (3) glaciation.
- Changes in pore-water pressure due to: (1) change in water table elevation, (2) artesian pressures, (3) deep pumping, (4) desiccation due to drying, (5) desiccation due to plants.
- Changes in soil structure due to: (1) secondary compression, (2) environmental changes, such as pH, temperature, salt concentration, (3) chemical alteration due to: weathering, precipitation of cementing agents, ion exchange.
- Changes in strain rate on loading.

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IAN SMALLEY

OVERFLOW CHANNEL

Ice-dammed lakes are often found in ice-free valleys tributary to glaciers. Meltwaters from snow and ice may be impounded between advancing glaciers and rock walls, between a main valley glacier and a tributary glacier, or by the advance of a glacier in a tributary valley across a valley not occupied by a glacier. Unless water drains beneath the glacier blocking the valley, meltwater accumulates and water levels rise, until the height of the lowest col is reached. At that stage, water overflows either to adjacent proglacial lakes, or to areas free of ice, creating overflow channels. The large volume of water that escapes ensures that the overflow channel is deepened in a comparatively short period of time.

When compared to normal regional drainage patterns, overflow channels are anomalous in terms of position, morphology and size. Characteristically overflow channels are trough-shaped, with flat floor and steep sides, forming an abrupt angle with higher ground above. For many, the longitudinal profile is undulating. Tributary valleys are rare or absent, but if present they display normal river valley morphology. Shorter overflow channels tend to be straight, or nearly so, while larger overflow channels may display a sinuous planform. Overflow channels are sometimes called spillways, although this term is also used for channels created by catastrophic OUTBURST FLOODS following the failure of an ice dam. Most of these channels are now dry, having been abandoned once the ice withdrew and the lakes they drained disappeared. However, on occasions they may be so deepened that the overflow channels retain drainage even after the ice melted and UNDERFIT STREAMS now flow in these channels. Not all anomalous drainage patterns can be explained by overflow, however. RIVER CAPTURE or subglacial drainage, for example, many provide alternative explanations.

Today, proglacial lakes and overflow channels are found in Norway, the Himalayan region, the Rocky and Andes mountains, Baffin Island, Iceland, and particularly Alaska. At the end of the last glaciation, significant volumes of water were impounded around the margin of continental ice sheets. Water levels in these lakes frequently changed configuration, depth and volume because of the interplay between ice margin position, subglacial topography, isostatic rebound and outlet erosion. As water levels changed, often abruptly, water was released into newly opened overflow channels. Thus, complex and extensive overflow channels resulting from late-Pleistocene glacial lakes are found in northern Europe and North America. Sparks (1960) provides detailed descriptions of overflow channels in northern England. Twidale (1968) describes the overflow channel that runs through central Sweden, from Stockholm, through two lakes Mälaren and Vänern and the Göta River, to Göteborg. Teller *et al.* (2002) describe the overflow channels of the largest late-Pleistocene lake, glacial Lake Agassiz, into the Mississippi and Hudson systems.

The term overflow channel has been extended to refer to abandoned channels in a floodplain that may carry water during periods of high flow, or at a dam site to refer to the spillway that can be opened to release lake water when water levels get high enough to threaten the safety of a dam.

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SEE ALSO: outburst flood

CATHERINE SOUCH

OVERLAND FLOW

Overland flow is the term used for water that flows over the surface of hillslopes. It is important because this route provides the fastest means by which rain falling on hillslopes can reach stream channels. Hence overland flow contributes significantly to the shape of a catchment storm hydrograph. Equally, it may be responsible for high erosion rates on hillslopes. Other terms that are used in this context are sheet wash (see SHEET EROSION, SHEET FLOW, SHEET WASH) and interrill flow, both of which denote unchannelled flow, and RILL flow, which is used to describe overland flow where it becomes concentrated into definable channels on the hillslope. Though interrill flow does not exhibit definable channels, it is common to observe that the flow is not of uniform depth. Instead, the flow converges and diverges around microtopographic obstacles forming anastomosing threads of deeper and faster flow within a layer of water that covers most of the surface. It is generally accepted that the presence of rills indicates that the flow is able to detach and transport sediment, whereas interrill flow, though capable of transporting sediment, does not have the erosive power to detach sediment. Instead, the sediment load of interrill flow is supplied to it by the detaching force of impacting raindrops. Three types of overland flow may be recognized. The first is that which is due to rain falling at an intensity in excess of the rate of infiltration into the soil. This type of overland flow is also termed Hortonian overland flow after R.E. Horton, who first described the process (Horton 1933). The second type of overland flow is termed saturation-excess overland flow, and the third is termed return flow.

Hortonian overland flow

According to Horton's description of the generation of overland flow, rain reaching the soil can be separated into two parts: one infiltrates into the soil, the other remains on the surface. The rate at which rain infiltrates and its relationship to the rainfall intensity is the basis of Horton's model for the generation of overland flow. Horton developed the equation

$$f = f_c + (f_o - f_c)e^{-kt}$$

to describe the way infiltration would change during a storm, in which f is the maximum instantaneous infiltration rate, f_c is minimum infiltration rate (infiltration capacity), assumed to be a constant for a given soil, f_0 is the initial (maximum) infiltration rate (at $t = 0$), k is a constant that varies with soil type, and t is time since the onset of rain.

In general, f_0 will be in excess of all but the highest rainfall intensities, so that initially all rain infiltrates. As the rainstorm proceeds, the pore spaces of the soil fill with infiltrated rainwater, cracks in the soil close and fine particles wash into the surface of the soil so that the instantaneous infiltration rate declines through time. Eventually, it may fall to the point where it is below the rainfall intensity, at which time some of the falling rain remains on the surface. The time taken for this to occur is known as the time to ponding and its completion can be recognized by glistening of the surface and the appearance of ponds of water in small depressions on an irregular ground surface. As these depressions fill with water they begin to overtop and interconnect, until the ground surface is covered by a connected series of these pools. The amount of water that is required for this to occur will vary with the surface irregularity of the hillslope, and is known as the depression storage. Once this stage is reached, flow from pool to pool begins to occur throughout the hillslope and water is discharged from the hillslope as overland flow. As the rainstorm proceeds, the rate of infiltration into the soil continues to decline so that the amount of water remaining on the surface increases. As the volume of water at the ground surface increases, so does the discharge from the hillslope. Once the infiltration capacity f_c of the soil has been approximated (assuming rainfall rate is constant), the discharge of water from the hillslope will also begin to level out towards an equilibrium value. The layer of moving water is known as the surface detention. The thickness of this layer and the time delay between the approximation of the soil's infiltration capacity and the achievement of equilibrium runoff will be a function of the runoff hydraulics (depth and velocity) and the length of the hillsope. Runoff hydraulics are, in turn, primarily controlled by the surface roughness of the ground, which determines the frictional resistance it affords to the flowing water. The relationship between water depth and velocity, on the one hand, and frictional resistance, on the other, can be expressed through the Darcy–Weisbach equation:

$$ff = \frac{8gds}{v^2}$$

in which ff is the dimensionless Darcy–Weisbach friction factor, g is the gravitational constant (m s^{-2}), d is water depth (m), s is slope (m m^{-1}) and v is water velocity (m s^{-1}).

Under the Hortonian model for the generation of overland flow it is assumed that flow will be generated more or less simultaneously over entire hillslopes. This is most likely to be the case where soils have very low initial soil moisture at the start of rainfall and/or the rainfall is intense and/or the soil has a very low infiltration rate. These conditions are most commonly met on bare soils (such as cleared agricultural land), in arid and semi-arid environments, and during convective thunderstorms during which peak rainfall intensities of 300 mm hr^{-1} can be attained and many minutes of rainfall at intensities exceeding 50 mm hr^{-1} are not uncommon. Bare soils may have quite low infiltration capacities because of crusting (either biological or mechanical) (see CRUSTING OF SOIL) of the surface. Kidron and Yair (2001) report infiltration capacities as low as 9 mm hr^{-1} on crusted dune soils in Israel, so that Hortonian overland flow can be generated at even quite low rainfall intensities.

Because Hortonian overland flow depends on the relationship between rainfall rate and infiltration rate discharge increases downslope. The ways in which this increase in discharge downslope affects flow width, depth and velocity (the HYDRAULIC GEOMETRY) is highly variable and depend primarily on the characteristics of the hillslope surface (Parsons *et al*. 1996). Both laminar and turbulent flow conditions are present and the flow may vary from laminar to turbulent both spatially and temporally. Horton (1945) used the term mixed flow to characterize this condition.

The downslope increase in discharge may be accompanied by a change from wholly unchannelled flow on the upper part of the hillslope, to a mix of channelled (rill) and unchannelled (interrill) flow on the lower part. The emergence of eroded channels under the operation of overland flow led Horton to term the upper part of hillslopes without rills the belt of no erosion. It is, however, not correct that no erosion takes place in this zone: simply that detachment in this zone is accomplished by raindrops and is spatially diffuse, as it is everywhere in interrill flow. Soil detachment by falling raindrops is controlled by the kinetic energy of the rainfall, but is diminished as the depth of interrill flow increases because the water dissipates some of the energy of the rainfall. The relationship may be expressed by the equation (Morgan *et al*. 1998)

$$D = \frac{k}{\rho_s} KEe^{-bd}$$

where D is the detachment rate, k is an index that varies with soil type, ρ_s is particle bulk density, KE is rainfall kinetic energy and b is a constant that varies with soil texture.

In rills, detachment is achieved by the shear stress exerted by the flow. Although there have been several attempts to quantify the threshold conditions for rill initiation by overland flow (e.g. Slattery and Bryan 1992), Nearing (1994) has pointed out that the shear stress exerted by shallow flow is of the order of a few pascals, whereas the shear strength of soils is of the order of a few kilopascals.

Saturation-excess overland flow

The acceptance of Horton's model for the generation of overland flow is at odds with the fact that it is very seldom observed in many environments, particularly those in which there is appreciable vegetation cover and/or rainfall is cyclonic, rather than convective. Kirkby and Chorley (1967) argued that rain falling onto an already saturated soil will also remain at the surface and become overland flow. Generation of this type of overland flow depends not so much on the relationship between rainfall intensity and soil infiltrability as on the amount of water that is already in the soil at the onset of rainfall, known as the antecedent moisture content, and the water-storage capacity of the soil. These amounts are both spatially and temporally variable. Antecedent moisture is likely to be highest on footslopes and in concavities, and areas with thin soils (such as spurs) have low total water storage capacity. Both of these areas will preferentially generate saturation-excess overland flow. Antecedent moisture will also depend on rainfall record prior to an individual storm event. Taken together, these two factors mean that, in contrast to Hortonian overland flow, saturation-excess overland flow is likely to be generated on only some parts of hillslopes (the concept of partial area contribution to overland flow – see Betson and Marius 1969), and be variable for two storms of similar characteristics (the concept of variable source area – see Dunne and Black 1970). Because saturation-excess overland flow is generated locally, particularly in areas close to rivers, it is an important control on catchment hydrographs. Conversely, because much is generated on low-angle footslopes, it is of much less importance for soil erosion on hillslopes.

Return flow

Water that infiltrates into the soil and moves downslope through the soil as throughflow or in pipes (see PIPE AND PIPING) may encounter saturated soil, thereby having its further downslope movement through the soil blocked. This water may be forced to the surface, where it is known as return flow, and travel further downslope as overland flow. Like saturation-excess overland flow, return flow is generated locally, often on footslopes, so its significance lies in its impact of catchment hydrographs.

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SEE ALSO: soil erosion

A.J. PARSONS

OVERWASHING

Overwashing is generally regarded as the process of sediment transport across a BEACH RIDGE or barrier beach (see BARRIER AND BARRIER ISLAND), with deposition as a washover deposit on the back slope of the ridge or in the lagoon landward of the ridge occurring during storms. Morton *et al.* (2000) have, however, reported frequent overwash events occurring during non-storm periods. The term ‘overwash’ is generally applied to the process, and ‘washover’ to the resulting despositional landform. Overwashing is a major process by which the back slope of a barrier beach is renewed with the barrier building to landward. Overwash-dominated barriers tend to be relatively narrow and flat, with low and unstable dune systems. Where sediment supply is abundant, overwash barriers may grow and stabilize, but where sediment supply is limited, overwash causes barriers to roll over with sediment being moved from the seaward slope to the landward slope, thereby causing the barrier to migrate landward. Overwashing is an important process on both sand-dominated barriers (where AEOLIAN PROCESSES may also be important) and gravel barrier systems.

Overwash can occur along a significant length of barrier crest producing a washover ramp on the landward side, but more commonly flow is concentrated in channels called overwash throats. In some circumstances and in the long term an overwash throat may lead to the development of a tidal inlet. Washover fans develop on the landward side of the barrier where flow is no longer constrained. Orford and Carter (1984) relate the spacing of overwash throats and washover fans to beach rhythmic morphology (see BEACH; BEACH CUSP; WAVE).

Overwashing (which generally leads to the lowering of a beach barrier) must be differentiated from overtopping which is the process by which barrier systems are built up due to swash transport of sediments to the top of a berm or barrier crest, thereby causing the barrier to build in height and width. Overtopping and overwashing may occur at the same locations, dependent upon wave energy conditions.

Although often considered as a somewhat different process, overwash sedimentation also occurs on CORAL REEF islands, and can be important for the maintenance and development of reef beaches and cays.

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SEE ALSO: beach ridge; storm surge

KEVIN PARNELL

OXBOW

An abandoned meander (see MEANDERING) loop along an alluvial river. It is the most common type of lake of fluvial origin. As a logical consequence of the lateral shifting and the general downstream migration of meanders, it develops from the interplay of channel erosion and accumulation separated in space. Oxbows are produced by meander cutoffs, which occur in two different ways. If a meander neck becomes narrow enough, streamflow is directed along the shortest route of greatest slope, instead of following the whole perimeter of the meander loop (neck cutoff). Alternatively, a new channel may develop along a swale between POINT BARS (chute cutoff). Natural LEVEE formation and FLOODPLAIN deposition soon build a silt or clay plug between the oxbow and the main channel, although a narrow batture (watercourse) may provide a connection. The American name emphasizes the crescent shape, while in many other languages an oxbow is called a 'dead arm' (e.g. in French: *bras mort*). Both types of cutoff cause channel shortening and scour upstream and deposition downstream. Thus the meandering river maintains its average SINUOSITY since each cutoff triggers the formation of further cutoffs on the long term (self-organization – Stolum 1996). The tranquil freshwater makes an oxbow a valuable aquatic habitat. Meander scars are oxbows completely filled up with mineral and organic matter. They remain discernible in the landscape for a long time.

Classic examples are found along several major rivers of the world, including the Mississippi, the Amazon and the large rivers of Siberia. Floodplains of some minor (regulated) rivers also abound in oxbow lakes.

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DÉNES LÓCZY

OXIDATION

Oxidation is the loss of a negative electron so an element becomes more positively charged, for example ferrous iron, Fe^{2+} (or Iron II) becomes oxidized to ferric iron Fe^{3+} or Iron III. This process commonly occurs in the presence of oxygen:

$$\underset{\text{Iron II}}{4FeO} + O_2 \rightarrow \underset{\text{Iron III}}{2Fe_2O_3}$$

Oxidation is thus a common weathering reaction when a mineral formed in an anoxic environment becomes exposed to air at the surface of the Earth. The process is often combined with HYDROLYSIS, for example the weathering of olivine:

$$\underset{\text{Olivine}}{3MgFeSiO_4} + H_2O \rightarrow \underset{\text{Serpentine}}{H_4Mg_3Si_2O_9} + SiO_2 + \underset{\text{Ferrous oxide}}{3FeO}$$

the FeO then becoming oxidized to iron oxide, as above.

During oxidation, the strength of the minerals is reduced which also makes mechanical breakdown much easier.

The loss of a negative electron can equally occur when iron in an acid solution becomes less acid. The latter process accounts for the deposition of reddish iron oxides in the less acid lower parts of soil profiles where there can be, in fact, less oxygen than nearer the surface.

Iron oxides are an important constituent of tropical soils, being produced as a residual mineral through prolonged, intense weathering. The simple iron oxide, haematite (Fe_2O_3) is bright red and leads to the distinctive colour of soils in tropical and subtropical areas. If haematite is subject to HYDRATION then limonite ($2Fe_2O_3{\cdot}3H_2O$) forms which is yellow in colour. The content of iron also appears to be a key factor in the formation and hardening of laterite (Thomas 1974). Many of the theories of laterite formation involve the movement of iron in solution by mobile groundwater, or by upward diffusion, and their subsequent oxidation and mobilization near the surface; the harder laterites having a higher iron content.

Oxidation is not only a weathering process in itself, it can produce further weathering agents. The oxidation of iron sulphides (pyrites) can produce sulphuric acid:

$$\underset{\text{Pyrite}}{2FeS_2} + 15(O) + H_2O \rightarrow 2Fe^{3+} + \underset{\text{Sulphuric acid}}{4SO_4^{2-} + 2H^+}$$

There is a notable example of this at Mam Tor in Derbyshire where pyrite oxidation and the production of sulphuric acid leads to the intense weathering of the illite and kaolinite present in the shale in which the pyrite occurs, leading to a marked rise in porosity and facilitating slope instability, contributing to the collapse of a road (Vear and Curtis 1981). The authors calculate that for 1.5 g of pyrite oxidized by a litre of acid-sulphate water, $0.0125\,\mathrm{g\,l^{-1}}$ H^+ is produced.

While geomorphologists often focus on the oxidation of iron, other compounds can also be oxidized, for example manganese, and a key process in the nitrogen cycle involves the oxidation of ammonium produced by the decomposition of organic matter to nitrate:

$$\underset{\text{Ammonium}}{4NH_4^+} + 6O_2 \rightarrow \underset{\text{Nitrite}}{4NO_2^-} + 8H^+ + 4H_2O$$

and

$$\underset{\text{Nitrite}}{4NO_2^-} + 2O_2 \rightarrow \underset{\text{Nitrate}}{4NO_3^-}$$

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STEVE TRUDGILL

OYSTER REEF

'Among organic reefs, those of the geologically young oysters are now second in size and distribution only to the coralline reefs' (Price 1968: 799). The reefs are built by the modern estuarine *Ostrea* and *Crassostrea* and the marine *Pycnodonte*. The tops of the reefs range from the intertidal zone to depths of as much as 12 m below sea level. Optimum temperatures are from 15–25 °C and optimum salinities occur in the central parts of bays and other estuaries midway between stream mouth and oceanic opening. In addition to forming reefs and visors, they can help to stabilize spits and other constructional features. Oyster reefs are widespread, and major locations include the Gulf of Mexico (*Crassostrea virginica*) and parts of China (*Crassostrea gigas*) (Wang Hong *et al*. 1995), but in some parts of the world they are being destroyed by human activities (including gastronomy and fishing); this has led to attempts to restore them or to provide artificial substances for their colonization (Coen and Luckenbach 2000).

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A.S. GOUDIE

P

PALAEOCHANNEL

When a channel ceases to be part of an active river system it becomes a palaeochannel. Palaeochannels vary greatly in age. Those from the recent geological past (perhaps tens to hundreds of years old) include meander cut-offs and longer reaches abandoned by AVULSION. Although these channels are now isolated from the active river flow, except during periods of floodplain inundation, they are scaled to present flow regime. Many of the more ancient palaeochannels indicate discharges greatly in excess of those occurring at present. The oldest palaeochannels known are found on Mars where huge flows of surface water carved a complex pattern of channels more than 3.5 billion years ago. Clearly, the Martian channels could not have formed in the planet's present waterless environment.

Where palaeochannels are well preserved they provide valuable information about past flow regimes. The basis of discharge reconstruction is given by established statistical relationships between channel forming (bankfull) discharge and aspects of channel morphology including cross-sectional area and meander wavelength as documented by the US Geological Survey in the 1950s and 1960s and amply confirmed since. Meandering rivers are particularly suited to this work because of their potential to preserve planform and, sometimes, cross-sectional geometry. The relationship between meander wavelength and stream discharge (Dury 1965) provides a useful, although often imprecise, approximation of palaeoflow. Rotnicki (1983) argued, on the basis of fieldwork on the Prosna River in Poland, that channel cross sections in meander neck cut-offs provided more reliable estimates because of the excellent preservation of channel dimensions at such locations.

The high degree of channel preservation in neck cut-offs results from their mode of formation. Upon initiation of a short circuit, sedimentation rapidly seals the cut-off ends to form an OXBOW lake. In the low energy environment of the lake, fine-grained sediments from suspension form a drape over the old riverbed. The infill sediment cast effectively preserves the former channel cross section, which can be revealed subsequently in a series of auger holes.

In Rotnicki's (1983) comparative study, estimates of BANKFULL DISCHARGE based on fourteen equations linking meander wavelength and discharge were compared with estimates based on preserved cross-sectional dimensions at cut-offs. For a measured bankfull discharge of $22.5\,m^3\,s^{-1}$ the estimates based on wavelength ranged from 0.2 to $34.1\,m^3\,s^{-1}$, or more than two orders of magnitude. The frequently used equations of Dury and Carlston gave errors of 36 per cent and 78 per cent respectively. Estimates of bankfull discharge based on cut-off cross sections and the Manning Equation reduced the error to 10 per cent.

Many palaeochannels also indicate past regimes of markedly different channel pattern and sediment load. The transition from late glacial conditions to those of the Holocene produced a strong channel response globally. In regions directly affected by ice, many large braided (see BRAIDED RIVER) and bedload-dominated proglacial channels gave way to meandering mixed and suspended load channels. In North America the Mississippi provides an excellent example. At the same time, in regions far removed from the great Quaternary ice sheets parallel changes occurred. For example, on the Riverine Plain in southeastern Australia, low sinuosity, aggraded sand-bed channels (PRIOR

STREAMS) were converted to highly sinuous systems dominated by fine-grained sediments. Here, changing upper catchment conditions including rising temperatures and treelines, and shrinkage of the winter snowpack, produced the channel response. Selected examples of palaeochannels reported in the scientific literature are presented below.

Underfit streams

From the late nineteenth century it was recognized that some sinuous valleys in Europe contain floodplains on which present-day rivers describe smaller wavelengths than the enclosing valley (Davis 1899). Such streams were described as being manifestly (obviously) underfit. The inference being that a reduction in discharge had resulted in a reduction in meander wavelength. George Dury's (1964a,b, 1965) detailed studies in Europe and North America demonstrated the fluvial origin of large meandering valleys and the widespread regional distribution of UNDERFIT STREAMS. Following the elimination of other possible causes, including headwater capture and the loss of glacial meltwater, Dury deduced that regional climatic change had been responsible for the observed reduction in discharge. Radiocarbon dating of valley fills indicated that the last major discharge shrinkage occurred between 10,000 and 12,000 years ago, at the beginning of the Holocene.

Dury estimated the discharges of the large palaeochannels largely on the basis of the statistical relationship between meander wavelength and discharge. Although the computed discharges exceeded those of the present rivers by up to a factor of 60, Dury argued that they could have been produced by glacial climates characterized by reduced evapo-transpiration and a 50–100 per cent increase in precipitation. Many of the reconstructed discharges approach the largest discharges ever recorded on Earth for catchments of equal drainage area and thus were considered by many workers to be excessive, especially in areas of low relief unsuited to the production of extreme flows. In particular, there was concern about the influence of parameters other than discharge on meander wavelength. Various studies have subsequently shown that meander wavelength alone does not provide reliable estimates of bankfull discharge.

Superflood palaeochannels

In the 1920s, J. Harlen Bretz (1923) first described superflood palaeochannels in the Columbia Plateau region of the northwestern United States. Here, space imagery reveals a complex network of large anastomosing (see ANABRANCHING AND ANASTOMOSING RIVER) channels carved into basalt bedrock and overlying LOESS and other sediments. Fluvial features include great waterfalls, potholes, longitudinal grooves carved into the bedrock, and boulders that were transported and deposited in flood bars and giant current ripples. Bretz attributed these features to a cataclysmic flow he called the Spokane Flood.

The eventual verification of Bretz's catastrophic flood hypothesis, despite vehement opposition at the time of its proposal, is considered by Baker (1978) to be one of the most fascinating episodes of modern science. On the basis of flow reconstructions by Shaw *et al.* (1999) it now appears that the giant scabland floods arose from a combination of sources including late Pleistocene ice-dammed Lake Missoula and large subglacial reservoirs that extended over much of British Columbia. An estimated total volume of $10^5\,km^3$ flowed across the Scablands and achieved a peak discharge of some $17 \times 10^6\,m^3\,s^{-1}$. The power per unit area of streambed generated by these flows was up to 30,000 times greater than that produced in the present Amazon.

Similar outburst floods (jökulhlaups) have been documented for spillways marginal to the Laurentide Ice Sheet and in Swedish Lapland. Cataclysmic flows generated by Pleistocene ice-dammed lake failures in the Chuja Valley in the Altas Mountains of south-central Siberia (Baker *et al.* 1993), which exceeded $18 \times 10^6\,m^3\,s^{-1}$, are comparable to the largest of the Channeled Scabland flows. Spectacular palaeochannel landforms in the Chuja Valley include scoured channels, giant bars and gravel wave trains. The impressive hydraulic parameters associated with the Chuja Valley floods include flow depths of 400–500 m, supercritical flow velocities of $45\,m\,s^{-1}$ and stream powers approaching $10^6\,W\,m^{-2}$. The outburst floods of the Channeled Scabland and Chuja Valley are Earth's greatest known terrestrial discharges of freshwater.

Landform assemblages characteristic of cataclysmic flooding are also present on Mars (Baker *et al.* 1993). The Martian outflow channels, which were first recognized on the basis of Mariner 9 and Viking space mission imagery, are much larger than those of the Channeled Scabland and may have experienced discharges as great as $10^9\,m^3\,s^{-1}$. The Martian palaeochannel systems are therefore not only the largest known, but also the oldest, dating from before 3.5 billion years ago.

Palaeochannels in Australia

Australia's modest stream discharges, compared to those of other continents, result from its predominantly subtropical location and low average relief. Not surprisingly, the presence of large palaeochannel systems in Australia's two inland drainage basins with areas exceeding 1,000,000 km^2, the Murray–Darling and Lake Eyre (Figure 112), has been of particular interest to geomorphologists seeking to reconstruct Late Quaternary hydrological regimes.

In the Murray–Darling Basin, palaeochannels of the Murrumbidgee River have been studied extensively since the late 1940s. Previously described as prior streams, large palaeochannels here form an impressive distributary system in a region now characterized by small meandering rivers. Research before 1970 subdivided the palaeochannels into two genetically different categories: older prior streams and younger ancestral rivers. Channels described as prior streams were aggraded bedload systems characterized by low sinuosity, high width to depth levees and source bordering sand dunes. The sinuous ancestral channels were characterized by floodplains of lateral migration and discharges much larger than the present rivers in this region (Plate 84).

Thermoluminescence (TL) dating by Page *et al.* (1996) resulted in a major revision of the prior/ancestral model. It was shown that four major surface palaeochannel systems (Coleambally, Kerarbury, Gum Creek and Yanco) operated between 100,000 and 12,000 years ago with frequent alternations between prior and ancestral modes of channel behaviour. Stratigraphic investigations showed that bedload aggraded channels (prior streams) typically were bordered by fining-upwards deposits associated with laterally migrating channels (ancestral rivers). Clearly, there was a need to revise the existing model in which prior steams preceded ancestral rivers. Page and Nanson (1996) proposed that the first three phases of Murrumbidgee palaeochannel activity were characterized by alternations from laterally migrating to vertically aggrading channel behaviour with each phase terminating in vertical aggradation and the formation of source-bordering sand dunes (Figure 113). Only the final (Yanco) phase failed to terminate in bedload aggradation, probably because the onset of Holocene climates reduced the size of flood peaks, greatly diminished the supply of bedload from the upper catchments, and resulted in streams evolving into their present highly sinuous suspended load morphology.

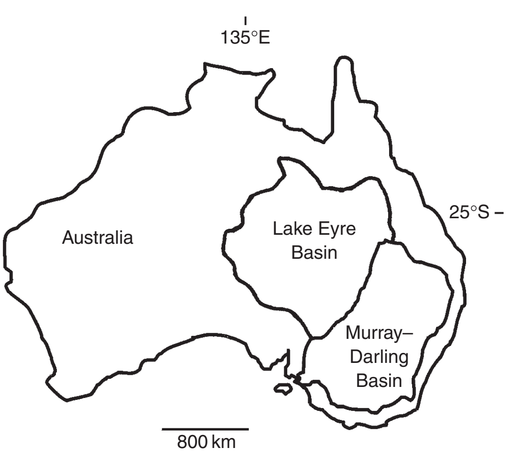

Figure 112 Map of Australia showing locations of Lake Eyre and Murray–Darling Basins

Plate 84 (a) Ancestral Green Gully palaeochannel, and (b) present channel of Murray River in southern Australia at same scale and drainage area

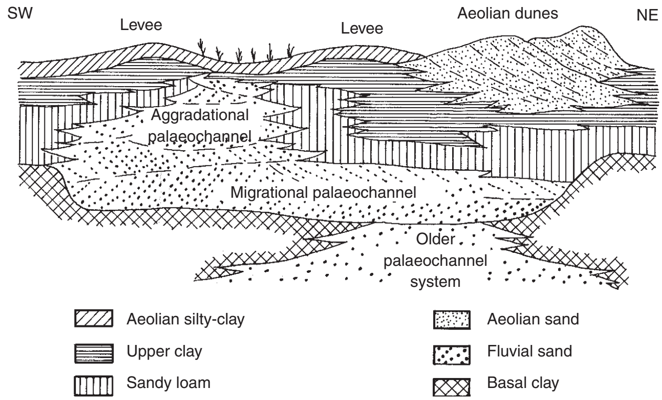

Figure 113 Stratigraphic model of Murrumbidgee River palaeochannels (Page and Nanson 1996: 943)

Discharge reconstructions at preserved channel cross sections of Murrumbidgee palaeochannels suggest that bankfull flows were between four and eight times greater than those of the present rivers.

The Channel Country of the Lake Eyre Basin (Figure 112), which includes Cooper Creek and the Diamantina River, comprises a vast system of low-gradient anastomosing channels dominated by fine-grained suspended load. The anastomosing channels, which date from about 80,000 years ago, are mud-lined, laterally very stable and underlain by extensive muddy floodplains. However, along the middle and lower reaches of Cooper Creek aerial photographs and subsurface exploration have revealed remnant scroll-bars and palaeochannels beneath the mud unit. The scrolls, which are scaled to river meanders larger than any present in the system today, were formed by mixed-load, laterally migrating rivers that deposited extensive sandy units with abundant flow structures (Katipiri Formation). TL dating by Nanson *et al.* (1988) showed that the Katipiri sands date from at least 250,000 years ago.

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KEN PAGE

PALAEOCLIMATE

The climate of the past: palaeo, from the Latin word meaning ancient or old; climate, refers to the interconnected group of Earth systems that control weather conditions (temperature, moisture, wind, etc. and the spatial/temporal variation in these factors) at the surface over an extended period of time. The study of past climates is referred to as palaeoclimatology. More specific definitions of palaeoclimate exist (e.g. climate prior to instrumental records) but in the present case a broad definition of palaeoclimate is taken as that meaning all climates prior to the present day. Further divisions of specific geological time periods may then be defined (e.g. Holocene, Quaternary, Permian) and some of these are discussed below briefly.

Climate has a significant influence on most geomorphic processes and an understanding of landscape systems cannot be achieved without knowledge of both present day climate and climatic history. Antecedence in geomorphic systems is often controlled to a large degree by palaeoclimate and interpretation of geomorphic records (e.g. through sedimentological records) is dependent upon an appreciation of climatic variations.

Climate varies on all temporal and spatial scales and the notion of scale is particularly important both in geomorphology and in climate studies. The nature of climate change at each spatial/temporal scale is determined to some extent by the factors forcing the change. Over long 'geological' timescales, the movement of tectonic plates, and associated volcanism, effect far-reaching but gradual changes in global climate. Over periods of hundreds of thousands of years, the influence on insolation of orbital variations (as described by the 'Milankovitch theory') may have driven the large-scale, high amplitude, changes in global climate of the Quaternary period (see below). Both tectonic and orbital forcing are examples of factors 'external' to the climate system. At shorter timescales (e.g. at the millennial scale), variations are probably due largely to 'internal' factors, such as the differences in response time of components within the climate system, and the subsequent chaotic dynamics of these strongly coupled (non-linear) systems. Internal controls often give rise to abrupt changes in climate as thresholds are reached and feedback systems evolve. Over these shorter timescales, the climate system is highly complex, meaning that cause and effect are often difficult to separate.

Information on past climate conditions can be obtained, in some cases, from historical records (e.g. farming records, instrumental data). However, such records are often sporadic, of questionable accuracy and limited in extent, both spatially and temporally. A second method is to employ the use of climate-dependent natural phenomena that leave traces in the geological record. Once calibrated, these traces can then be used as proxies for past climates. Calibration involves defining the dependence of each proxy on climate, sometimes using modern analogues, sometimes using theoretical calculations (the principal of uniformitarianism then applied). Broadly, there are two kinds of proxy data: (1) episodic/discontinuous records (e.g. flood deposits, glacial advances), which result from the integration of climatic conditions prior to the event; and (2) continuous/incremental records (e.g. constant accumulation of marine mud) which preserve a quasi-continuous record of environmental/climatic conditions. Some examples of proxy records are:

1 *Glaciological*: composition and macrostructure of ice cored from both large ice sheets (e.g. Antarctica, Greenland) and from mountain glaciers – provides information on both regional air temperatures and on global/regional atmospheric composition;
2 *Geological*: ocean sediment cores (geochemistry – particularly oxygen isotope profiles – and species composition of micro fauna), glacial features, sedimentary deposits (e.g. loess, sand dunes), chemistry of speleothems;
3 *Biological*: pollen recovered from terrestrial/marine sediments, remains of insects and micro fauna, composition and structure of tree rings.

The degree to which these proxies are affected by global/regional/local influences depends upon many factors (some of which are context specific) and, to some extent, what we see depends upon how we choose to look at the climate record. Broadly, globally integrated climate signals can be obtained from marine sediment geochemistry, loess/palaeosol deposits and ice sheet cores. However, all proxies are affected by both random fluctuations (noise), non-climate related processes and lags in response times, and all have different sensitivities to climatic conditions. These factors underline the importance of multi-proxy datasets in palaeoclimate research.

As with other geographical matters, it is not easy to discuss palaeoclimates as sets of primary data or even calibrated indices (e.g. temperatures) and further interpretations are required (sometimes quantitative, sometimes qualitative, e.g. 'wetter' or 'warmer'). For conceptual ease, palaeoclimates are commonly discussed in terms of simplifications such as by warmer/colder, drier/wetter or in other even more generalized terms such as reference to glacial/interglacial conditions. Care is needed when using such terms as they often have different meanings in different parts of the world. However, on the regional scale, these terms are indeed useful and are often retained in the palaeoclimate literature.

Climate has varied widely over the history of the Earth. Geological evidence from Precambrian times (approx. 860–550 Ma) points to large-scale glaciations, although the extent of ice cover during this period is debated. Some evidence suggests the Earth was almost completely ice covered, with glaciers reaching to the tropics, while other evidence puts the effected land masses at higher latitudes during these times. Since the Cambrian, the Earth's mean temperature has varied considerably, between the icy conditions of the Permian ice age to the more tropical temperatures of the late Cretaceous. Over the last ~55 Ma, there is strong evidence for a cooling trend at both poles and across the lower latitudes, culminating in the glacial conditions of the last ~2 Ma, the Quaternary period. While the deep geological past sets the wider stage on which contemporary processes operate, it is perhaps the Quaternary period, with its high amplitude, high frequency climatic changes, that is of most significance in understanding the geomorphic setting of most present- day landscapes.

Quaternary palaeoclimate

During the Quaternary period (~2 Ma) the Earth's climate has been through many oscillations in climate, that have operated over a range of temporal and spatial scales. Perhaps the most characteristic feature of the Quaternary has been the oscillation between glacial and interglacial conditions. During glacial conditions, large continental ice sheets grew on Northern Europe/Eurasia and Canada/North America, causing SEA LEVELS to fall by up to 120 m compared to the present day. During the warmer interglacials, ice sheets were restricted to the poles and to Greenland and temperatures were similar to those experienced at present. Insolation forcing due to orbital variation (the Milankovitch theory) at periods of roughly 100, 40 and 20 ka are believed to play a key role in these climate oscillations, although other factors cannot be ruled out. At lower latitudes, the effects of insolation forcing were different, affecting, for example, the intensity and distribution of precipitation (particularly the monsoon systems).

Other factors such as heat distribution due to the ocean current circulation and atmospheric gas composition are likely also to have played key roles. Some lines of evidence (e.g. ice and marine core data) suggest rapid high magnitude changes also occurred during the Quaternary, with mean regional temperatures changing by as much as 10 °C over decades or centuries. Such changes would have significant effects on geomorphic processes and these effects are often recorded in the sedimentological and morphological record.

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SEE ALSO: El Niño effects; Holocene geomorphology; ice ages

RICHARD BAILEY

PALAEOFLOOD

Palaeoflood literally means 'ancient flood', but the word does not necessarily connote a specific age, and is often used for any flood not systematically gauged. The characteristics of ungauged floods may be inferred using historical, botanical or geological evidence (Wohl and Enzel 1995). Historical evidence comes from qualitative flood records kept by humans. High-water marks on buildings or canyon walls, diary entries, newspaper reports or damage reports for insurance purposes may all be used to estimate the magnitude and date of occurrence of floods. Such records may extend back 2,000 years in countries such as China.

Botanical evidence of past floods comes from vegetation growing along the riparian corridor (Hupp 1988). Flood-borne debris may impact riverside trees hard enough to destroy a portion of the tree's cambium and leave a corrasion scar that can be dated using annual growth rings in some tree species. Maximum scar heights may be used to estimate minimum peak stage. Flood debris may also bend or break trees near the channel. Many tree species can survive such damage and develop adventitious sprouts, usually within a year of the damage, which can also be dated using tree rings. Anomalies in the width or symmetry of annual growth rings result from changes in water availability, tilting of the tree or stripping of the tree's leaves, all of which may be associated with floods. The age structure of riparian vegetation indicates minimum time since initial deposition or scour of alluvial surfaces. Each of these botanical indicators may provide chronologically precise information on a range of flows for species with annual rings, but the use of these indicators is limited by the presence and age of the vegetation.

Geological evidence of ungauged floods may come from dimensions of relict channels, the size of fluvially transported sediment, or erosional and depositional features that indicate maximum flood stage. Palaeochannels may be preserved as exposed cross sections, abandoned channels on the surface or exhumed channels (Williams 1988). Form parameters including drainage density, terraces, channel pattern, meander wavelength and channel cross-sectional dimensions have been used to infer flow parameters including mean velocity and discharge by means of empirical equations developed for active channels. Many of these form parameters are best preserved along low gradient, unconfined alluvial rivers where continued lateral movement of the channel has left abandoned channels relatively well preserved. Along these rivers, form parameters commonly record low magnitude floods such as average discharge or mean annual discharge. Estimates of flow parameters using channel characteristics may be inaccurate because of deficient regression equations based on limited data; misapplication of available equations caused by extrapolation to different conditions of channel pattern and climate; inadequate preservation of abandoned channels; improper algebraic manipulation of the empirical equations; and uncertainty in the definition of some variables (Williams 1988).

Sediment characteristics may be related to flow parameters by first relating particle size to some index of local transport capability, such as STREAM POWER, and then transforming the transport variable into a discharge estimate using a hydraulic flow equation such as the Manning equation. Gravel and finer sediments may be used in the aggregate to reflect average flow. This approach is commonly used for lower gradient alluvial channels. Coarser sediments are often treated as individual particles, with a focus on the flow competence necessary to transport the largest particles present (O'Connor 1993). This approach is more commonly used for higher gradient confined channels such as bedrock canyons. Use of both finer and coarser sediments to estimate flow parameters relies on empirical relations developed between transported particles and observed, calculated or inferred flow conditions.

Erosional and depositional features may provide palaeostage indicators that record the maximum stage of individual flows. Erosional features include lines scoured into valley-wall soil and colluvium; truncation of landforms such as debris flow fans impinging on the channel; or vegetation limits below which individual species of vegetation are absent (Jarrett and Malde 1987). Depositional features include silt lines of very fine sediment and organics adhering to the channel banks; accumulations of organic debris from fine particles to logs; and slackwater deposits of sediment settling from suspension in areas of flow separation such as tributary mouths or channel-margin alcoves or caves (Kochel and Baker 1982). Palaeostage indicators are best preserved along confined channels with resistant boundaries where an increase in discharge produces a large increase in stage, and changes in channel geometry during and between floods are minimized; and in drier climates where the indicators are less likely to be weathered or obliterated by non-fluvial processes. Flood chronologies may be established from palaeostage indicators using both absolute geochronologic methods such as radiocarbon or thermoluminescence, and relative methods such as stratigraphic position or soil development. Combined with surveyed channel geometry, the stage indicators can be used to estimate flood magnitude (Webb and Jarrett 2002). Palaeostage indicators are commonly used to estimate the largest floods along a channel.

The majority of palaeoflood studies address floods that occurred during the late Pleistocene and Holocene. The late Pleistocene was characterized by immense outburst floods such as those in the Channeled SCABLAND and Siberia produced by the release of meltwater ponded along the margins of the continental ice sheets.

Geological methods used to estimate palaeoflood magnitude on Earth have also been applied to channels on Mars (Baker 1982).

Palaeoflood studies are distinguished from other types of fluvial palaeohydrology in that they usually focus on maximum flows along a channel rather than the entire range of flows. Palaeoflood studies may be a part of studies focusing on channel change as recorded in terraces (see TERRACE, RIVER), ARROYO formation or COMPLEX RESPONSE, or palaeoflood data may be used to examine issues of flood-frequency analysis, flood hydroclimatology and the geomorphic effectiveness of floods.

Flood-frequency analysis is largely based on the measured or extrapolated recurrence interval between discharges of a given magnitude. Measured recurrence intervals are limited by the time span of systematic discharge measurements, which is rarely longer than a hundred years. Extrapolated recurrence intervals may come from extending an existing flood-frequency curve beyond the time span of measurement, or from combining records from neighbouring regions and using the cumulative record length. Both approaches assume that the statistical properties of the hydrologic time series do not change with time, a condition known as stationarity. However, changes through time in the type or frequency of flood-producing storms, or changes in rainfall-runoff generation resulting from land use, are widespread (Hirschboeck 1988). Extending the systematic flood record with palaeoflood information avoids the problem of nonstationarity in the past because palaeoflood indicators record actual rather than hypothetical past floods (Baker *et al*. 2002). Palaeoflood records can also help to constrain the estimate of the probable maximum flood, the largest probable flood that could theoretically occur in a drainage basin. Statistical incorporation of palaeoflood data into systematic data relies on recognition of differences in the two types of data. For example, systematic data may include all floods above a fixed magnitude threshold, whereas the magnitude threshold for palaeoflood data may have varied through time (Blainey *et al*. 2002).

Palaeoflood indicators that record changes in flood frequency through time can also indicate changes in climatic circulation patterns (Redmond *et al*. 2002). And records of the magnitude and frequency of large floods may be used to infer rates of geomorphic change for channels dominated by floods (Wohl 2002) (see FLOOD).

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SEE ALSO: flood

ELLEN E. WOHL

PALAEOHYDROLOGY

Palaeohydrology is the study of past occurrences, distributions and movements of continental waters. It is the highly interdisciplinary linkage of scientific hydrology with the sciences of Earth history and past environments (Schumm 1967). The linkage extends in both directions in that modern hydrological data can be used to create the means of reconstructing past environments (Schumm 1965), while data from past hydrological processes can be used to calibrate and test modern hydrological models (Baker 1998).

The term *palaeohydrology* was first used by Leopold and Miller (1954) in their study of past hydrological conditions associated with a sequence of late Quaternary alluvial terraces in Wyoming. Nevertheless, it is applicable to all elements of the hydrological cycle. Thus, many aspects of cave development in karst aquifers preserve indicators of paleohydrology for those aquifers. Similarly, past changes in lake levels can be documented in terms of a hydrological balance. All these branches of palaeohydrology derive from long traditions in geology and related Earth sciences. For example, Patton (1987) documents the interest by nineteenth and early twentieth-century geologists in past changes in river processes, as evidenced in deposits, terraces and other landforms. Particularly important was the example of Bretz (1923), who discovered the catastrophic flood origin of the Channeled Scabland region in the northwestern United States. Subsequent palaeohydrological quantification (Baker 1973) showed that immense catastrophic flood discharges generated the scabland features during the late Pleistocene bursting of ice-dammed glacial Lake Missoula.

Modes of palaeohydrological inference

There are three general modes of reasoning in palaeohydrology. In one mode general theories of hydrology are used to infer specific effects that can then be discerned in evidence of past hydrological processes. This is the classical deductive mode of rational inquiry. An example would be the problem of the catastrophic flooding associated with the failure of ice-dammed glacial lakes. The palaeohydrologist can use an existing theoretical model for how such a dam fails. Of course, the effective use of this model requires that the correct mode of dam failure be matched with the model (Figure 114). With this condition satisfied, the model may be capable of predicting the hydrograph of the resulting flood. Matching the predicted hydrograph properties to preserved field evidence then constitutes a kind of reconstruction of the past hydrological process.

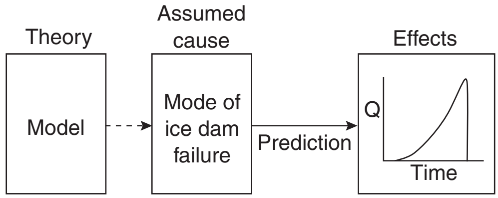

Figure 114 Schematic representation of the deductive mode of palaeohydrological inference applied to the problem of predicting an outburst flood hydrograph from a general theory for the failure of an ice-dammed lake

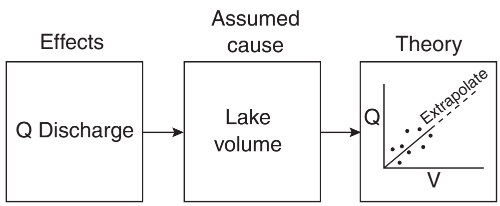

Figure 115 Schematic representation of the inductive mode of palaeohydrological inference applied to the problem of estimating the relationship between lake volume and peak outflow discharge for the failure of ice-dammed glacial lakes

Another common mode of palaeohydrological inference uses empirical relationships that are developed from numerous observations of related hydrological phenomena. This mode of scientific reasoning is inductive. Returning to the problem of ice dam failure, one can collect data on modern glacial lakes. By relating the peak outburst discharges to the associated lake volumes, one can derive an empirical relationship between these two variables (Figure 115). This relationship, extrapolated to the evidence for past lake volumes (or peak discharges) can then be used to estimate the associated discharge (or lake volume). Of course, this exercise must presume that the past phenomena fall in the same class as the data set on modern outburst floods. This is a limitation on all inductive reasoning, because nature is not constrained to behave as we presume it should from our limited set of observation.

Finally, a third mode of reasoning that is used extensively in palaeohydrology is retroductive, or abductive inference (Baker 1996, 1998). For the flood problem, retroductive inference can be accomplished by studying evidence or signs of the

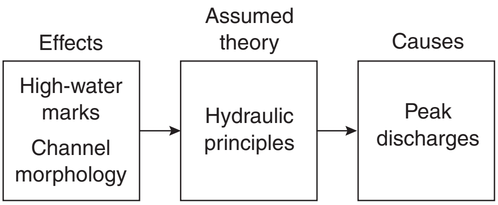

Figure 116 Schematic representation of the retroductive or abductive mode of palaeohydrological inference applied to the problem of estimating peak palaeoflood discharges from various evidence of palaeoflood stages, utilizing hydraulic theory

past floods. These might include the slackwater deposits emplaced marginal to flood channels, or other high-water marks for the past flow stages, as done in *palaeoflood hydrology* (Baker 1987). Then, using a hydraulic model, it is possible to associate the past flood effects with the causative discharges (Figure 116). Thus, retroductive reasoning proceeds from effect to cause, in contrast to the deductive reasoning that proceeds from cause to effect.

Fluvial palaeohydrology

The most basic relationships between river morphology and hydrology involve the supply of water and sediment from upstream of a channel reach of interest. The important dependencies are summarized in the following relationships:

$$Q_W \propto \frac{w, d, \lambda}{S}$$

$$Q_S \propto \frac{w, d, S}{d, P}$$

where Q_W is a measure of the mean annual water discharge and Q_S is a measure of the type of sediment given by the proportion of bedload (usually sand and gravel) to the total sediment load (which may include considerable clay and silt). Q_W and Q_S are the controlling, independent variables. The dependent variables include the channel width w and depth d, the slope or gradient of the channel S, the sinuosity of the reach P, and the meander wavelength λ. For palaeohydrological applications the above relationships are usually quantified by empirical equations, using the inductive approach.

An example of the foregoing reasoning is in regard to the phenomenon of *underfit streams*. These are streams for which some practical measure of the modern river, usually the meander wavelength λ, is too small in relation to the valley that contains the stream. Long recognized as being caused by stream capture, underfitness was also recognized in the context of climatic change by Dury (1954, 1965). Dury (1965) reasoned that, because meander wavelength is directly proportional to bankfull channel width, and because bankfull width is a function of discharge to the 0.5 power (Leopold and Maddock 1953), the wavelength of modern river meanders λ must be proportional to modern bankfull discharges q_b. Applying the same arguments to the enlarged valley meanders of wavelength L formed by ancient discharges Q, Dury (1965) finds

$$Q/q_b = L^2/l^2$$

Dury's study of many rivers in the United States and Europe showed that the ratio L/l varies from 5 to 10, which implies that the ancient discharges Q were 25 to 100 times larger. The immense climatic implications of such large changes led many to question Dury's estimates. In subsequent work it was discovered that the discharges responsible for valley meanders, which are often developed in bedrock, may have very different relationships to channel size than the empirical relationships that apply to modern alluvial rivers.

A significant discovery in fluvial palaeohydrology came when attempts were made to apply Dury's theory to underfit streams on the Riverine Plain of southeastern Australia. The modern Murrumbidgee River is underfit relative to the very large meanders of an ancestral channel. It has a much narrower channel and a much smaller meander wavelength. In addition, there are *prior channels*, which constitute an older system of paleochannels filled with sediment that is much coarser than that conveyed by either the modern Murrumbidgee or by its ancestral stream. Because the prior channels are much wider and have much greater meander wavelengths than the modern river, Dury's theory predicts that they should have experienced much larger bankfull discharges. However, Australian soil scientists insisted that the conditions at the time of prior stream activity were extremely arid. The apparent paradox was resolved when Schumm (1968) showed that the prior channels were formed by relatively high-gradient, low-sinuosity, coarse-sediment-transporting streams. From the proportional expressions above it is clear that the discharge factor relates to parameters other than meander wavelength and channel width, as presumed in Dury's theory. Slope, sediment size, sinuosity and the width-to-depth ratio are all factors, and these combine to produce the result of prior channel development during a drier climatic period.

While much of the foregoing concerned a regime approach to fluvial palaeohydrology, there are other procedures. The sizes of bedload particles moved during past flow events can be related to various measures of the event magnitude, including flow velocity, bed shear stress and power per unit area of bed (Costa 1983). One can also determine the past stages of flow events from a variety of palaeostage indicators, including the study of flood slackwater deposits (Baker 1987). These various techniques have now achieved global application both for the practical study of flood risk assessment, and for the academic study of extreme river processes that defy direct measurement in the field.

Lacustrine palaeohydrology

Ideally, the water balance for a closed basin can be described by the expression

$$dV/dt = d(\mathrm{P_L+R+U})/dt - \mathrm{d(E+O)}/dt$$

where V is the water volume in the lake, t is time, P_L is precipitation input to the lake, R is runoff from the tributary basins that feed the lake, U is subsurface (groundwater) flow into the lake, E is evaporation from the lake, and O is the subsurface flow out of the lake. For any given lake stage, the hydrological balance can be considered in equilibrium, so that

$$dV/dt = 0$$

Subsurface inflow and outflow are generally rather small for many lakes, or they may be very difficult to estimate. By ignoring these factors, the equilibrium water balance equation can be simplified in relation to the area of the lake A_L and the area of the tributary catchment A_C from which water drains into the lake, as follows:

$$A_C\, P_L + A_C\,(P_C k) = A_L\, E_L$$

where P_C is the mean precipitation per unit area over the catchment, k is a runoff coefficient such that $P_C k$ will equal the runoff per unit area from the catchment, P_L is the mean precipitation per unit area over the lake, and E_L is the evaporation per unit area from the lake. Usually only A_L and A_C are known for ancient lakes, leaving a problem in estimating the relative influences of evaporation versus precipitation on the overall lake balance, as follows:

$$A_L/A_C = P_C k/E_L - P_L$$

Note that the area of the lake can expand if the evaporation E_L is reduced, if the runoff from the catchment $P_C k$ increases, if the precipitation over the lake P_L increases, or if some combination of these changes occurs. Because evaporation depends on temperature and other climatic factors, its determination may require some independent means of estimating the past climate. Additional complexities occur for precipitation. Thus, the relative simple appearance of expressions for lake palaeohydrology can be misleading in regard to the problem of actually estimating ancient lake balances.

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SEE ALSO: cave; palaeochannel; palaeoflood; pluvial lake; prior stream; scabland; underfit stream

VICTOR R. BAKER

PALAEOKARST AND RELICT KARST

Palaeokarst refers to KARST landforms that are completely decoupled from the hydrogeochemical system that formed them, as distinct from *relict karst* that is removed from the morphogenetic situation in which it was formed, but remains exposed to and may be modified by present geomorphic processes (Ford and Williams 1989). The terminology associated with palaeokarst can be complex and ambiguous, but a definitive discussion and explanation is provided by Bosak *et al.* (1989). Figure 117 illustrates the main geomorphic relationships encountered.

Palaeokarst is usually found buried unconformably beneath other rocks, the cover beds being younger than the karst. This is sometimes referred to as *buried karst*. When the burial is relatively recent, it tends to be by unconsolidated allochthonous clastic sediments such as alluvial, volcanic, marine or glacial deposits. Relict karst is still subject to modification by modern solution processes beneath the covering sediments and tends to be only partly buried.

Old and deeply buried palaeokarst arises from tectonic subsidence. It can also involve geological deformation. The caprock constitutes a confining formation, and the palaeokarst is interstratified between it and an underlying non-karst formation. This is a form of interstratal karst, but unlike currently active interstratal karst, the palaeokarst is older than the confining cover

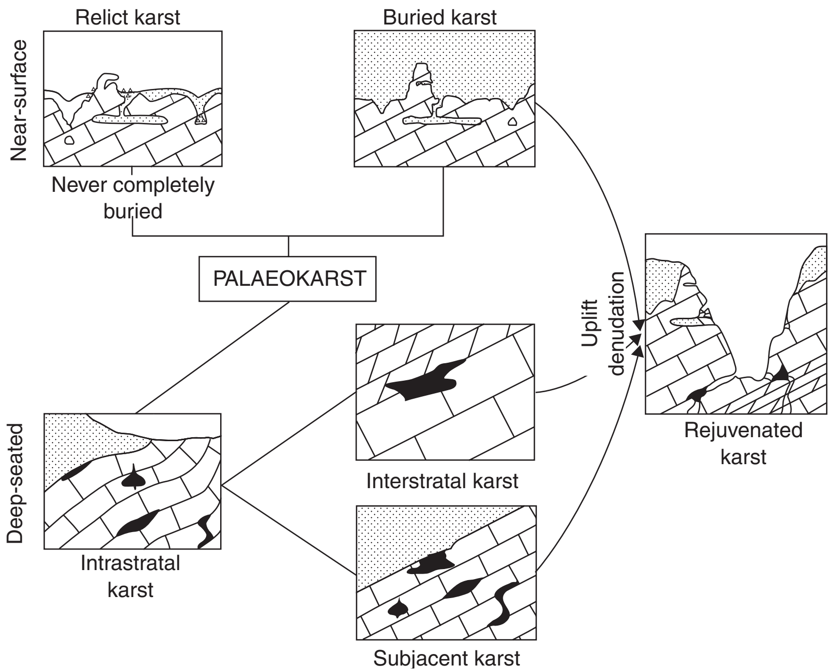

Figure 117 Main geomorphic relationships of palaeokarst

rocks and unconnected to the modern hydrogeological system. There is also an unconformity between the karstified rocks and the caprock. However, sometimes the palaeokarst is quite subtle and without major landforms, and is only recognizable by a disconformity in the carbonate sequence marked by a thin layer of insoluble residue. Such a situation represents a relatively brief interval of subaerial weathering followed by marine transgression.

A distinction is sometimes made between *interstratal* and *intrastratal* karst. The former develops along bedding planes and unconformities, whereas the latter is not restricted to such boundaries between strata. Karst beneath cover beds is sometimes referred to as *subjacent* karst, although if it is currently active it does not constitute palaeokarst. Deeply buried palaeokarsts serve as excellent traps for migrating hydrocarbons and contain some of the world's major oil and gas reserves. Later uplift may result in exposure of the cover beds to erosion and the exhumation of the palaeokarst. When this occurs it can sometimes be reintegrated into the modern hydrogeochemical system and therefore becomes rejuvenated.

Relict karst can arise in two ways: its hydrogeological context may change or its climatic (morphogenetic) situation may alter (Ford and Williams 1989). The first case is commonly found underground as a consequence of the incision of cave streams, because this leads to the de-watering and abandonment of high level cave passages, thus leaving them relict. They are not totally removed from the active hydrogeological system, because they remain in the vadose zone and receive percolation water and accumulate speleothems but, like river terraces in the case of surface rivers, they are removed from the streams that formed them.

The second case results from climate change on a timescale of 10^5 years or more. Climate change associated with major latitudinal shifts of climatic zones has resulted in landforms developed under one morphogenetic system (say humid subtropical) being exposed later to radically different process conditions (perhaps arid, cool temperate or even periglacial). This can arise from global changes to the Earth's climatic system, as experienced in the transition from the Tertiary to the Quaternary, or from continental-scale movement over millions of years arising from plate tectonics, which can result in latitudinal displacement and in wholesale uplift of very large tracts of land (including its karst), such as in Tibet. This leads to the karst being forced out of equilibrium with its process environment. Such landscapes in a different morphogenetic context than the one in which they were developed are sometimes referred to as fossil karst. Shorter term climatic changes, as experienced in glacial–interglacial cycles, can also have profound effects on landscapes, exposing them to polygenetic conditions without necessarily making them relict, but forcing frequent readjustments to new process regimes.

Although most karst is developed by processes associated with the circulation of cool meteoric waters, some is produced by dissolution by hydrothermal waters and some by hot hypogean solutions associated with the intrusion of magma bodies. Deep subjacent karst is formed where heated water is circulated in a confined aquifer. These karsts are often encountered during mineral exploration, because the cavities produced are often heavily mineralized (Bosak 1989) frequently with sulphide ores. When removed from the situation in which they were produced (which was often at depth and many millions of years ago), such karsts constitute *hypogene palaeokarst*.

Palaeokarst is widespread throughout the world and occurs in carbonate rocks to at least Cambrian age. Contributors to the book edited by Bosak (1989) provide the best international review currently available.

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PAUL W. WILLIAMS

PALAEOSOL

A palaeosol is a soil that formed on a landscape of the past (Retallack 2001). Soils are products of the physical, chemical and/or biologic weathering of sediments and rocks (see SOIL GEOMORPHOLOGY). Palaeosols typically occur at (1) major unconformities and (2) within basin-fill deposits representing aggradational systems. Although alluvial palaeosols are probably the most common type of palaeosol, they also occur in palustrine, aeolian, deltaic and coastal sediments, as well as in carbonate deposits (Kraus 1999). Palaeosols are especially abundant in Quaternary deposits and have been identified in rocks as old as 3.5 billion years (Retallack 2001).

Palaeosols are identified by a wide range of features including root traces, burrow fills, mottles, nodules, peds, clay films, cemented horizons (i.e. CALCRETE, SILCRETE, FERRICRETE), slickensides and matrix microfabrics. Concentrations of these features are used to identify palaeosol horizons, and individual profiles consist of vertically stacked horizons. Palaeosols should show vertical and lateral variations that mimic those observed in modern soils (see CATENA). One major difficulty in recognizing palaeosols is the effect of burial diagenesis. Diagenetic processes such as compaction, cementation and mineral transformations can significantly alter the texture, mineralogy and chemistry of palaeosols.

Palaeosols provide important records of past environments. Palaeosols at major unconformities are used to interpret past climates and changes in base level, and serve as important lithologic markers for correlating sedimentary deposits. In thick successions of sedimentary rocks, alluvial paleosols record the mode and tempo of basin filling (Kraus 1999). Weakly developed palaeosols are associated with rapid sediment accumulation rates and form close to ancient channel systems. In contrast, well-developed palaeosols reflect slow sediment accumulation rates and settings where sediment input is negligible. Rapid subsidence and sedimentation produce vertically stacked profiles whereas cumulative profiles reflect slow but steady rates of concurrent sedimentation and pedogenesis. Palaeosols also provide opportunities to study landscape development at a variety of spatial scales. At local scales, palaeosol properties vary according to changes in grain size and topography. At more regional scales, palaeosols reflect differences in climate, topography, tectonic setting and lithology.

One of the most promising applications of palaeosol research involves paleoclimate studies. Field-based and stable-isotope studies of iron- and carbonate-rich palaeosols have been used to document increases and decreases in atmospheric oxygen and carbon dioxide concentrations, patterns of global cooling and warming, and ancient mean annual temperature and precipitation (Retallack 2001). Mass balance studies of palaeosols have been used to quantify chemical weathering trends and ancient floodplain hydrology. Finally, palaeosols contribute to the record of ecosystem evolution. The colonization of land by plants and the development of forest and grassland ecosystems are recorded by the development of new palaeosol morphologies such as the mollic horizon (Retallack 2001).

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SEE ALSO: calcrete; catena; chemical weathering; chronosequence; climatic geomorphology; duricrust; ferricrete; silcrete; soil geomorphology

ANDRES ASLAN

PALI RIDGE

A Hawaiian term for a steep slope or large cliff. Palis are steep-faced scarp ridges between stream valleys, commonly composed of basalt, and typically over 1,000 m in height. Various mechanisms have been suggested for their origin, such as being the eroded wall of a dissected shield volcano, being shaped by higher past sea levels, extreme fluvial downcutting, and catastrophic landslides. It is likely that several of these processes are involved in the formation of a pali.

Further reading

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STEVE WARD

PALSA

Palsas are small mounds of peat rising out of mires in the subarctic region characteristic to the discontinuous circumpolar permafrost zone provided that the peat layer is thick enough. They contain a permanently frozen core of peat and/or silt, small ice crystals and thin layers of segregated ice, which can survive the heat of summers. An insulating peat layer is important for preserving the frozen core during the summer. The peat should be dry during the summer, thus having a very low thermoconductivity, and wet in autumn, when freezing starts, giving a much higher thermoconductivity. This allows the cold to penetrate so deep into the peat layers that they do not thaw during the summer.

Palsas can be classified according to their morphology: dome-shaped, elongated string-form, longitudinal ridge-form, and extensive plateaux palsas as well as palsa complexes with many basins, hollows and ponds of thermokarst origin (Plate 85). The diameter of dome-shaped palsas ranges from 10 to 150 m and the heights from 0.5 up to 12 m. Longitudinal ridge-form palsas could be up to 0.5 km long and 6 m in height. Palsa plateaux rise 1–1.5 m above the surface of the surrounding peat surface and can cover areas of several square km.

Once a palsa hummock rises above the mire surface peat formation on its top ceases almost entirely. The surface peat on an old palsa is produced mainly by *Bryales* mosses, lichens and *Ericales* shrubs. It could also be by wind eroded old moss peat. Below the dry surface peat is the original mire peat formed by *Sphagnum, Carex* and *Eriophorum* remains. It is normally permanently frozen forming the permafrost core. In Finnish Lapland the summer thawing forms only a 50 to 60-cm thick active layer on the palsa surface. On the southern slopes of palsas the active layer gets deeper and on the edges the permafrost table is almost vertical. To date a palsa formation, samples should be collected from the contact of normal mire peat and of the dry peat formed on the palsa after its formation.

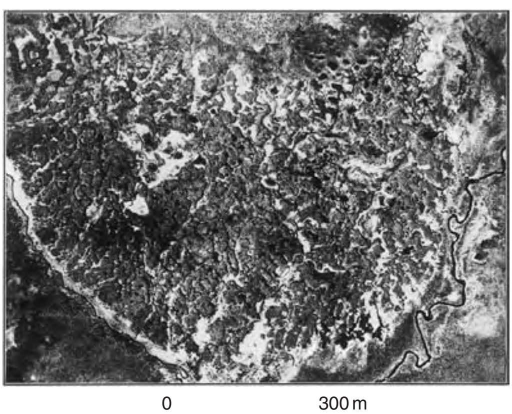

Plate 85 Aerial photograph (nr. 8634 17) of Linkinjeäggi palsa mire, Utsjoki, Finland. Published with the permission of Topografikunta

Low air temperatures together with low precipitation and a thin snow cover are found to be the most prominent factors for palsa formation. The hypothesis that palsas are formed in places with thin snow cover has been proved experimentally by cleaning the snow off the mire surface several times during three winters; a permafrost layer formed in the peat and a man-made small palsa.

Wind drift controls the thickness of snow cover on the mire surface. Thin snow cover allows the frost to penetrate deep into the peat, and in these places the frost fails to disappear completely during the seasonal thawing and part of it remains under the insulating peat. In the following winters the unthawed layer of frost becomes thicker and the mound starts to rise. The wind then carries away snow from the exposed hump more easily and the freezing process accelerates. The freezing front sucks moisture and segregated ice lenses are formed in the frozen core. This process increases the water content of the frozen core which can be 80–90 per cent of the volume.

The concept of cyclic palsa development is based on field observations and experimental studies in Finnish Lapland (Figure 118):

(A, B) The formation of a palsa begins when snow cover is locally so thin that winter frost penetrates sufficiently deeply to prevent summer heat from thawing it completely. The surface of the bog is then raised somewhat by frost processes.

(C) During succeeding winters the frost penetrates still deeper, the process of formation accelerates and the hump shows further upheaving due to freezing of pore water and ice segregation. As the surface rises, the wind becomes ever more effective in drying the surface peat and keeping it clear of snow.

(D, E) When the freezing of the palsa core reaches the till or silt layers at the base of the mire then the mature stage of palsa development begins. By this time the palsa stands well above the surface of the mire, typically displaying a relief of about 7 m in western Finnish Lapland.

(F) Degradation now starts, and peat blocks from the edges of the palsa collapse along open cracks into the pools which often surround the hummocks. During later stages,

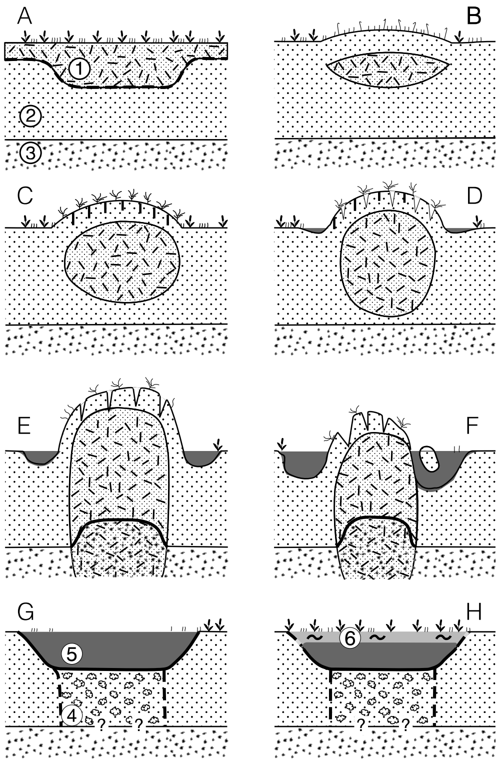

Figure 118 A general model of the formation of the frozen core (1) of a palsa in a mire (2) with a silty till substratum (3). A: the beginning of the thaw season. B: the end of the first thaw season. C: embryo palsa. D: young palsa. E: mature palsa. F: old collapsing palsa surrounded by a large water body. G: fully thawed palsa giving a circular pond on the mire (5). The thawed peat is decomposed (4). H: new peat (6) formation starts in the pond (Seppälä 1982, 1986, 1988)

the vegetation may be removed so that the palsa surface is exposed to deflation and rain erosion.

(G) Old palsas are partially destroyed by thermakarst, and become scarred by pits and collapse forms. Dead palsas are unfrozen remnants: either low (0.5 to 2 m high) circular rim ridges; or rounded open ponds and pond groups; or open peat surfaces without vegetation.

(H) From such pools a new palsa may ultimately emerge after a renewed phase of peat formation, and the cycle of palsa development recommences from the beginning.

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MATTI SEPPÄLÄ

PAN

Also called *playas, pfannen, sabkhas, chotts, kavirs*, etc. are closed topographic depressions that are features of low-angle surfaces in the world's drylands (Jaeger 1939). Their characteristic morphology has often been likened to a clam, a heart or a pork chop. They are especially well developed on the High Plains of the USA, in the Argentinian Pampas, Manchuria, the West Siberian steppes and Kazakhstan, western and southern Australia and the interior of southern Africa (Goudie and Wells 1995). Pans evolve on susceptible surfaces. In southern Africa, for example, they are best developed on the sandy Kalahari Beds and on fine-grained Ecca shales. They also occur in particular topographic situations – deflated lake floors, old drainage lines, in interdune swales, in the noses of parabolic dunes and on coastal plains (e.g. the Carolina Bays of the eastern seaboard of the USA). They are sometimes, though by no means invariably, associated with LUNETTEs (Sabin and Holliday 1995). They are often oriented with respect to regional wind trends, and tend in many cases to have bulbous lee sides. In areas like the Pampas, the High Plains of the USA and the interior of South Africa there are literally tens of thousands of pans, and they may cover as much as a quarter of the ground surface.

The origin of pans has intrigued geomorphologists for over a century. Hypotheses have included deflation, excavation by animals, and karstic (see DAYAS) and pseudo-karstic solution. Arguments on this issue are recurrent and recent years have seen some important contributions to the debate (e.g. Gustavson *et al.* 1995). What is becoming clear is that a range of processes has been involved in the initiation and maintenance of pans and that no one hypothesis can explain all facets of their own long histories and their variable sizes and morphologies.

An integrated model of pan development is as follows (Goudie 1999). First, pans occur preferentially in areas of relatively low effective precipitation. This predisposing condition of low precipitation means that vegetation cover is sparse and that deflational activity can occur. Moreover, once a small initial depression has formed, and the water in it has evaporated to give a saline environment, the growth of vegetation is further retarded. This further encourages deflation. The role of deflation in the removal of material from a depression may be augmented by animals, who tend to concentrate at pans because of the availability of water, salt licks and a lack of cover for predators. Trampling and overgrazing expose the soil to deflation and the animals would also physically remove material on their skins and in their bladders. Aridity also promotes salt accumulation so that salt weathering could attack the bedrock in which the pan might be located. It is also important that any initial depression, once formed and by whatever means, should not be obliterated by the action of integrated or effective fluvial systems. Among the factors that can cause a lack of fluvial integration are low angle slopes, episodic desiccation and dune encroachment, the presence of dolerite intrusions and tectonic disturbance. This model of pan formation is similar to that developed for the USA High Plains by Gustavson *et al.* (1995).

In addition to their occurrence in deserts, various types of oriented lake are also a feature of some tundra areas (Carson and Hussey 1962).

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A.S. GOUDIE

PARAGLACIAL

Paraglacial is a term that was introduced by June Ryder (1971) to describe alluvial fans in the interior of British Columbia that had accumulated through the reworking of glacial sediment by rivers and debris flows following late Wisconsinan deglaciation. She mapped alluvial fan distribution throughout south-central British Columbia, noted that they were essentially inactive at present and concluded that they must have been dependent on the reworking of till, glacifluvial and glacilacustrine deposits by streams and debris flows in the earliest Holocene. She showed that fan accumulation was initiated soon after valley floors became ice free and continued until after the deposition of Mazama tephra (*c.*6,000 yrs BP).

The paraglacial concept was formalized by Church and Ryder (1972). They defined paraglacial as non-glacial processes that are directly conditioned by glaciation and added that 'it refers both to proglacial processes and to those occurring around and within the margins of a former glacier that are the direct result of the former presence of ice'. In this remarkable paper, they synthesized evidence from their field areas in

British Columbia (Ryder) and Baffin Island (Church) and they used the contemporary Baffin Island environment as an analogue for the early Holocene environment in south-central British Columbia. They concluded that, although fluvial sediment transport rates were likely to be greatest immediately after deglaciation, fluvial reworking of glacigenic sediments was likely to continue as long as such sediment was accessible to rivers. They identified three aspects of the influence of paraglacial sediment supply on fluvial transport: (a) the dominant component of reworked sediment may shift from till to secondary sources, such as alluvial fans and valley fills; (b) regional uplift will condition the timing of changes in the balance between fluvial deposition and erosion such that the cascade of sediment evacuation can be interrupted by sediment deposition; and (c) consequently, the total period of paraglacial effect is prolonged beyond the period of initial reworking of glacigenic sediments.

Slaymaker (1977) and Slaymaker and McPherson (1977) noted that in British Columbia primary upland denudation rates are low and that a large component of contemporary sediment load is derived from secondary remobilization of late Pleistocene and Holocene deposits. Slaymaker (1987) also showed that in the British Columbia and Yukon region, medium-scale (100–10,000 km^2) river systems exhibit the highest specific sediment yield, in contradiction to the conventional model of sediment yield vs basin area relations. Church *et al*. (1999) confirmed this result.

Church and Slaymaker (1989) emphasized the generality of the definition, specifying that it is applicable to all periods of glacier retreat and that a paraglacial period is not restricted to the closing phases of glaciation but may extend well into the ensuing non-glacial interval. The essence of the concept is that recently deglaciated terrain is often initially in an unstable or metastable state and thus vulnerable to rapid modification by subaerial agents. Effectively then the 'paraglacial period' is the period of readjustment from a glacial to a non-glacial condition. Different elements of paraglacial systems adjust at different rates from steep, sediment-mantled hillslopes (a few centuries) to large fluvial systems (>10,000 years). Increase in specific sediment yield with basin area for basins smaller than *c*.30,000 km^2 shows that specific sediment yield equals area raised to the power of 0.6. Isometry would dictate an exponent of 0.5 (because specific sediment yield can be reduced to a length dimension); hence sediment is recruited to streams at a rate greater than expected simply from an increase in area. The additional sediment is derived from erosion of both *in situ* and reworked glacigenic deposits along riverbanks. Effectively, these rivers are degrading through valley fill deposits forming entrenched trunk streams flanked by Holocene terraces. For basins greater than 30,000 km^2 specific sediment yield tends to decline as conventional models predict because non-alluvial riverbanks are protected from erosion.

These data demonstrate that the timescale of paraglacial sediment reworking in British Columbia includes the whole of Holocene time. Church and Slaymaker (1989) estimate that ultimate dissipation could take several tens of thousands of years. This implies that interglacial fluvial systems were still relaxing from the previous glacial period when the succeeding glacial period arrived. They also imply that there is no equilibrium between hydro-climate, denudation rate and sediment yield because all 'fluvial sediment yields at all scales above *c*.1 km^2 remain a consequence of the glacial events of the Quaternary'.

Owens and Slaymaker (1992) have examined sediment accumulation rates in three small lake-drained basins of less than 1 km^2 over the last 6,000 years and confirmed that these rates are 1–2 orders of magnitude lower than those of larger basins. Souch (1994) has traced the paraglacial signal downstream through a system of lakes progressively further from the glacial sources. Church *et al*. (1999) have expanded the analysis of suspended sediment yields across Canada and described seven Canadian regions with adequately monitored sediment data. Five of these regions were shown to have trends comparable with those of British Columbia; one, southern Ontario, is influenced by intensive land use disturbance and the data show no trend; one, the eastern Prairies, is a region of net fluvial aggradation and specific sediment yields decrease with basin area in accordance with conventional models. Evidently, paraglacial effects persist throughout the majority of Canada's regions.

Ballantyne (2002), in a magisterial summary and extension of the paraglacial concepts developed in British Columbia, points out that between 1971 and 1985, the paraglacial concept was largely ignored outside North America. Since 1985, he sees four trends: (a) an extension in the geomorphic contexts in which the paraglacial concept has been explicitly used; (b) a focusing of research on present-day paraglacial processes and land systems; (c) use of the paraglacial concept as a framework for research across a wide range of contrasting deglacial environments; and (d) a growing awareness of the palaeo-environmental significance of paraglacial facies in Quaternary stratigraphic facies. The working definition that he adopts for 'paraglacial' is 'non-glacial Earth-surface processes, sediment accumulations, landforms, land systems and landscapes that are directly conditioned by glaciation and deglaciation'.

The new perspective given by Ballantyne (2002) is most remarkable in its overview of the wide range of geomorphic contexts in which the paraglacial concept is already explicitly being used. These contexts are, in addition to the original debris cone, alluvial fan and valley fill deposits: (a) rock slopes; (b) sediment-mantled slopes; (c) glacier forefields; (d) glacilacustrine systems; and (e) coastal systems.

Wyrwoll (1977) was the first to identify rock slope response in a paraglacial context. Ice downwasting and retreat has resulted in the debuttressing of rockslopes and yields three responses: large-scale catastrophic rock slope failure; large-scale progressive rock mass deformation and discrete rock fall events.

The work of Ballantyne and Benn (1994) is significant in identifying sediment-mantled slopes in a paraglacial context. They note the processes of reworking sediment-mantled slopes yielding intersecting gullies, coalescing slope foot debris cones and valley floor deposits of reworked drift. Gully erosion and debris flow activity are the most obvious paraglacial processes invoked in this environment.

Matthews (1992) is credited with the first explicit identification of glacier forefields (forelands) in a paraglacial context. Effects conditioned by the former presence of a glacier include unconsolidated diamicton, steep slopes, unvegetated surfaces and the acceleration of mass movement, frost action, fluvial and aeolian processes.

Leonard (1985) was one of the early investigators of the paraglacial response of lake sediments. Such work accelerated in the 1990s and is now one of the most commonly used ways of assessing the changing rates of sediment production during the Holocene, specifically estimating the duration of the paraglacial effect in specific lake-drained basins.

The extension of the paraglacial concept to coastal systems is perhaps the most dramatic extension of the concept. Forbes and Syvitski (1994) defined paraglacial coasts as 'those on or adjacent to formerly glaciated terrain, where glacially excavated landforms or glacigenic sediments have a recognizable influence on the character and evolution of the coast and nearshore deposits'. They specifically exclude the effects of glacio-isostatic rebound and glacio-eustatic sea-level change on the grounds that these effects are more widely or even globally distributed.

It is clear from Ballantyne's discussion that the paraglacial concept has even wider significance than had previously been imagined. The data bring into question the possibility of any equilibrium or balanced condition in landscapes that have undergone Quaternary glaciation.

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SEE ALSO: alluvial fan; glacifluvial; glacilacustrine

OLAV SLAYMAKER

PARALIC

Term referring to environments by the sea where shallow waters predominate, though nonmarine. Paralic environments are particularly associated with intertongued marine and continental deposits situated on the landward side of a coast. This includes lagoonal, littoral, alluvial and shallow neritic environments. Paralic sedimentation incorporates basins, swamps (paralic swamps), deltaic zones, heavily alluviated shelves and platform marshes. The word paralic is derived from the Greek word *paralia* meaning sea coast.

Paralic environments typically exhibit localized, abruptly changing facies tracts with a large variety of lithologies. The deposits are distinct by their thick terrigenous accumulations of clays, sands and silts (orthoquartzite to subgreywacke), intimately mixed with estuarine, marine and continental deposits. Paralic sediments can offer important stratigraphical information concerning the long-term changing coastal environment. Often the deposits are zones of subsequent coal formation (termed paralic coal), while petroleum accumulation is frequent within paralic basins. Paralic ecosystems are characterized by a large spectrum of biological species that are strictly bound to the particular environment, and are able to remain stable despite changing environmental conditions (Guelorget and Perthuisot 1992).

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STEVE WARD

PARNA

A deposit of dust (suspended wind-blown mineral material) differentiated from LOESS by its higher clay content. The term was coined for deposits found in the interior of southeastern Australia and is attributed to an aboriginal word meaning 'sandy and dusty ground' (Butler 1956: 147). Parna can be regarded as synonymous with desert loess. High clay content (30–70 per cent) lead to differentiation of parna from the glacial loess of Europe, but is a feature of desert loess in Africa and elsewhere. High clay content, and particularly the inferred or observed presence of clay in the form of detrital aggregates, remains the main criterion for recognition of parna although the quartz fine sand or silt component is the most readily recognizable feature. These, and other, properties are inferred to arise from the origin of parna from the deflation of soils which have already experienced considerable weathering. Thus other deposits of clay-rich aeolian sediment, derived from deflation of lakes (see LUNETTE), for example, are not considered to be parna, although there is inconsistency in the application of the term. Other properties of parna, such as colour, calcium carbonate, salts, texture and structure vary with soil drainage in a catenary relationship. The depth and number of parna layers are variable and relate strongly to local topography and post-depositional erosion as well as proximity to the source areas. Parna layers were deposited during arid climate phases of the late Quaternary.

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PAUL HESSE

PASSIVE MARGIN

In plate tectonic theory, oceans are spreading from mid-ocean ridges, and being consumed by subduction at active margins. Passive continental margins are those that are not also edges of plates. They are also known as 'trailing edges' and 'Atlantic-type margins'.

They are presumed to be initiated as rift valleys, and when the rifts turn into oceans by seafloor spreading they become continental margins. The new margins may undergo some changes, but may also inherit landforms from pre-breakup times. In contrast to active margins which have many volcanoes, volcanicity is rare in passive margins: only east Australia has abundant volcanoes. In India the vast flows of the Deccan traps accompanied creation of the passive margin.

Based on morphotectonics there are two main types of passive margin: (1) passive margins without significant vertical deformation; and (2) passive margins with a marginal swell and Great Escarpment. We have no good explanation for the two types, or their distribution. Why does eastern Australia have a marginal swell-type margin, but most of the south coast is without vertical deformation? Why does most of southern Africa have Great Escarpments, while East Africa does not?

Passive margins without significant vertical deformation are formed by simple pull-apart of a continent. The Red Sea is an example of the early stage of the process. The Great Australian Bight exemplifies a later stage. Horizontal Tertiary limestones underlie the flat Nullarbor Plain, which is almost an old seafloor. In Patagonia (Argentina) the Atlantic is bordered by an extensive plain cut across ancient rocks. The offshore zone is characterized by many listric faults.

Margins with a marginal swell are the dominant type of passive margin (Ollier 2003), and include the Drakensberg, the Western Ghats, the Appalachians, parts of Greenland, Brazil, Antarctica, and elsewhere. The basic geomorphology of such margins is shown in Figure 119.

Plateau are upland areas with relatively flat topography and most are erosion surfaces. They may be extensive or dissected until only fragments are left. They occur on a wide range of rock types including horizontal strata, metamorphic rocks, granite and massive lava flow sequences.

The *marginal swell* is a widespread swell or bulge along the edges of a continent (*Randschwellen* in German; *bourrelets marginaux* in French). The whole land surface has been warped into an asymmetrical bulge, with the steeper slope

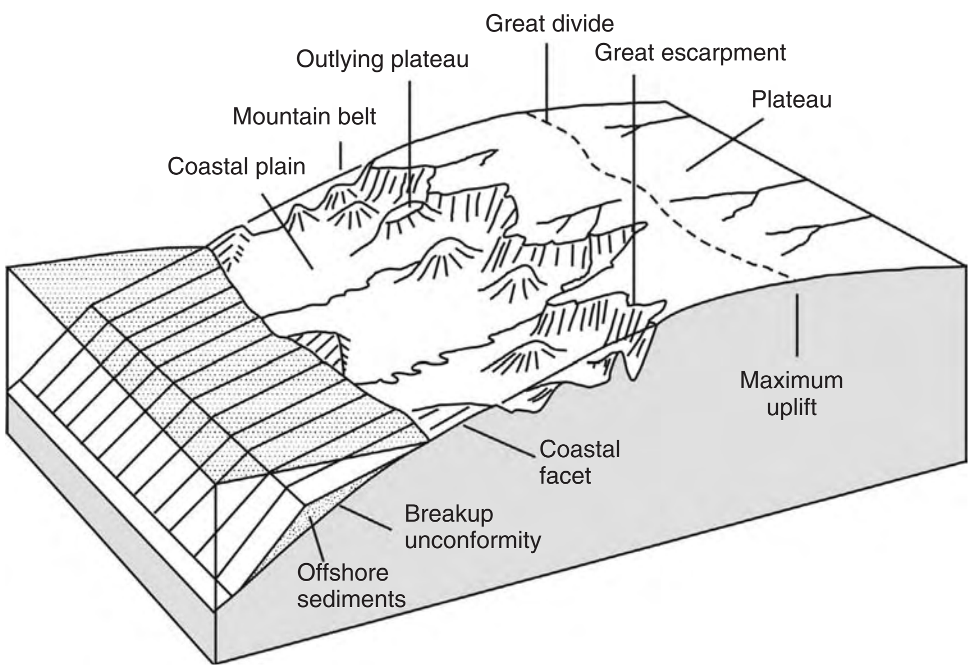

Figure 119 Morphotectonic features of a passive continental margin with marginal swell

to the coast (though the 'steep' slope is still only a few degrees). The marginal swell formed after the planation surface of the plateau, and after formation of major valleys.

Great Escarpments are scarps hundreds or thousands of kilometres long, and up to a thousand metres high. They occur on all sorts of rocks and are not structurally controlled. This is demonstrated especially in the Western Ghats of India, where the Great Escarpment in the north is cut across horizontal basalt, and continues with no change of form into the Precambrian gneisses and granites of southern India. Great Escarpments run roughly parallel to the coast, and they separate a high plateau from a coastal plain. The top of the Great Escarpment can be very abrupt. They are undoubtedly erosional. In places they are rather straight, but elsewhere can be highly convoluted. In some instances the top of the escarpment is the drainage divide between coastal and inland drainage (Brazil, Namibia); in other places the major drainage divide of the continental margin may be hundreds of kilometres inland of the Great Escarpment (east Australia). Many large waterfalls are found where rivers cross the Great Escarpment.

Mountainous areas, often quite rugged, form below the Great Escarpment, where the old plateau surface has been largely destroyed. Occasionally a patch of plateau is isolated to form a peninsula or isolated tableland.

Coastal plains lie between the mountains and the sea.

The landforms on such margins depend to some extent on present and past climates. In southern Norway the landscape is dominated by fjords and glaciated valleys, but the major features of plateau and Great Escarpment were present earlier. Glaciation straightened and deepened the valleys, but they originated before glaciation (Lidmar-Bergström *et al*. 2000). Greenland and Antarctica have some marginal swell-type margins that have been much modified but not obliterated by glaciation. In marked contrast is the Great Escarpment of Namibia, created by fluvial erosion but now in a largely desert environment.

A wedge of sediments is deposited offshore from the continental margin above an unconformity called the breakup unconformity (meaning related to the breakup of a supercontinent). The sediments record the history of uplift in their hinterland. In Scandinavia and southern Africa interpretation of the offshore sediments suggests that there were two main uplift phases – Palaeogene and Neogene. Individual river sources of sediment may be indicated: a delta developed after 103 Ma near the mouth of the present Orange River, South Africa.

There are two main models of passive margin evolution. One school, placing emphasis on fission track data and similar methods, believes there was a continuing uplift towards the margin, where the continental rim ended at a massive fault. Slope retreat moved from the initial faulted margin to the present Great Escarpment. The alternative is warping of the palaeoplain to below sea level. Valleys eroding the steeper, coastal side coalesced to form a Great Escarpment, which then retreats. This model equates the breakup unconformity with the plateau surface, and equates the marginal swell with the raised shoulder of present-day rift valleys (e.g. the Lake Albert rift, Uganda). This implies that the marginal swell dates back to the earliest days of continental breakup.

Some passive margins have simple drainage patterns with streams flowing in opposite directions away from a ridge at the top of the Great Escarpment (Brazil, Western Ghats). On some marginal swells major rivers were in existence before continental breakup and can still be traced in the modern landscape (Australia, South Africa). Drainage may be modified and even reversed. Original drainage divides may relate to the original tectonic movement that made a marginal swell, with the crest of the swell being the drainage divide. The location of divides can be modified by drainage evolution, especially headward erosion of coastward flowing rivers.

What caused uplift of the marginal swell is not known, though a wide range of proposals have been made, and it may not even be the correct question to ask. Some passive continental margins might have been high originally, like the high plateau bounding many present-day rift valleys. A secondary mechanism is isostatic adjustment to erosion of the land, loading by offshore deposition and (in places like Greenland and Antarctica) loading by ice sheets.

The geomorphic evolution of some margins has been traced back to the Mesozoic, as in eastern Australia. Elsewhere landscape evolution is thought to be Miocene and younger, as in the Piedmont of the USA. There may be more than one period of movement. Several passive margins are now thought to have Mesozoic beginnings modified by further movement in the Neogene.

Several passive margins do not fall into the two groups outlined above. Some margins are dominated by deposition of sediment as the crust sinks. The Gulf Coast of the United States, and the coastal part of the Chaco-Pampean Plain in Argentina are examples. The country east of Perth in Western Australia is dominated by the north–south Darling Fault. This makes a topographic fault scarp, bounding an erosion surface cut across Precambrian rocks. West of the fault, the Perth Basin has about 11 km of Silurian to Cretaceous sediments showing long-continued downfaulting. The basin is not a rift valley as no fault further west has been located, so this is a faulted passive margin. The southern coast of Western Australia, west of the Great Australian Bight, exhibits a simple warp, bending the West Australian planation surface to sea level. Ancient valleys that flowed across the land from Antarctica before breakup can be traced across the warp as lines of salt lakes, and the drainage is reversed to form the present south flowing rivers (Clarke 1994). Despite the long time available (Antarctica started to separate about 55 Ma) there is no sign of initiation of a Great Escarpment, suggesting that it requires a greater relief than this margin offers.

Considering the relationship of big rivers, deltas and global tectonics, Potter (1978) pointed out that the twenty-eight biggest rivers in the world all drain to passive margins. Twenty-five of the world's largest deltas are also found on passive margins.

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SEE ALSO: active margin; isostasy

CLIFF OLLIER

PATERNOSTER LAKE

A body of water set in a formerly glaciated environment, often divided by either moraine deposits and/or rock bars, and aligned with similar neighbouring lakes. They are often linked by a stream, rapids or waterfalls running through the valley so that it resembles, when viewed in plan, a string of beads. The term paternoster lake is derived from this pattern, with each lake resembling a paternoster (bead) in a rosary. Paternoster lakes are formed by the plucking and scouring out of a valley bed by a glacier, though they may also form through the damming effects of glacial deposits (by moraines, rock bars or by riegels). The varying rock resistance means that the glacier will erode away the weaker rock more quickly, forming depressions in the valley floor. Water then accumulates in these depressions upon glacial retreat, leaving a series of usually elongated lakes, reflecting the direction of scour from which they developed. The number, size and shape of paternoster lakes varies as a function of weakness, jointing and lithology of the underlying rock, alongside the varying characteristics of the glacier, valley steepness, and extending or compressive flow. Paternoster lakes are common in Sweden, draining into the Baltic. Also, Llyn Dinas and Llyn Gwynant, Snowdonia (UK), are examples of paternoster lakes.

STEVE WARD

PATTERNED GROUND

Patterned ground consists of a range of phenomena – circles, nets, polygons, steps and stripes – developed in surface materials. Such phenomena occur in a wide range of environments and have a large number of causes.

Particularly in the seasonal tropics, swelling clay and texture-contrast soils develop micro-relief consisting of mounds and depressions arranged in random to ordered patterns. These are normally termed GILGAI. Most mechanisms of gilgai development involve swelling and shrinking of clay subsoils under a severe seasonal climate.

In some arid and periglacial regions patterned ground is in the form of DESICCATION CRACKS AND POLYGONS. These result because of the volume reduction that takes place in fine-grained, cohesive sediments as they dry out by evaporation of water. This creates sufficient tensional stress for rupture to occur and cracks to be formed.

Elsewhere in dry regions, patterned ground can be associated with the presence of salt, particularly on the floors of playas and on SABKHAS (Hunt and Washburn 1960). Thrusting structures can develop which are called tepees, because of their resemblance to the shape of the hide dwellings of early American Indians (Warren 1983).

In other dryland regions patterns can be produced by vegetation banding. From the air many dryland surfaces can be seen to be characterized by alternating light and dark bands called BROUSSE TIGRÉE (tiger bush). The banding reflects differences in the proportions of grasses and shrubs. This in turn is related to the action of sheet flow on low angle surfaces (0.2–2 per cent) in areas with 50–750 mm mean annual rainfall (Mabbutt and Fanning 1987).

Organic processes also create patterns through the building of mounds by such organisms as communal rodents and termites (see TERMITES AND TERMITARIA). In the case of the MIMA MOUNDS of the USA and the Heuweltjies of southern Africa their mode of origin is uncertain (Reider *et al.* 1996).

However, patterned ground (Plate 86) is especially prevalent in periglacial regions (see PERIGLACIAL GEOMORPHOLOGY; ICE WEDGE AND RELATED STRUCTURES) and in areas underlain by PERMAFROST. A great diversity of forms and processes are involved (Washburn 1956), including thermal contraction cracking, seasonal frost cracking and desiccation cracking. Circular forms are produced by FROST HEAVE (cryoturbation). Periglacial areas also show the development of Earth hummocks (Thufur) (Schunke and Zoltai 1988), and cryoturbation plays a role in their formation. Relict periglacial patterned ground phenomena developed during former cold phases are widespread in mid-latitude areas (Boardman 1987).

Plate 86 Late Pleistocene patterned ground developed under periglacial conditions in the Thetford region of eastern England. The stripes, which have analogues in present-day Alaska, are formed by alternations of heather (*Calluna vulgaris*) and grass

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A.S. GOUDIE

PEAT EROSION

Peatlands cover large tracts of the microthermal northern hemisphere, in countries like Canada, Russia and Finland, but mostly these are low-lying and the peat remains largely intact. Upland blanket mire is much more rare and because of higher rainfall and greater slope angles, erosion is more likely. Some 8 per cent of the land surface of the British Isles is covered by blanket peat, mainly in the north and west. These blanket bogs form the largest single contribution (10–15 per cent) to a globally scarce resource. These areas of blanket peat are important for many reasons: water catchments, hill farming, shooting, recreation and landscape.

The blanket peat of the southern Pennines is unquestionably the most degraded in Britain with gullying affecting three-quarters of the blanket peat. These peatlands lie close to large urban areas (Manchester, Leeds, Sheffield), important sources of air pollution, and, compared to other areas of blanket bog, are climatically marginal, being more southerly than most and in areas receiving barely 1500 mm annual rainfall (Tallis 1997). Peat erosion has been studied over the last century, but it was largely Margaret Bower (1960, 1961) who stimulated recent work. She identified two types of gully system: a dense network of freely and intricately branching gullies on very flat ground (less than 3°); and, more linear gullies with much less tendency to branch on sloping land. Erosion rates from the heavily gullied blanket peat are high for the UK. Labadz *et al.* (1991) used reservoir sedimentation surveys to establish the long-term sediment yield: in total over 200 t $km^{-2} yr^{-1}$ including an organic fraction of almost 40 t $km^{-2} yr^{-1}$. These high sediment yields mean that many of the small reservoirs built in the nineteenth century are now largely full up with sediment and effectively useless for water storage. Whilst the peat erosion rates seem relatively small, given the low bulk density of peat, these do in fact represent large volumetric losses, implying that most of the gullies may have developed during the past three centuries.

John Tallis, in particular, has studied the history of peat erosion in the southern Pennines. His analysis of pollen profiles shows that there was drying out of the mire surface during the 'Early Mediaeval Warm Period' in the twelfth and thirteenth centuries. This was followed by a cooler, wetter period, and it seems possible therefore that climate change could have triggered gully erosion at that time (and perhaps in earlier dry phases too). More recently, human-induced pressures on the blanket peat have probably been more important, sometimes working in tandem with climate change. Fire (accidental or deliberate) and overgrazing by sheep are the most important direct pressures leading to erosion; the loss of pollution-sensitive mosses, particularly *Sphagnum*, is also likely to have been significant. Complete loss of *Sphagnum* soon after the start of the Industrial Revolution in the eighteenth century may well have initiated more widespread gully erosion than might have developed because of climatic change alone.

In intact peat, the water table remains close to the surface except during severe droughts. Most storm runoff is produced by saturation-excess overland flow therefore, although locally pipeflow may be important (Holden and Burt 2002). On flat ground a hummock and pool micro-topography often develops. If the peat dries out, gullies begin to form, further lowering the water table, especially during summer. As the peat re-wets in autumn, there is an increased tendency for leaching of dissolved organic carbon (DOC) discolouring local water supplies and leading to significantly increased costs of water treatment (Worrall *et al.* 2002). Together, enhanced export of particulate and dissolved carbon means that the blanket peat no longer continues to build up a store of carbon and increasingly becomes a source of carbon export instead. From the 1950s, many areas of blanket peat were drained (using narrow slot drains or 'grips') in an attempt to increase productivity. More recently, landowners are beginning to fill in the grips, in an attempt to restore habitat and reduce DOC export.

Further north in the Pennine Hills, there are clear signs today (2002) of revegetation of previously heavily eroded blanket peat. This may indicate that previous pressures leading to erosion have been reduced. In the southern Pennines, some gullies have begun to infill over the past twenty years but generally there is little sign of revegetation, perhaps showing that the combined influences of sheep grazing and air pollution continue to hinder recovery there.

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TIM BURT

PEDESTAL ROCK

A pedestal rock is an isolated erosional rock mass comprising a slender stem, neck or column supporting a wider cap. Also known as mushroom rocks balanced rocks and perched blocks, by local names such as *loganstones* (south-west England) and *hoodoo rocks* (North America), and by non-English equivalents such as *rocas fungiformas, roches champignons, Pilzfelsen*, pedestal rocks are developed in various climatic and lithological contexts; but especially well in sandstone, granite and limestone.

They are due to differential weathering and erosion of the cap and stem. Some are structural, the caprock being inherently more resistant than that of the stem. Pedestal rocks have been attributed to various epigene effects, and certainly, some standing in rivers and on the coast, especially where limestone is exposed, are due to physical, biotic and biochemical attack around water level. The occurrence of pedestal bedrock shapes just below the surface, however, shows that moisture attack there produces incipient indents, alcoves and concave shapes in the bedrock surface. Subsurface weathering all around the base of a block or boulder followed by lowering of the surrounding area produces a mushroom form.

Epigene effects such as differential wetting and drying on different aspects contribute to development and maintenance of the form after exposure. Sand blasting may be responsible for pedestal rocks in an immediate sense, but may exploit bedrock already weakened by weathering. Pedestal rocks are convergent forms, but most are of two-stage or etch origin.

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C.R. TWIDALE

PEDIMENT

A pediment is a gently inclined slope of transportation and/or erosion that truncates rock and connects eroding slopes or scarps to areas of sediment deposition at lower levels (Oberlander 1989). They have been reported from polar, humid and arid zones (Whitaker 1979), but they are most widely reported and studied in dryland environments, and are generally perceived as a phenomenon of DESERT GEOMORPHOLOGY.

Pediments are part of a family of landforms developed in the piedmont zone, an area of diverse geomorphology juxtaposed between uplands dominated by sediment erosion and lowlands dominated by sediment transport and deposition. Piedmonts may, therefore, be subjected to EROSION, transport or depositional process domains. These domains can vary spatially and temporally, giving rise to a complex variety of piedmont landforms, including ALLUVIAL FANs, BAJADAs and pediments.

Pediments have been variously defined in the literature. These definitions range from very general, such as ‘a terrestrial erosional surface inclined at a low angle and lacking significant relief’ (Whitaker 1979), to more specific, such as ‘surfaces eroded across bedrock or alluvium, usually discordant to structure, with longitudinal profiles usually either concave upward or rectilinear, slope at less than 11°, and thinly and discontinuously veneered with rock debris’ (Cooke 1970). Although gradients can range between 0.5° and 11°, pediments steeper than 6° are rare in the natural environment (Dohrenwend 1994). Where two pediments meet across a divide, breaking up the continuity of a mountain mass, a pediment pass is formed, often creating useful trafficable routes through upland regions. A pediplain is formed by the coalescence of numerous pediments. It should be distinguished from a PENEPLAIN, where slope decline is thought to be the major process, rather than slope retreat.

Pediments were first described by Gilbert (1877), but the term ‘pediment’ was coined by McGee (1897), and is derived from the architectural term for a low-pitched gable, especially the triangular form used extensively in classical architecture. Subsequently, the term has been applied to a variety of geomorphological forms, giving rise to considerable confusion and problems of definition (Whitaker 1979).

Difficulties are encountered when trying to define the outlines of pediments for the purpose of GEOMORPHOLOGICAL MAPPING. This also makes it difficult to derive their MORPHOMETRIC PROPERTIES, such as length, area, mean gradient, etc. (Cooke 1970). The upper margin is generally agreed to be at the piedmont angle (the junction between pediment and mountain front, usually defined as the line of maximum change in gradient in the slope profile), or at the watershed if a tributary upland area is absent. However, the downslope margin is more difficult to define. Cooke (1970) suggests this boundary should be placed where alluvial cover becomes continuous; Howard (1942) and Tator (1952, 1953) suggest it should be placed where the depth of alluvial cover equals the depth of stream scour (15 m). Other researchers have placed this boundary where the thickness of alluvial cover exceeds a small proportion (e.g. 1 per cent) of total pediment length (Dohrenwend 1994).

A variety of pediment types have been recognized and classified. Three different pediment forms can be identified using simple geomorphological criteria; an apron pediment is the common form that extends between an upland source area and a lowland depositional area; a pediment dome is formed by coalescing pediments, when the upland area has been removed; a terrace pediment is developed adjacent to a relatively stable base level such as a through-flowing stream. Other classifications distinguish between covered and exposed forms: a mantled pediment is one where crystalline bedrock is veneered by a residual weathering mantle and which is inferred to have been formed by subsurface weathering of the crystalline bedrock and wash removal of the resulting debris; a rock pediment is thought to be formed by removal of the overlying debris from a mantled pediment; and a covered pediment is developed discordantly across sedimentary rocks, having a veneer of coarse debris.

Oberlander (1989) makes an important distinction between two fundamental types of pediment; those that truncate softer rocks adjacent to a more resistant upland, and those where there is no change in lithology between upland and pediment. The first form has been widely reported, most notably along the northern margin of the Sahara Desert. These landforms have been widely studied by French geomorphologists, who term them GLACIS D'ÉROSION. These landforms truncate weak materials, and tend to be veneered by alluvial gravels, indicating the importance of fluvial processes in their creation. The second form, which has proved much more difficult to explain, is referred to by Oberlander (1989) as a 'true' pediment. Here, the pedimented surface has been cut across a lithology of similar resistance to the adjacent upland, usually a relatively resistant igneous or metamorphic rock. They typically lack the alluvial cover of the glacis-type, and show little clear evidence of fluvial processes in their formation. In the absence of an obvious mechanism a number of theories have been proposed for their formation and development, but they remain ill-defined and controversial.

At a basic level, pediments are normally viewed as a result of erosion of upland areas. This material is transported across the pediment into the lowland depositional area, and the retreating upland leaves behind an enlarging transportational pediment surface. However, problems arise when attempts are made to identify specific pediment-forming processes. Numerous processes have been proposed, but their significance is very difficult to demonstrate in practice. These proposed pediment-forming processes can be considered under three headings; surface WEATHERING, subsurface weathering and fluvial processes.

Surface weathering processes cover a wide range of SUBAERIAL processes leading to the breakdown of bedrock and REGOLITH. However, these processes do not explain the formation of a distinct piedmont angle on rocks with the same resistance to weathering. This has led Mabbutt (1966), and others, to emphasize the importance of subsurface weathering in the formation of pediments. This is largely based on the observation that pediments are widely developed on granitic bedrock, a lithology particularly susceptible to subsurface weathering. Perhaps most important here is the nature of the material produced by deep weathering of granite. The well-sorted, sand-sized GRUS forms non-cohesive channel bank material that are highly susceptible to lateral channel shifting and planation, resulting in a limited amount of channel incision (Dohrenwend 1994). The fine-grained grus can also be transported down the low pediment slopes. Mabbutt (1966) attributes the formation of a piedmont angle to slope foot notching (weathering in the subsurface layer at the base of the mountain front). However, much of this model of formation is based on assumptions based on form and occurrence, rather than on observed and well-characterized processes that can be easily validated.

Fluvial processes have been widely implicated in pediment formation, with the major emphasis being given to lateral planation. Streams debouching from the upland drainage basins are thought to erode back the mountain front by lateral channel migration. Channel incision is thought to be limited due to the high sediment load of these streams. Other research has focused on the importance of sheet floods, but this process occurs rarely in the natural environment, and its significance in pediment formation must be questioned. Sheet flooding cannot produce a planar surface, because a planar surface is necessary for sheet flooding to occur (Cooke *et al.* 1993). This vital distinction between pediment-forming and pediment-modifying processes was emphasized by Lustig (1969), who suggested that contemporary pediments were the wrong place to look for an explanation of how they were formed, since the pediments would already have to exist for these processes to operate. He suggested that geomorphologists should instead concentrate on studying the erosional processes operating in the adjacent upland drainage basins, as this is where erosion is most active. Other workers have suggested that much of the erosion leading to formation of pediments takes place in embayments, formed where streams debouch from the upland area (Parsons and Abrahams 1984). As with the subsurface weathering model detailed above, the fluvial models must be treated with caution due to the difficulties of linking observed forms with clearly defined physical processes.

The main difficulty in explaining development of pediments is the problem of maintaining parallel rectilinear retreat of permeable slopes in a SAPROLITE-mantled landscape. Oberlander (1989) proposes that rectilinear retreat occurs because sediment transport processes are limited by deep permeability of grus, eluviation of fines by throughflow, and accelerated subsurface weathering by soil moisture, concentrated at the base of slopes. Twidale (1978) suggests that lithological and structural features within granitic massifs (petrological variations and differences in joint density) are important controls on pediment morphology, but other work has failed to demonstrate any clear relationships. The importance of tectonics in pediment formation is also uncertain, although, in general, pediments appear to be best developed in areas of long-term stability (Dohrenwend 1994).

With improvements in dating techniques, there is a growing amount of evidence indicating that some pediments are extremely ancient. In the Sahara and Mojave desert (Cooke and Reeves 1972; Plate 87), lava flows can be seen to bury existing pediment surfaces. This raises the possibility that they may be relict landforms formed under different climatic conditions pertaining in Tertiary or even Late Mesozoic times (Oberlander 1989). Specifically, the arid-zone processes acting on contemporary desert pediments may not be appropriate to explain landforms that developed over timescales that embraced humid as well as arid phases. A variety of conditions have been proposed as being optimal for pediment development of crystalline rocks; these include seasonally wet, low-latitude forest, savanna and cold-winter deserts affected by cryogenic processes. Oberlander (1989) suggests that pedimentation is currently active in parts of central Arizona, which appears to replicate conditions in the Mojave Desert in the Miocene. It seems likely that many pediments must be regarded to some extent as relict landforms, currently being modified under very different environmental conditions from those that pertained during their initial formation.

Plate 87 Pediment in the Mojave Desert, southwest USA

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SEE ALSO: alluvial fan; desert geomorphology; glacis d'érosion

KEVIN WHITE

PENEPLAIN

Peneplain is the term coined by W.M. Davis to mean a surface of low relief worn down to near sea level and formed through erosion over protracted spans of time. His own words are:

> Given sufficient time for the action of denuding forces on a mass of land standing fixed with reference to a constant base-level, and it must be worn down so low and so smooth, that it would fully deserve the name of a plain. But it is very unusual for a mass of land to maintain a fixed position as long as is here assumed. . . . I have therefore elsewhere suggested that an old region, nearly base-levelled, should be called an almost-plain; that is a peneplain.
>
> (Davis in Chorley *et al*. 1973: 190)

The peneplain is thus not the end-product of a cycle of erosion and, if keeping with Davis's way of thinking, it should not be confused with an endless and featureless plain as is often implied. Rather, a peneplain is a regional landscape at the penultimate stage of development which is yet to be eroded down to a true plain. In another place Davis himself says:

> At a less advanced stage of degradation, the land will still possess low, unconsumed hills along the divides and subdivides between the broad-floored rivers. It will then be almost-a-plain, or a peneplain. A peneplain will be hardly above sealevel at its base, but if the area is large it may attain altitudes of 2,000, 3,000, or 4,000 feet far inland near the river heads, and its residual mounts and hills may rise still higher, although with gentle slopes.
>
> (Davis in King and Schumm 1980: 8)

The processes leading to a peneplain would be mainly subaerial, chiefly fluvial and gravity-driven hillslope processes. They ought to be in action long enough to obliterate the effect of unequal rock resistance, so only the hardest rocks would form bedrock-built hills rising above the peneplain, the monadnocks. Otherwise, gentle slopes would be underlain by deeply weathered rock, with the thickness of weathering mantle being in excess of 10 m. As far as the relative relief of a peneplain is concerned, Davis seemed to be rather vague in defining any critical hill heights or slope angles. Therefore, there are two crucial characteristics of a peneplain. One is its temporal context within the cycle of subaerial erosion. To be a peneplain, the surface of low relief must have formed in the course of protracted denudation. The second prerequisite is grading down to sea level.

Much of the substantial confusion around the term results from the fact that subsequent workers did not always keep with Davis's original definition and used the term in various contexts. For example, peneplains were often equated with PLANATION SURFACES, or a particular mode of slope evolution was implied for a peneplain. Many geomorphology textbooks contrast peneplains formed mainly through slope downwearing and consequent relief reduction, with pediplains formed by slope backwearing and relative relief maintained high until a rather late stage of development. In other cases the condition of being located close to, or graded to, sea level was ignored. In consequence, flattened summit surfaces within mountain ranges were frequently called peneplains despite the fact that neither their origin nor the age were known sufficiently to warrant the use of the term in its strict, Davisian sense. Davis himself suggested the term 'pastplain' to describe a peneplain which has been uplifted and now shows the initial stage of dissection.

Free usage of the term and its obvious connection with the Davisian model of cyclic landform development and the DENUDATION CHRONOLOGY approach, themselves strongly criticized since the 1960s, had eventually led to its declining popularity and gradual abandonment. Preference was given to more neutral 'planation surfaces' in describing landscapes, whereas in the field of theory a search for non-cyclic models of geomorphic development was pursued strongly.

Nonetheless, Fairbridge and Finkl (1980) proposed to return to 'peneplains', but realizing the potential for confusion and misuse they suggested disassociation from restrictions implied by the Davisian definition. Instead, they preferred to give the term a non-genetic meaning, simply to describe a near flat surface regardless of its origin, setting and evolutionary stage. This point has apparently been taken forward by Twidale (1983) who describes peneplains as 'rolling or undulating surfaces of low relief', without referring to their position in respect to BASE LEVEL. Moreover, he argues that there is no means to decipher the mode of past slope evolution leading to the present-day peneplain; hence the argument focused on the backwearing-or-downwearing issue is by and large pointless. On the other hand, he firmly adheres to the Davisian understanding of the peneplain as a landscape at the penultimate stage of evolution and introduces 'ultiplains' as the true end-products of relief development. By contrast, Phillips (2002) in his most recent review offers a broader definition of the peneplain in which the condition of being at any certain stage of a cycle has been made redundant. Given all these divergent views, the term is very difficult to recommend for routine application in describing and explaining landforms.

There has been much debate as to whether peneplains and peneplanation, in the truly Davisian sense, really occur. Phillips (2002), following his many predecessors, claims that contemporary peneplains eroded to near sea level are almost non-existent and seeks the reasons in constant variations in tectonic forcings, climate and base level, especially in the Quaternary. All these changes would not allow peneplanation to last for long, and induce surface dissection rather than planation. On the other hand, Twidale (1983) gives a number of examples of almost perfect rock-cut plains, but demonstrates their antiquity at the same time. Many of these plains date back to the Cretaceous or even beyond. In Fennoscandia, a peneplain of subcontinental extent undoubtedly existed at the end of the Precambrian (Lidmar-Bergström 1995). Further examples of past plains, or palaeoplains, have been reviewed by Ollier (1991). It appears that reconciling the evidence for little or no peneplanation at present and widespread planation in the geological past is one of the challenges for evolutionary geomorphology.

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SEE ALSO: Cycle of Erosion; denudation chronology

PIOTR MIGOŃ

PERIGLACIAL GEOMORPHOLOGY

Perhaps surprisingly, there is no agreement as to what exactly constitutes terrain which can be regarded as periglacial for there are no quantitative defining parameters which have gained universal acceptance. However, most would accept the proposition that there are two possible approaches to the demarcation of what is periglacial, and that both can be justified.

One would emphasize the requirement for intense frost action in the form of frequent FREEZE–THAW CYCLES and deep seasonal freezing. If these criteria are used to delimit the distribution of the periglacial domain, it follows from this that some 35 per cent of the Earth's continental surface (mainly in the northern hemisphere) falls into the periglacial category. The other approach would stipulate that the presence of perennially frozen ground, i.e. PERMAFROST, is paramount. Permafrost may be defined as any earth material that has maintained a temperature at or below 0 °C for a minimum period of two years. Note that there is no reference to water content or lithology in this definition. If permafrost is the fundamental attribute, then a more rigorous climate than that needed for frost action alone is required to qualify for periglacial status, as permafrost demands a mean annual temperature of below 0 °C. As a result the global periglacial area would be substantially less at around 20 per cent of the total terrestial surface. Nevertheless, from both perspectives the total area regarded as periglacial remains a considerable part of the Earth's terrestrial environment. By way of comparison, the area of permanent snow and glacial ice is only 3 per cent.

Relief also has an influential role in determining the distribution of periglacial regions. Both freeze–thaw cycles and permafrost are related to climate and this is influenced by both latitude and altitude. The outcome is that the most widespread periglacial areas are mainly lowlands in northern Eurasia and North America and these incorporate both tundra and boreal forest landscapes. Mountain temperatures are influenced by elevation which is sensitive to the lapse rate. This can produce sufficient cooling at lower latitudes to over print the more temperate conditions in the adjacent lower areas. As a consequence alpine periglacial areas can occur even at the equator. Usually they reveal a tundra zone with the lower limit roughly corresponding with the upper treeline.

Basic periglacial processes

Periglacial geomorphology focuses primarily upon those terrestrial surface processes, sediments and resultant landforms which characterize the cold non-glacial areas of the Earth's surface. Basic to comprehension of these is a knowledge of the somewhat anomalous physical behaviour of water substance. First, there is a 9 per cent increase in volume as the phase changes from liquid to solid and conversely by a decrease from solid to liquid which occurs during freezing and thawing. This is accompanied by a large latent heat of fusion (84 calories per gram) not much less than the heat required (100 calories per gram) in changing the temperature of the liquid phase from the solid to gas transitions. The net result is that the rate of both freezing and melting are delayed more than might be expected. Second, volumetric changes occur during temperature variations in the solid (frozen) state when cooling produces contraction and vice versa. This in itself is not unusual but to avoid confusion it has to be seen as being totally independent on the 9 per cent shift at the liquid–solid–liquid phase change. Third, maximum density is achieved in the liquid phase, some 3.98 °C above the freezing point, ensuring that the solid phase floats on the liquid. This has profound implications for life as it means that water is present beneath lakes and rivers with ice covers. Even in the harshest climates water bodies over 3 m deep do not freeze to their beds as annually developed ice rarely exceeds 2 m in thickness. Fourth, within the sediment pore the freezing point of water can be lowered down to −22 °C in extreme cases. This is especially effective in fine-grained sediments (clays and silts) where the movement of thin films of water occur even though the ground is technically frozen. This facilitates the aggradation of ice masses with volumes well in excess of the pore capacity. All these factors contribute to the landscape-forming processes associated with periglacial environments and collectively these determine the nature of periglacial processes. Driving these processes is temperature change and in its turn weather and climate.

Palaeoperiglacial activity

Apart from the unusual physical behaviour of water substance, a further complicating factor in understanding periglacial features is the temporal dimension, particularly that allied to climatic change. This can be illustrated by the example of Britain. If delimiting areas subject to periglaciation on the basis of freeze–thaw cycles is accepted, then the higher summits of Britain remain within the ambit of periglacial activity. This is the viewpoint taken by Ballantyne and Harris (1994) in their major regionally based synthesis. However, adoption of the alternative less permissive approach based on the presence or absence of permafrost, inevitably means that the British Isles cannot be regarded as periglacial since the last permafrost finally dissipated some 11,500 years ago towards the end of the Last Glacial stage. Since then through the ensuing Flandrian interglacial (postglacial) even the climatically most extreme locations have been unable to sustain any permafrost. This was the position taken by Worsley (1977) in reviewing British periglaciation when it was concluded that all the periglacial evidence in Britain was effectively relict.

Prior to 11,500 years ago, during the Last Glacial stage, most of Britain had experienced episodic extensive permafrost. Indeed, in many areas of the world the Last Glacial stage witnessed the dramatic expansion of the periglacial realm by up to 50 per cent. But this was probably never so dramatic as in western Europe where the mean annual temperature dropped by some 20 °C during the time when the Gulf Stream was inoperative. An underemphasized aspect of the global glacial stages is that if sea level is taken as a proxy for palaeoclimate then the durations of very low sea levels (maximum glacial ice volumes) were relatively short. Similarly Quaternary marine oxygen isotope ratio records are interpreted as reflecting in large part the degree of global glaciation and the negative ratio peaks in the curves are taken to correspond with the periods of most extensive glacial ice cover (corresponding to the lowest sea levels). In many areas of the world covered by glaciers in the Last Glacial stage, the stratigraphic evidence indicates that the glacial advance which culminated in the most extensive ice cover occurred late in the stage. These data imply that for much of the glacial stage those areas which were to become glaciated were cold but non-glacial in character, i.e. periglacial processes rather than glacial processes prevailed. Hence many of the glaciated landscapes bear a partial periglacial imprint. Naturally those areas immediately outside the maximum ice extent limits witnessed a periglacial regime for much of the glacial stage and hence the effects of periglaciation are clearer. Finally, the earlier phases of ice retreat from the maximum limits were primarily the result of reduced snowfall rather than a temperature increase. This enabled the perglacial environment to extend into the areas recently vacated by the ice. PARAGLACIAL conditions follow ice withdrawal and to an extent these might be regarded as part of periglaciation.

Historical development of the periglacial concept

The term periglacial was first coined in 1909 by the Polish geologist and pedologist W. Łoziński in his account of the mechanical weathering of sandstones and blockfield (see BLOCKFIELD AND BLOCKSTREAM) production under inferred cold climates in the Carpathian Mountains. Three years later Łoziński introduced the concept of a periglacial facies produced by mechanical weathering, although he did not give it any quantitative climatic parameters. However, it is clear that Łoziński was proposing the periglacial concept in an attempt to reconstruct the palaeoenvironmental context of his facies. He envisioned it as diagnostic of the former processes operative in terrain immediately adjacent to glaciers and ice sheets of Pleistocene age rather than as a function of contemporary activity. His second paper was published as part of the proceedings of an international geological congress held in Stockholm in 1910 and this ensured that the term periglacial was widely disseminated. Other activities at the congress included a field excursion to Spitzbergen where periglacial facies could be related to contemporary environmental processes, and thereby gave further impetus to the scientific study. Strictly therefore, periglacial, as originally envisaged by Łoziński, should refer to the area or zone formerly subject to arctic-type climatic conditions peripheral to a glacier.

From the standpoint of modern usage, an erroneous impression might be given that periglacial features are exclusively associated with the area around glacial margins. On the contrary, some of the major areas of permafrost today have never been glaciated or indeed been peripheral to former ice sheets. The prime example of this is east Siberia where the permafrost can exceed over 1 km in depth. This is probably explained by the fact that it has experienced the longest history of sustained permafrost development anywhere on the Earth. A further disadvantage is that there might be the assumption that areas peripheral to glaciers experience a rather less severe climate and that a climatic deterioration would necessarily lead from the periglacial to glacial, with glaciation representing the ultimate severe climatic state. Despite a number of workers having argued for the term periglacial to be abandoned because of its imprecision, its usage is now widespread and a degree of permissiveness in its definition is automatically accepted by its users. It is interesting to note that there was a change in the title between the first and second editions of A.L. Washburn's synthesis of periglacial environments (Washburn 1973, 1979). In the latter the term GEOCRYOLOGY was introduced (i.e. the investigation of frozen Earth materials). The word is derived from the Russian equivalent and although it can include glaciers it is usually directed towards frozen ground.

Although Łoziński formalized the notion of a periglacial zone early in the twentieth century, as with many concepts in Earth science earlier workers anticipated later formulations. For example, the commencement of government-sponsored geological mapping in England in the 1830s soon led to the identification of a relict mantle of rock debris by De la Beche overlying the slopes of Cornwall and Devon in the south-west peninsula. Field relationships demonstrated to him that this 'head' blanket of angular fragments had been derived from mechanical weathering of the bedrock cropping out on the slopes above and that there had been an ubiquitous downslope movement which had tended to even out any terrain irregularities, thereby anticipating something akin to SOLIFLUCTION.

Unique periglacial processes and landforms

The research literature arising from investigations of periglacial geomorphology has given rise to a wide range of specialized terms and names of unique landforms. These have come from a number of language sources and some duplication and confusion have arisen because of inconsistent usage.

This was exemplified in the planning of this encyclopedia since the draft list of topics specified both altiplanation and CRYOPLANATION as separate entries. In reality there is no difference between the two. Altiplanation terraces were first described by H.M. Eakin in 1918 following field mapping in part of eastern Alaska where he encountered benches and summit surfaces cut in bedrock largely independent of lithology and structure. Similar relationships had earlier been identified in Russia and the term 'goletz' terrace applied to them. Other terms which have been used include equiplanation and nivation terraces. Bryan (1946) undertook a wide-ranging review of the then existing periglacial terminology and amongst others proposed a new term – cryoplanation – to express the unified concept of frost action and frost-related downslope movement of debris to produce a degradation system eroding and lowering hillslopes. This is now the internationally agreed term for such features.

Fortunately the confused nomenclature has been subject to clarification in a comprehensive glossary produced by a very experienced interdisciplinary team of Canadian permafrost workers (Harris *et al.* 1988). This is an excellent source of current usage, definitions and synonyms used in periglacial geomorphology. It also has the additional merit of thoughtfully discussing many of them and the authors are not reticent in recommending the abandonment of some cherished terms!

Periglacial geomorphogy, like any morphogenetic geomorphology, should consider all the geomorphological agents which contribute to the landscape character. Naturally there is a tendency to concentrate upon those agents which are either unique to or are readily associated with it at the expense of those which are common to a range of environments. To illustrate this point there are over fifty entries in this encyclopedia which are relevant to periglacial geomorphology. Yet only half of these are likely to be discussed or referenced to the periglacial realm. Examination of most periglacial environments, in the field or from maps and air photographs, normally reveals an essentially fluvial landscape displaying a 'normal' drainage network. There are some exceptions and significant parts of the periglacial regions display desert landscapes. This is not surprising considering some of the most arid areas of the world are underlain by permafrost. But these deserts are cold. There is tendency to regard deserts as hot places for this is where the vast majority of desert geomorphologists work.

Applied periglacial geomorphology

Wherever there is ice within the subsurface there is always the possibility that it might melt. Under natural conditions this is an ongoing process and can occur through a range of incidents such as forest fires, coastal and riverbank erosion, or climatic amelioration. Indeed the prospect of global warming carries severe implications for the entire permafrost world.

Over the last century there has been progressive settlement of the permafrost terrain by people from more southerly regions who had an expectation of a similar range of facilities to those south of the permafrost region. This movement was given a particular impetus by the Second World War and operation of defence facilities during the Cold War. Later economic exploitation of mineral resources and hydrocarbon exploration placed further demands for transport, urbanization and allied installations.

Construction of all kinds on permafrost terrain is potentially hazardous if the natural ground thermal equilibrium is disturbed as this will induce melting of the ground ice and cause thaw consolidation. Under the stress of war a number of mistakes were made in road and pipeline construction but experience was gained from tackling the challenges presented by permafrost. A landmark publication was the compilation by Muller (1947) of the then state of the art understanding of permafrost and its allied engineering problems. This drew extensively upon Russian experience. It led to the founding of the US Army's Cold Regions Research and Engineering Laboratory which has subsequently been one of the leading institutions engaged in periglacial research. Similar laboratories were established in Yakutia and Canada with primarily civilian missions.

In Canada in the 1950s, Aklavik, the pre-existing administrative centre in the Mackenzie delta region, was suffering from annual breakup floodings. A decision was taken by the Federal Government to construct a new town to replace it which would incorporate the 'best practice' in permafrost construction (Johnston 1981). A number of sites were assessed in terms of their periglacial geomorphological attributes with that at Inuvik selected for development. Inuvik has since become the show piece of how a small town offering the facilities of the non-periglacial world can be created without significant environmental damage. There all the buildings are well insulated and usually placed on piles which penetrate pads of non frost-susceptible materials carefully placed on the original vegetation. A 1 m high air gap through the tops of the piles enables the maintenance of the natural ground thermal regime. Using the same approach, a system of water, sewage and heating pipes were installed in a duct network (utilidor). In some instances, such as power generating units, piles were not feasible and thick pads of granular materials, through which ventilation pipes were inserted, have succeeded in achieving the same objectives.

A vastly improved appreciation of periglacial LANDSCAPE SENSITIVITY has largely ensured that land use activities can be undertaken without major disastrous consequences. Even so construction has to be closely monitored by environmental managers versed in the basics of periglacial geomorphology and in the field of hydrocarbon exploration a number of drill sites have been closed in the summer for fear of excessive disturbance to the ground ice within the permafrost.

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SEE ALSO: active layer; alas; cryostatic pressure; frost and frost weathering; frost heave; hummock; ice wedge and related structures; icing; loess; needle-ice; nivation; niveo-aeolian activity; oriented lake; palsa; patterned ground; pingo; protalus rampart; rock glacier; thermokarst

PETER WORSLEY

PERMAFROST

Permafrost is defined as ground (soil or rock) that remains below 0 °C for at least two years, and the term is defined purely in terms of temperature rather than the presence of frozen water (Permafrost Subcommittee 1988). Permafrost may, therefore, not contain ice, or may contain both ice and unfrozen water. In many cases, however, ground ice forms a significant component of permafrost, particularly where the substrate comprises fine-grained unconsolidated sediments. The geothermal gradient below the ground surface averages around 30 °C km^{-1} (Williams and Smith 1989) and this increase in temperature with depth determines the thickness of the permafrost (Figure 120). Seasonal temperature fluctuations lead to above zero ground surface temperatures in summer, and the downward penetration of a thawing front. The surface layer that freezes and thaws seasonally is called the active layer, and its thickness depends on the ground thermal properties and on the ratio of the summer thawing index (the accumulated degree-days above freezing) to the winter freezing index (accumulated degree days below freezing). The annual cycle of winter cold and summer warmth is propagated downwards into the permafrost, but rapidly attenuated, so that it becomes undetectable below around 15 m (Figure 120). This is termed the depth of zero amplitude (Brown and Péwé 1973). Longer term changes in ground surface temperatures cause downward propagation of a thermal perturbation, and in many permafrost sites today the geothermal gradient is non-linear, with warm-side deviation that increases towards the surface (Lachenbruch and Marshall 1986), indicating warming over the past century or more (Figure 120).

In northern Canada, permafrost is up to 600 m thick (Figure 121) and its thickness decreases southwards as the climate becomes warmer. Eventually, local variation in ground conditions leads to breaks in the continuity of permafrost, and a complex pattern of discontinuous permafrost results. Under still warmer climatic conditions

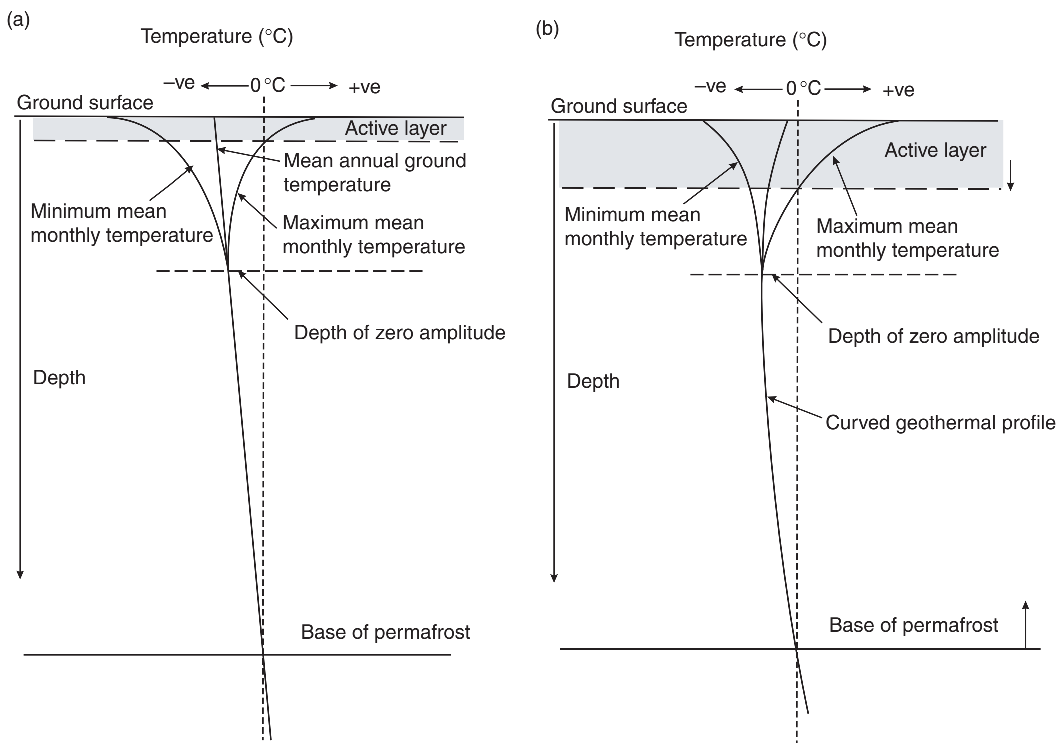

Figure 120 Thermal profiles in permafrost: (a) equilibrium and (b) during thermal adjustment to surface warming

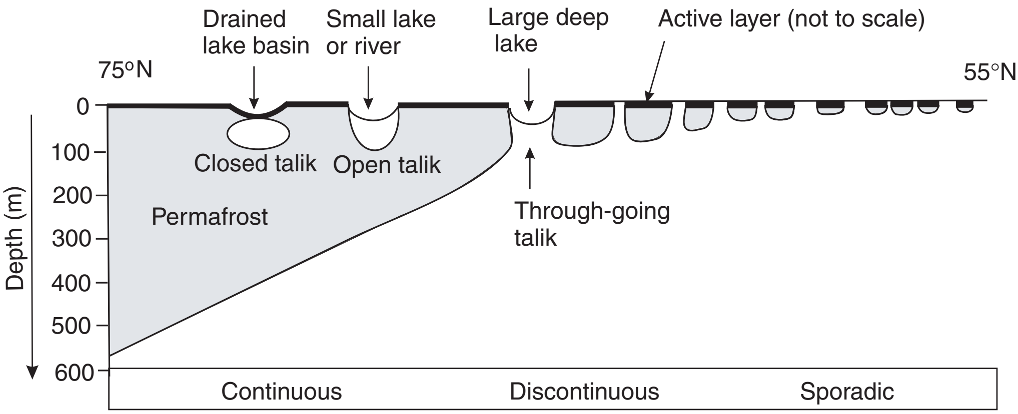

Figure 121 Typical permafrost characteristics along north–south transect, north-west Canada (after Lewkowicz 1989)

permafrost may form only isolated patches (often related to areas with peat cover), and is described as sporadic. Siberian permafrost is generally colder than North American and in places is in excess of 1,000 m thick (Williams and Smith 1989). However, it takes many millennia for thick permafrost to adjust to surface warming, and it is likely that Siberian permafrost remains chilled by the severity of the last Quaternary cold stage, and is not in thermal equilibrium with present-day conditions.

Taliks are unfrozen zones within permafrost terrain that generally occur beneath large bodies of water such as lakes or rivers that do not freeze to their beds in winter. The unfrozen lake or river water is warmer than 0 °C and therefore constitutes a heat source causing a thaw bulb to develop in the underlying permafrost. Drainage of lakes causes downward advance of permafrost, creating a closed talik, entirely surrounded by permafrost (Figure 121). Hydrochemical taliks may be cryotic (below 0 °C), but remain unfrozen due to the flow of mineralized ground water, while hydrothermal taliks may remain non-cryotic due to heat supplied by groundwater flow.

Mean annual permafrost ground surface temperatures are usually higher than the corresponding mean air temperature by a few degrees, so that defining permafrost distribution on the basis of air temperature can be misleading. However, Brown *et al*. (1981) used a mean annual air temperature of approximately −8 °C to delimit the boundary between continuous and discontinuous permafrost in North America, and −1 °C to define the southern limit of discontinuous permafrost. Williams and Smith (1989) stress the multitude of factors that influence the development and survival of permafrost, and point to a gradual southward transition from continuous to discontinuous to sporadic permafrost, with local factors leading to wide variations in associated mean air temperatures.

Where permafrost is developed in unconsolidated sediments, it commonly contains ground ice. Mackay (1972) has provided a classification of ground ice, identifying four categories: pore ice, segregation ice, vein ice and intrusive ice. Both pore ice and segregation ice occur in seasonally frozen soils, but vein ice and intrusive ice occur only in permafrost. Pore ice refers to the ice occupying the pore space in ice-cemented permafrost, and is particularly important in sands and gravels. In fine-grained soils (silts and clays) and porous rocks, much of the pore water occurs as thin films within which capillary and adsorption effects lower the freezing point by several degree celsius (Burt and Williams 1976; Williams and Smith 1989). Progressive freezing of such water results in development of cryosuction, causing water to migrate towards the freezing front. Here it freezes to form lenses of clear ice (segregation ice), increasing ice contents to well in excess of the natural saturated moisture content. Ice segregation during freezing of fine soils causes a significant increase in soil volume and upward frost heaving of the ground surface. Vein ice is the ice that accumulates within permafrost as ice wedges as a result of thermal contraction cracking (see ICE WEDGE AND RELATED STRUCTURES).

Finally, intrusive ice may form layers up to several metres in thickness as a result of pressurized water flow towards the freezing zone. The pressurized water may be derived from groundwater flow beneath permafrost (open system), or arise from porewater expulsion ahead of a penetrating freezing front in saturated coarse sands and gravels (closed system). Expulsion of water results from the expansion that occurs as pore water changes phase from water to ice. Freezing of pressurized water close to the ground surface in both open and closed systems is responsible for the formation of distinctive conical hills, or PINGOs, the pingo ice being a common form of intrusive massive ground ice (Mackay 1998). Not all massive ice bodies within continuous permafrost originated as intrusive ice, however. In Siberia and parts of northern Canada, ice bodies are considered by some to represent buried glacier ice (Astakhov *et al.* 1996; French and Harry 1990).

The presence of ice-rich permafrost results in high terrain sensitivity to surface thermal disturbance. Permafrost degradation caused by climate warming leads to slope instability and differential settlement as ground ice thaws (French and Egginton 1973). The resulting irregular surface relief is termed THERMOKARST. Widespread thaw settlement in the Arctic has been predicted due to twenty-first-century global warming (Nelson *et al.* 2001). In high altitude mountains, such as the Rockies, Himalayas and the European Alps, discontinuous permafrost is commonly present. Mountain permafrost distribution is generally complex, reflecting altitude, aspect and ground cover, particularly snow cover in winter. Terrain sensitivity to atmospheric warming is again high, and the presence of steep mountainsides increases potential hazards from landslides, debris flows and rockfall (Harris *et al.* 2001).

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CHARLES HARRIS

PHYSICAL INTEGRITY OF RIVERS

Human activities have dramatically altered the forms and processes of the Earth's river systems. In the northern third of the globe, almost 80 per cent of the rivers are segmented by dams (Dynesius and Nilsson 1994), while in technologically advanced countries such as the United States more than 98 per cent of rivers are significantly impacted by human activities (Echeverria *et al.* 1989). The recognition that the far-reaching effects of channelization, levee building and dam construction have affected biodiversity of aquatic and riparian environments has produced governmental policies

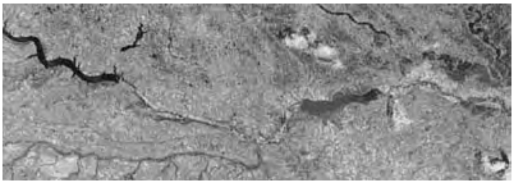

Plate 88 US Geological Survey LANDSAT image of the middle Missouri River in north-central United States. North is at the top of the image, which extends about 175 km east–west and about 88 km north–south. The Missouri River, largely lacking physical integrity, flows from the left (west) side of the image to the right (east) side. The dark, wide areas are reservoirs behind large dams, while the remaining connections between the dams and the next reservoir downstream are channels where fluvial processes are controlled by the dams. The Niobrara River enters the view from the lower left (south-west) corner of the image. It retains some of its physical integrity and does not include large dams

in many countries to restore rivers and their environments to more natural conditions. Although it is rarely possible to restore rivers to primeval anatural conditions, many nations have policies to promote the physical integrity of rivers, a term that often appears in legislation. Thus, the term integrity has its origins in legal language and usage.

From a scientific and engineering perspective, a physical integrity for rivers refers to a set of active fluvial processes and landforms wherein the channel, floodplains, sediments and overall spatial configuration maintain a dynamic equilibrium, with adjustments not exceeding limits of change defined by societal values (Plate 88). Rivers possess physical integrity when their processes and forms maintain active connections with each other in the present hydrologic regime (Graf 2001). Each term in this definition has particular meaning for the geomorphologist:

- streams and rivers: those parts of the landscape with confined water flow;
- fluvial processes and forms: those features related only to the fluvial domain;
- channel, near-channel landforms, sediments: channel area that is active in the present regime of the river (having a return interval of interaction with flow of 100 years or less), near-channel landforms include the functional surfaces that interact with fluvial processes and channels in the present regime, sediments that are active in the present regime;
- configuration: planimetric and cross-sectional arrangement of functional surfaces, landforms and sediments;
- dynamic equilibrium: the tendency for parameters describing the river to change annually about mean values which also change over periods of decades or centuries;
- limits of change defined by societal values: dimensional and spatial changes in forms and processes within ranges that are acceptable for economic, social or cultural reasons; changes greater than limits imposed by society result in re-engineering the channel to protect lives and property;
- present hydrologic regime: decade or century-long behaviour of daily stream-flow values for magnitude, frequency, duration, seasonality and rates of change.

Measurement of physical integrity for rivers depends on use of a few easily defined, readily assessed indicator parameters. These parameters must have strong roots in the geomorphological literature so that researchers may take advantage of existing knowledge and theory, but the parameters must also be understandable by decision-makers and the public. The parameters must be few in number, and be readily available or easily measured by non-specialists, because river management entails contributions from a variety of observers. Although the range of choices for such parameters in the literature is large (Leopold 1994), the following are the most commonly used and are measured at cross sections: daily water discharge, active channel width, sinuosity, pattern and particle size of bed material.

Of these indicator parameters, daily water discharge is the most important. These data, often collected and made widely available by governmental agencies, provide insight in to the primary driving forces and masses that control the river system. Human interactions with rivers often directly impact the water discharge, with subsequent effects rippling through the geomorphic system. Active channel width is the most easily assessed morphological variable for rivers, and it is the variable that is most sensitive to changes imposed by human activities through discharge adjustments. Classic hydraulic geometry usually shows that width adjusts more than depth or velocity with changes in discharge (Knighton 1998).

SINUOSITY and channel pattern are easily measured on aerial photography for small or medium-sized rivers, or on satellite imagery for large rivers. These parameters reflect upstream impacts of human activities that alter the delicate balance among water, sediment and channel form. Bed material size is readily assessed in field measurements and is sensitive to sediment supply as well as transport capacity (total amount of sediment the river is capable of transporting) and competence (the maximum size of particle that can be transported). Sediment discharge data are also informative, but such data are often not available because they are expensive to measure.

Physical integrity for rivers is important because it underpins biological integrity of river environments. Biological integrity, often characterized by biodiversity and sustainability of ecosystems, depends on the physical substrate of water, sediment and landforms. Efforts to restore rivers to more natural conditions are often thought of in biological terms by planners and managers, but restoration of the underlying physical system must occur first before the biological components of the system can assume more natural conditions.

The most significant human activity in reducing the physical integrity of rivers is the installation of dams. Dams segment the original river system and at least partially control the flow of water and sediment in downstream reaches. Dams reduce peak flows for flood protection, but high flows are important in activating functional surfaces near the channel. As a result, the floodplains downstream from many dams become inactive, causing substantial ecosystem changes that include wholesale vegetation changes. Birds and animals dependent upon the pre-dam vegetation face a loss of habitat under post-dam conditions when the physical and biological integrity of the river decline. In dryland settings, dams often divert the entire flow of the river, resulting in the dessication of reaches that once supported riparian habitat for diverse flora and fauna. In some urban areas, the original stream is replaced by a totally artificial channel without floodplains or other active features, and engineering works often seek to replace braided channels with single thread channels. The restoration of rivers with physical integrity in such cases is a scientific and engineering challenge (Brookes and Shields 1996; Petts and Carlow 1996) that must balance competing objectives subject to social valuation.

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SEE ALSO: floodplain

WILLIAM L. GRAF

PHYSIOGRAPHY

A word that has obscure origins, although it was in common currency in eighteenth-century Scandinavia, and in regular usage in the English-speaking world in the nineteenth century (see Stoddart 1975, for a historical analysis). Dana defined it in 1863:

> Physiography, which begins where geology ends – that is, with the adult or finished earth – and treats (1) of the earth's final surface arrangements (as to its features, climates, magnetism, life, etc.) and (2) its systems of physical movements and changes (as atmospheric and oceanic currents, and other secular variations in heat, moisture, magnetism, etc.).

One of the most notable exponents of physiography was the British naturalist T.H. Huxley, who published a highly successful text, *Physiography*, in 1877. Huxley's *Physiography* has some geomorphological content including chapters on 'the work of rain and rivers', 'ice and its work', 'the sea and its work', 'slow movements of the land' and 'the formation of land by Animal Agencies'. In the USA W.M. Davis preferred the term to GEOMORPHOLOGY, but he used it without the catholicity of meaning that it had for Huxley. Various other American geomorphologists, including J.W. Powell and N. Fenneman, divided up the USA into what they termed Physiographic Regions, Provinces or Divisions (see Atwood 1940).

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A.S. GOUDIE

PIEZOMETRIC

A term used in the study of groundwater hydrology referring to the underground water pressure, though the term has particular relevance to underground aquifers and stability analysis of slopes (and PORE-WATER PRESSURE). Underground flow is largely unseen and so instruments called piezometers are used to provide an indication of pressure potential in the water table at particular depths. There are several types of piezometers, including well piezometers, standpoint piezometers and hydraulic piezometers, the commonest being a standpoint piezometer. A typical standpoint piezometer consists of a generally imperforated tube (typically 1–3 m length) inserted into the layer or horizon of interest, with a porous screen at one end to permit flow (about 25–50 mm diameter). Clean sand is placed around the screen, and the borehole surrounding the piezometer tube is filled with a seal (e.g. cement) to ensure that the pressure value given reflects only that on the screen tip. More comprehensive descriptions of piezometer instrumentation is provided in Goudie (1994: 237).

The piezometer gives the potential pressure in a soil or rock by measuring piezometric head, referring to the energy possessed by the water (also termed potentiometric head, hydraulic head and pressure head). Thus, the piezometric head at a point on the water table is the level above an appropriate datum (for instance sea level) that the water table reaches. Differences in head between points in the same water table will result in the transportation of energy (and underground flow) from the point of high piezometric head to that of lesser piezometric head. The velocity of flow between the two points should be directly proportional to the difference in head between, as long as all else remains equal (known as the piezometric gradient). The piezometric gradient can be influenced by external factors, such as precipitation (high amounts of precipitation resulting in a greater piezometric gradient).

By interpolating measurements of piezometric head, an imaginary surface, termed a piezometric surface (though the term potentiometric surface is preferable) can be formed. This represents the distribution of potential energy within the water body. A piezometric surface that lies above ground level will result in flowing water on land. This is termed an artesian well, and is typical of synclinal structures. Insufficient piezometric pressure to reach above ground level is termed subartesian. In unconfined aquifers, the slope of the piezometric surface defines the hydraulic gradient.

Maps of the piezometric surface can be developed, joining together points of equal piezometric head (using contours known as equipotential lines). Flow takes place perpendicular to these lines (down the piezometric gradient), and largely parallel to the overlying surface (Jones 1997: 93). Contour maps provide indications of the piezometric gradient and the pattern of the subsurface flow, and can also be used in stability analysis of soils and rocks.

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STEVE WARD

PINGO

A pingo is a perennial PERMAFROST mound or hill formed through the growth of a body of ice in the subsurface. The term 'pingo' has been taken from a local Inuktitut word (meaning conical hill) used in the Mackenzie Delta region of the western Canadian Arctic. The term is now used globally to refer to this particular type of ice-cored mound, although in Siberia the Yakutian term *bulgannyakh* is sometimes used.

There are estimated to be some 5,000 or more pingos worldwide with the highest concentration occurring in the Tuktoyaktuk Peninsula area of the western Canadian Arctic, where there are around 1,350 pingos (Mackay 1998). Other areas with considerable numbers of pingos include the Yukon Territory (Canada), the Canadian Arctic Islands, northern Québec (Canada), northern Alaska (USA), northern areas of the former Soviet Union, the Svalbard archipelago (Norway) and Greenland. A few examples have been noted from Mongolia and the Tibetan Plateau.

Pingos can attain heights in excess of 50 m, with basal diameters of over 600 m. The ice volume of large pingos can exceed 1 million m^3. Whilst pingos can form almost conical hills, they may have more irregular forms and can be oval or elongate rather than circular in plan.

In order for a pingo to form and grow, water under pressure must be delivered to a position beneath the surface within a continuous or discontinuous permafrost environment. This water is frozen to form the ice core which is often described as intrusive or injection ice. As the pingo ice core grows the material overlying it (the pingo 'skin' or overburden) is forced upward forming the mound or hill.

Pingos can be classified as either hydraulic (previously termed 'open system' or 'east Greenland' type) or hydrostatic ('closed system' or 'Mackenzie Delta' type). This classification is based on the origin of the groundwater feeding the growth of the pingo ice core.

Hydraulic pingos are initiated by water under pressure of a hydraulic head/potential coming towards the surface in a valley bottom or lower valley side position and they are, therefore, features of high relief environments (for example, east Greenland, Alaska, Svalbard). The position of the upwelling of the ground water may shift over time and in this way a group or complex of pingos may develop within a relatively small area. The largest documented group is the 'Zurich Pingo group' in the Karup Valley of Traill Island, east Greenland (Worsley and Gurney 1996) which has some eleven pingos in various states of growth and decay.

Hydrostatic pingos are initiated by the drainage of a deep lake in a continuous permafrost environment. Following lake drainage the unfrozen saturated sediments that were beneath the lake are aggraded by permafrost and the pore water is progressively squeezed out. It is this water which feeds the growth of the ice core. In general the size of the pingo will be governed by the size of the lake basin from which it grows and usually one pingo will grow in the centre of the former lake basin. If the lake which drains has an irregular form and has two or more basins then one pingo may form in each of the basins.

Studies of the internal structure of hydrostatic pingos in the Tuktoyaktuk Peninsula have shown that beneath the pingo ice core there may be found a pressurized sub-pingo water lens and it is this water which feeds further growth of the pingo. Once such a pingo has become established there is little increase in its diameter and all subsequent growth is upwards, increasing its height. Similar details have not been proven for hydraulic pingos.

One of the longest surveys of a growing pingo was conducted on Ibyuk Pingo in the Tuktoyaktuk Peninsula (Mackay 1998). Although this large pingo was already some 47 m high at the initiation of the survey, the pingo was still seen to grow higher at an average rate of 2.7 cm per year during the survey period 1973–1994. Using this growth rate, along with other data concerning the geomorphological evolution of the area, suggests that Ibyuk may be of the order of 1,000 or more years old (Mackay 1998).

Growth of the pingo ice core leads to the progressive stretching of the overburden causing it to fail through cracking (the generation of dilation cracks) and slumping. The sediment cover at the summit will be thinnest due to this stretching and slumping and pronounced radial cracks may form here. The thinning and rupture of the thermally protective overburden will lead to the decay of the ice core and this will often result in the development of a crater at or near the summit which may contain a pond in summer. When pingos ultimately collapse, whether in a permafrost environment or due to climate change which sees the decay of the permafrost, they do so from the top down and invariably leave a circular or oval rampart surrounding a depression which may contain a pond or marshy area.

Since pingos only form in a permafrost environment, evidence of their previous existence can be used to infer the former presence of permafrost and hence they are extremely useful for palaeoclimatic reconstruction. The remains of pingos of Pleistocene age are often referred to as 'relict pingos' and such features have been documented from North America and western Europe. That these features have always been correctly interpreted, however, is still a matter of some dispute.

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SEE ALSO: ice wedge and related structures; palsa

STEPHEN D. GURNEY

PINNING POINT

Pinning points are topographic constrictions at which glaciers halt during advances or retreats. They are places where troughs shallow and/or narrow, bifurcate, join another valley or bend sharply. They operate at a range of scales; the fluctuations of CALVING GLACIERs in FJORD systems and ice-contact lakes are sensitive to pinning points, and the topography of land masses and continental shelves affects the behaviour of the floating extensions of ICE SHEETs (ice streams and ice shelves). At non-calving GLACIERs ice is lost primarily through melting so that ice losses are closely tied to climate change. At calving glaciers, however, calving may represent a significant proportion of total ablation, yet calving is only indirectly affected by climate. Calving rates increase with water depth and with the cross-sectional area of the calving terminus, so the mass of ice lost through calving is determined primarily by these non-climatic factors. This means that the fluctuations of calving glaciers can be largely controlled by the topographic geometry of the valley. At pinning points calving rates are reduced, and they therefore represent places of enhanced stability where ice losses are balanced by ice supply. If a glacier retreats from a pinning point into deep water it must continue to retreat until calving rates decrease to match ice supply at a pinning point upstream. Equally, a calving glacier may be unable to advance during periods of regional glacier growth if such an advance would take its terminus into deep, open water. Glacier stillstands are common at pinning points and large MORAINEs may be constructed at these locations. Because these halts are determined by topography and not by climate, such moraine systems may have limited palaeoclimatic significance.

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SEE ALSO: mass balance of glaciers

CHARLES WARREN

PIPE AND PIPING

Natural soil pipes are linear voids formed by flowing water in soils or unconsolidated deposits (Plate 89). They occur throughout the world and vary from a few millimetres to several metres in diameter (Figure 122). Attempts have been made to provide a quantitative distinction between pipes and other soil macropores based on size, but none has been entirely satisfactory. The most fundamental property of soil pipes is that they actively drain water through the soil, which means that ‘connectivity’ and a drainage outlet are generally more critical than size in defining a pipe.

As pipes develop, they tend to create subsurface drainage networks akin to surface streams. Horizontal networks up to 750 metres long with

Plate 89 Pipe outlets in a riverbank in the English Peak District

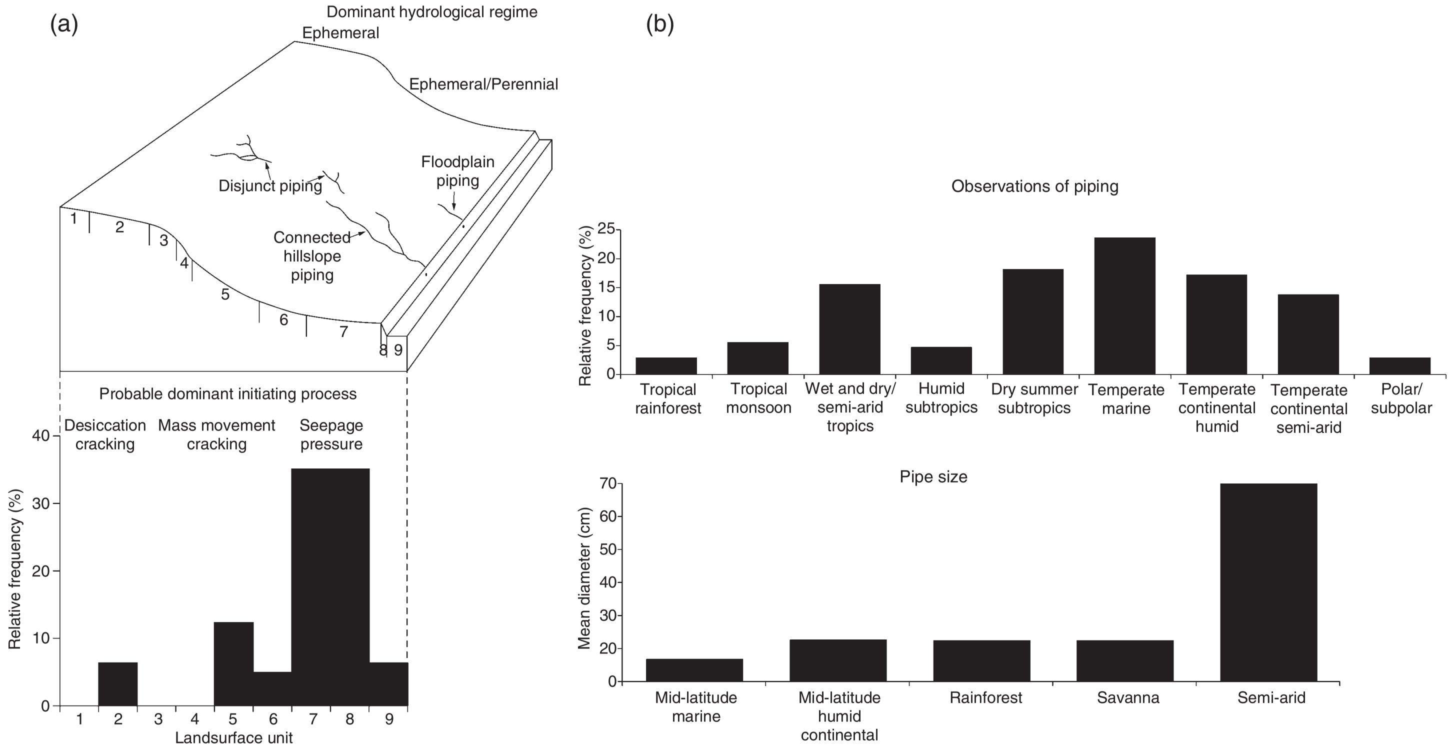

Figure 122 Geomorphic and climatic distribution of pipes: (a) frequency of piping in different landsurface units (Conacher and Dalrymple's (1977) NULM classification); (b) frequency and mean size of pipes in different climatic zones

many branching tributaries, reminiscent of dendritic stream networks, have been reported in shallow upland soils in Britain. In contrast, in badlands or deep loess deposits the networks tend to be more three-dimensional, as in the Loess Plateau in northern China or badlands in Alberta, Arizona, South Africa and Spain. Sometimes these consist of a number of horizontal networks formed above impeding layers with lower permeability and greater resistance to erosion linked by pipes eroding through the layers.

Pipes may be both a cause and a product of GULLYing. Gullies or channels may be formed when pipe roofs collapse. Pipe collapse provides an alternative to the 'classical' theory of channel extension based on headcutting by OVERLAND FLOW entering open channels (Jones 1987a). Conversely, gullies may trigger pipe formation by increasing the surrounding hydraulic gradient. American engineers have been concerned about pipes causing riverbank collapse (Hagerty 1991a,b). In such cases, traditional tests of bank stability based on the properties of the bank material may not give a good indication of risk, as the triggering mechanism may owe more to the source of pipeflow many tens of metres away.

The relationship between piping and LANDSLIDES is equally ambivalent. Pipes may initiate slides if they become blocked or water pressures exceed a critical threshold. Conversely, many peat bogs in Ireland that have well-developed piping do not display the periodic 'bog burst' phenomenon found in those lacking pipes. The pipes may prevent the buildup of water pressure. Again, pipes can develop in mass movement cracks following landslides.

These relationships indicate that shallow subsurface erosion processes can be significant agents in landscape development, sometimes with cyclical collapse and renewal. The term 'pseudo-karst' has been used for such landscapes, because the KARST-like features are predominantly formed by mechanical erosion rather than solution. The term SUFFOSION aptly evokes the mechanical winnowing and scouring. Clearly, the role of subsurface erosion is still undervalued in geomorphology, particularly in modelling.

The mechanics of pipe initiation were first defined by Karl Terzaghi in the founding years of the science of soil mechanics, because piping was the cause of many failures of earth dams. The process begins when seeping water produces sufficient force to entrain material at the seepage outlet point, which may be on a hillslope, a cliff face, riverbank or dam toe slope. The essential feature of Terzaghi's mechanism is that the *water pressure* renders the soil particles or aggregates weightless by counterbalancing gravity, so that the soil seems to 'boil' away (Terzaghi and Peck 1966). Seepage is increasingly focused on the head of the hole as it eats back into the bank. This may be called 'true' piping in the engineering sense. It is quite possible, however, for piping to begin at the upslope end, even on earth dams where it has been called 'rainfall erosion tunnels'. This second process involves progressive expansion of an existing conduit mainly through the shear stress exerted by the flowing water. Cracks caused by desiccation are commonly exploited by rainwater, which preferentially selects and erodes those cracks that run downslope and keeps them open underground even after re-expansion closes the cracks at the surface. Mass movement cracks, tree roots and animal holes can also be exploited. This process has often been distinguished as 'tunnel erosion'. However, the exact mode of initiation may not always be apparent from looking at the resulting landforms in the field. Indeed, in many cases the two processes interact in a very complex manner, as even those who have advocated distinguishing between them admit (Dunne 1990).

Reports of pipes and tunnels from almost every climatic region of the world clearly demonstrate that there is no single, unique set of initiating factors. One of the few truly universal contributory factors is a sufficient water supply to create the necessary pressure or shear stress. This water supply may be very spasmodic and some of the larger pipes occur in drylands, where short, intense rainstorms or the occasional rapid snowmelt event are the main cause.

Steep hydraulic gradients are also commonly needed, typically generated by a combination of water surplus and local relief. Piping is therefore most common where there is a high local relative relief, be it on upland hillslopes, in deeply incised badlands or simply in a riverbank.

Impeding layers within or beneath the soil, which concentrate vertical seepage and divert flow horizontally downslope, are also amongst the most universal preconditions. These are most effective where the layer conducting the flow is more erodible and/or contains pre-existing macropores, especially desiccation cracks, to speed the flow. There is a clear distinction here between the most common initiating factors in humid lands and those in the drylands. Whilst prior cracking aids pipe development in both climatic realms, most dryland piping is linked with dispersible soils, which are not common in humid lands.

Individual soil particles are more easily washed away than soil aggregates. Soil dispersal is predominantly a chemical process. It generally depends on the three-way balance between the total concentration of dissolved cations in the water flowing through the soil, the percentage of soluble or exchangeable sodium in the cation content of the soil itself and the type of clay minerals. In soils bonded by clays, concentrated salt solutions or weakly soluble salts like gypsum, dispersion may be accomplished by the dilution of the bonding salts by seepage water with low salinity. Alternatively, it may be produced by the chemical exchange of cations between the water and the soil, especially the replacement of divalent cations like calcium with monovalent cations like sodium in the soil. Sodium increases the repulsive forces between mineral particles and is the main dispersing or deflocculating agent. However, deflocculation does not necessarily increase erodibility and pipe development. Deflocculated soils may be stable if the sodium content of the soil is high enough to so thoroughly disperse the soil that permeability is reduced below a critical level, and to so reduce the structural stability of the soil that any voids that do develop quickly collapse and fill in. Experimental studies have identified an area of potential instability, swelling clays and deflocculation where piping is likely to be initiated in the transition zone between stable dispersion (high soil sodium and low salinity water) and stable flocculation (low soil sodium and saline seepage water). The exact boundaries between these zones vary according to the predominant clay mineral species. Montmorillonite is generally the least stable, with the broadest zone of instability. Nevertheless, evidence from earth dam engineers in America has revealed that the situation can be rather more complex, and many cases of piping have been observed in earth dams that fall well within the so-called stable zones (Sherard and Decker 1977).

Clay minerals affect the mechanical properties of the soil as well as dispersibility. Minerals of the smectite group like montmorillonite are noted for their higher rates of expansion and contraction. This increases susceptibility to desiccation cracking and infiltration rates. However, the expansive and dispersive properties of montmorillonites depend on the variety of the mineral. Sodium montmorillonite is more expansive than calcium or aluminium montmorillonite, and montmorillonites with cation substitution in their tetrahedral layer rather than their octahedral layer will not display the usual dispersive properties.

The minimum rate of seepage required to initiate piping depends on many properties of the soil, e.g. a threshold of $0.1\,\mathrm{mm\ s^{-1}}$ in non-dispersive silts against only 0.001 in dispersive clays. Soils that contain impeding horizons, are subjected to periodic desiccation and intense rainstorms, have high susceptibility to cracking, especially clayey or organic soils, and/or contain highly erodible, especially dispersible, layers are the most prone to piping.

Human interference has often increased the incidence of piping. Deforestation and devegetation decrease evapotranspirational losses, increasing water surplus. They expose soils to more desiccation and reduce the stabilizing effect of roots. In these situations, piping generally combines with other processes of accelerated erosion causing land degradation and increasing flood hazard (Jones 1981).

Considerable research has been undertaken into methods of rehabilitating farmland damaged by piping in Australia and New Zealand (Crouch *et al.* 1986). These have successfully included planting trees and grasses with high evapotranspiration rates and deep, stabilizing root systems, as well as adding soil conditioners to improve crumb structure. Even so, agricultural land is still being lost to piping, especially in drylands and through over-irrigation, e.g. in Arizona and Spain. Nearly half the farmland in the San Pedro valley, Arizona, has been lost to piping (Masannat 1980).

Research on the hydrology of piping has shown that pipes can contribute significant amounts of water, especially to upland rivers. Nearly half of floodflow and baseflow in one Welsh headwater tributary is derived from ephemeral or perennially flowing pipes (Jones 1987b). The pipes generally flow when the water table rises to pipe level and their discharge is a variable mix of new rainfall and 'old' water pushed out by the new. Monitoring in Canada, Japan, China, India and Britain generally confirms their role in speeding runoff and increasing floodflows. Most contributions fall between 20 and 50 per cent of streamflow, though amounts vary considerably in both time and space, even within the same basin, and in some cases are insignificant (Jones and Connelly 2002). Comparison with the response characteristics of other hillslope drainage processes suggests that pipeflow falls between diffuse throughflow and saturation overland flow in both timing and volume.

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J. ANTHONY A. JONES

PLANATION SURFACE

Topographical surfaces which are nearly flat over longer distances are called in geomorphology 'planation surfaces'. Some relief is allowed, especially in the form of isolated residual hills, but otherwise slope gradients should be very low and drainage lines should not be incised. Ideally, planation surfaces should cut across bedrock structures.

Thus, the descriptive definition is a simple one, but the issue of planation surfaces is one of the more controversial in geomorphology. As the name implies, the state of low relief has been achieved in the course of planation of a formerly higher relief by means of various exogenic agents of destruction (Plate 90). There are at least a few points persistently disputed, including the nature of process, or processes, leading to planation, the meaning of planation surfaces in long-term landscape evolution, and the possibility of producing and maintaining flat surfaces without recourse to erosional processes acting over protracted time spans (Figure 123).

Several mechanisms have been proposed to account for the origin of near-level surfaces. Accordingly, specific types of planation surfaces are distinguished. These include PENEPLAINS formed by peneplanation (Davis 1899),

Plate 90 Planation surfaces ideally should truncate geological structures, as it is in the case of a coastal surface in the Gower Peninsula in south Wales. The actual origin of the surface, whether wave-cut or subaerial, is the subject of debate

pediplains formed by pediplanation (King 1953), etchplains formed by etchplanation (see ETCHING, ETCHPLAIN AND ETCHPLANATION) (Büdel 1957). Other, more localized modes of formation are planation by sea waves in the coastal zone, by frost processes in the periglacial environment (see CRYOPLANATION), by areal glacial erosion, or by ubiquitous salt weathering in some desert and coastal situations.

The discussion between protagonists of peneplanation and pediplanation is now to much extent historical. In short, the principal difference between the two models resides in the way of slope development. Peneplains develop primarily through downwearing, i.e. slope lowering. In the course of peneplanation divides are lowered, slopes become gentler with time, and the landscape is progressively graded towards BASE LEVEL. By contrast, in the pediplanation model slope retreat away from drainage lines, i.e. backwearing, plays a crucial part. Higher ground may persist for much longer than the peneplanation model would imply, but their areal extent diminishes in time as bounding escarpments retreat. In front of scarps gently inclined surfaces of PEDIMENTs form and then coalesce to form a regional, ever-growing planation surface of the pediplain type. Another difference is that pediplains are not necessarily graded towards base level and they may form stepped landscapes and develop simultaneously at different altitudes. In both theories, planation surfaces are the ultimate products of long-term landform development and need a long time to form, perhaps of the order of $>10^7$ my. Neither peneplanation nor pediplanation are geomorphic processes per se; rather, they include a variety of superficial processes, including fluvial erosion, surface wash and various categories of mass movement.

Etchplanation was initially seen as a specific variant of peneplanation, applicable to low latitudes, where deep chemical weathering is ubiquitous. However, it was shown later that the mechanism of planation is fundamentally different. Etchplains form in the subsurface, through rock decay which is intense enough to overcome local differences in rock resistance against weathering. This leads to the development of a planar boundary between weathered material and solid rock beneath (see WEATHERING FRONT). Subsequent removal of weathering products exposes the planar 'etched' topography, which now forms an etchplain. Etchplains are thus two-stage features, as opposed to peneplains and pediplains. Although almost featureless etchplains, fulfilling the descriptive criteria for a planation surface, have been described from several areas, the view seems to prevail now that long-term etching leads to diversification rather than to planation of relief. Many low-latitude surfaces of low relief are cut across the weathering mantle, but the hidden topography of the weathering front is much more varied.

As the residual topography rarely gives a clue to the mode of formation of surfaces of low relief, it is preferable to call them simply 'planation surfaces', without genetic connotations. Moreover, it is likely that plains of long geomorphic history have been shaped by various processes, alternating over time, hence they would be 'polygenetic' surfaces rather than any 'monogenetic' peneplains, pediplains, or etchplains (Fairbridge and Finkl 1980). For example, pedimentation may be a means of stripping products of deep weathering to expose an etched surface.

Marine action used to be a favoured mode of planation, and in the nineteenth and early twentieth centuries many flat surfaces, especially in Britain, were identified as 'abrasion surfaces', even if no marine sediments could have been demonstrated. Later studies have shown that wave-trimmed surfaces of regional extent are unlikely to exist, and abrasion platform would have only limited extent (King 1963). Demonstration of a subaerial origin for many surfaces previously claimed as of marine origin, undermined the concept even further.

Periglacial planation is supposed to be achieved by means of simultaneous action of frost weathering of bedrock, rock cliff development and retreat, and mass movement, chiefly solifluction. The resultant cryoplanation is essentially a variant of pediplanation, applied to high latitudes and the Pleistocene. As with marine planation, cryoplanation is now generally seen as unlikely to account for the origin of more extensive surfaces, which are mostly inherited from pre-Pleistocene times.

Extensive level terrains in the Canadian and Fennoscandian Shield were long believed to have been shaped by powerful glacial erosion exerted by consecutive ice sheets during the Quaternary. Later research has shown that areal glacial erosion is not as common as formerly thought (Sugden and John 1976). By contrast, remarkable flatness of basement surfaces is inherited from protracted pre-Quaternary subaerial development, whilst much of these surfaces are exhumed Precambrian features (Lidmar-Bergström 1997).

The recognition of the powerful role of salt weathering in low-lying desert environments has led to the proposal that this process may have been crucial in producing flat topography of some coastal plains or closed depressions such as Quattara in Egypt. The term 'haloplanation' has been suggested (Goudie and Viles 1997).

Planation surfaces occupied a central position in geomorphology in the days, when establishing DENUDATION CHRONOLOGY of a given area was considered the main objective of geomorphology and cyclic development of landforms served as a paradigm. The first step in geomorphic research was to identify planation surfaces in the present-day landscape, or more often their remnants surviving on divides after the landscape had been dissected. It was followed by recognition of

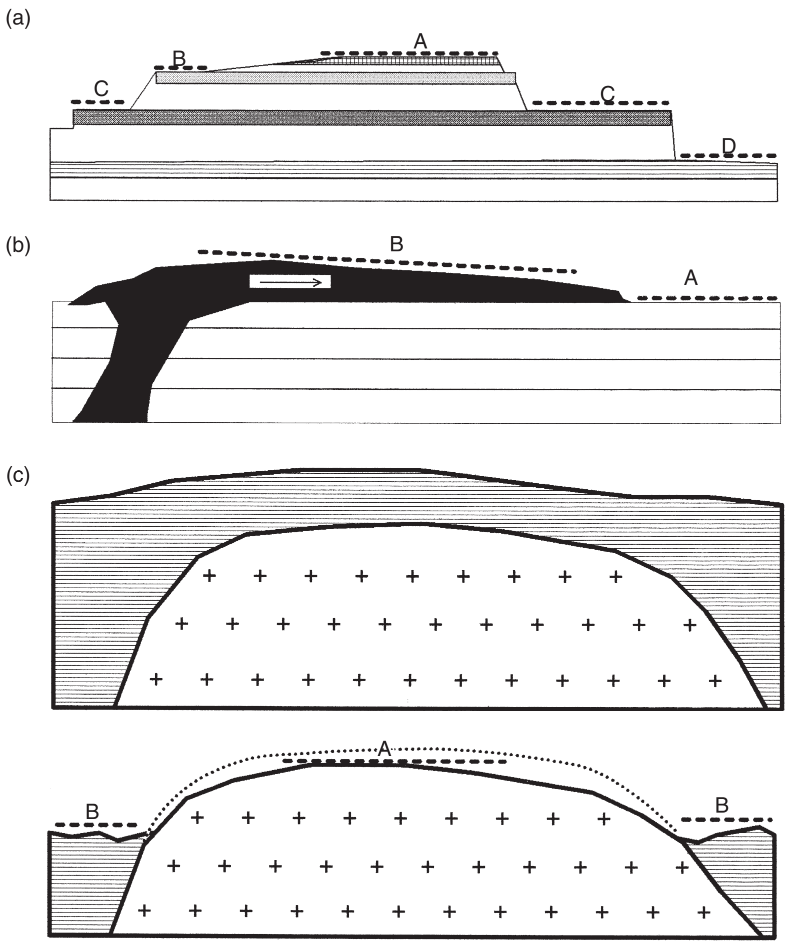

Figure 123 A diagram to show that not all plains are products of protracted planation. In areas built of flat-lying sedimentary rocks benches are distinctively controlled by structure (a). Flatness of surfaces underlain by lava flows may be largely primary and relate to the low viscosity of lava (b). The surface 'B', although located at higher altitude, is actually younger than the true planation surface 'A'. Surfaces of low relief common for many granite batholiths may be related to the original dome form assumed during emplacement (c). Different altitudes of surfaces cut in granite ('A') and country rock ('B') reflect different resistance of each rock complex rather than being indicators of different ages

altitude range of their occurrence, correlation of surfaces over wider areas and classification by height. The number of planation surfaces indicated the number of erosional cycles experienced by the landscape. As this approach frequently ignored influences differential tectonics may have had on landform development, relied on highly uncertain correlation procedures, and suffered from deficiency of accurate dating of surfaces, it began to be severely criticized in the 1960s and lost much of its popularity. Thus, whereas the issue of planation surfaces, their origin and chronology, features prominently in older regional studies, its importance has been greatly reduced in modern geomorphology. In addition, preoccupation with planation surfaces in historical geomorphology created an incorrect view that a search for ancient landscapes is essentially a search for planated relief. Recent work indicates that many palaeosurfaces preserved in geological and geomorphic record had a very complex topography, far from any state of advanced planation.

However, planation surfaces have retained significance in morphotectonic studies and are widely used as indicators of uplift and subsidence histories (Ollier and Pain 2000). By analysing spatial and altitude patterns of their distribution one can infer magnitudes of surface uplift, recognize direction of tilting and amount of warping, or locate fault zones in areas where conventional geological evidence is not at hand. In this context, the origin of a planation surface is usually not critical to the argument.

It is important to remember that the surface morphology alone may not give the clue to the reasons of flatness. Examination of geological structure underlying a flat topography can suggest origins alternative to the one implied by the name 'planation surface'. Moreover, if geological control can be demonstrated, tectonic quiescence is not a prerequisite any longer as inherent in the 'classic' modes of formation.

In many platform areas sedimentary strata lie horizontally over very long distances and topography may be adjusted to this negligible dip, especially if resistant rock layers occur at the surface. Likewise, backslopes of CUESTA ridges may show close adjustment to the dip of strata. In elevated plateaux there may exist several levels of 'planation surfaces' separated by escarpments, but in reality these are structural surfaces, each following a more resistant layer.

In formerly volcanic areas, topography may be adjusted to the geometry of lava flows. Basaltic lava, because of its low viscosity, may extrude in sheets of more or less uniform thickness over large areas. Flatness of the top surface of a flow will be then inherited from the time of extrusion and cooling. In case of multiple flows, denudation may expose top parts of each major lava flow and produce stepped topography, reminiscent of a generation of planation surfaces of various ages. In fact, all benches have structural foundation and may all have similar ages.

Remnants of ancient planation surfaces have also been sought in areas underlain by igneous rocks, especially by granite, which indeed often display a gently rolling topography or resemble laterally extensive, low-radius domes. A variety of structural controls is possible, including adjustment of form to the original roof of the intrusion, or to flat-lying joints.

That planation surfaces exist on Earth is indisputable. Examples are known from throughout the geological record, from Precambrian up to the present. In places such as the Fennoscandian Shield, Laurentide Shield or the Middle East, extensive surfaces of extreme flatness truncating various bedrock structures and disregarding differential rock resistance existed by the end of Precambrian. They were subsequently buried by Cambrian sediments and form sub-Cambrian planation surfaces, nowadays partially exhumed. Another generation of extensive planation surfaces evolved in the Mesozoic. Many upland areas are typified by the occurrence of surfaces of low relief at different altitudes, which probably have formed during the Cenozoic. It is the meaning, the mode of origin and age range of their formation which remain contentious and, paradoxically, underresearched issues.

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PIOTR MIGOŃ

PLATE TECTONICS

Plate tectonics is a unifying theory that explains many of the major features of the Earth's lithosphere such as VOLCANOes, rift (see RIFT VALLEY AND RIFTING) zones and mountain belts. The theory asserts that the lithosphere – the crust and uppermost mantle – is divided into rigid bodies or 'plates' that move horizontally and interact at their boundaries to produce these features. The theory evolved from the earlier concepts of continental drift and SEAFLOOR SPREADING.

In the eighteenth century it was noted that the coastlines of the Atlantic Ocean fit together like a jigsaw puzzle. In 1915 Alfred Wegener published several geological arguments to hypothesize that the continents had drifted apart from a supercontinent named Pangaea. He pointed to features that could be aligned by closing the Atlantic including Palaeozoic fold belts like the Appalachian and Caledonian mountains (Figure 124), metamorphic shields like northern Scotland and Labrador, major faults, palaeoclimatic indicators such as Carboniferous subtropical coal beds and tropical evaporites and desert sandstones, tillites from a Carboniferous ice cap with radiating striations and glacial erratics, and unique fossils of ferns and reptiles. Shortly thereafter he noted that seismic velocities in oceanic rocks were faster than in continental rocks, indicating that the less dense continents would float on top of oceanic rocks.

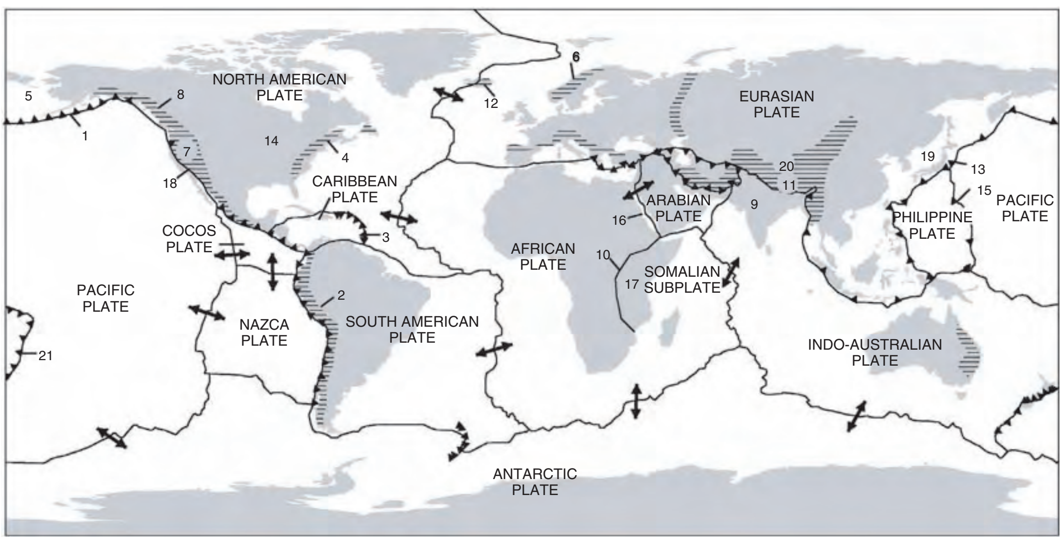

Figure 124 Plate boundaries of the Earth's crust. Features mentioned in the text: 1: Aleutian island arc; 2: Andes mountains; 3: Antilles island arc; 4: Appalachian mountains; 5: Bering Sea; 6: Caledonian mountains; 7: Columbia River basalts; 8: Cordilleran mountains; 9: Deccan basalts; 10: East African rift; 11: Himalayan mountains; 12: Iceland; 13: Japanese island arc; 14: Keweenawan aulacogen; 15: Mariana trench; 16: Mount Kilimanjaro; 17: Red Sea; 18: San Andreas fault; 19: Sea of Japan; 20: Tibetan plateau; 21: Tonga trench

Wegener's arguments were widely discredited for several decades. One exception was Alexander Du Toit who observed in 1937 that Pangaea split into the supercontinents of Laurasia (North America and Eurasia) and Gondwanaland (South America, Africa, India, Australia and Antarctica) in Triassic time, and that they split into the present continents in Jurassic time.

Geologic mapping increased dramatically worldwide in the 1950s, adding abundant evidence in support of continental drift. Critical geophysical arguments were added. For example, the advent of computers and related statistics enabled Bullard *et al.* (1965) to show the high probability of fit along the Atlantic continental margins. K-Ar radiometric age DATING METHODS and aeromagnetic anomaly maps confirmed the match of exposed and buried lithologies across the South Atlantic. Seafloor exploration confirmed that seafloor and continental rocks differed. Palaeomagnetic studies proved most persuasive. By measuring a rock unit's fossilized remnant magnetization in many samples, its location can be calculated relative to the Earth's palaeomagnetic pole with a known error. By comparing poles from coeval rock units on two continents, the relative rotation and displacement between them can be determined. Such studies statistically proved continental drift (Irving 1958).

Arthur Holmes in 1928 realized that radioactive decay heat would cause the Earth's mantle to convect and push continents along the Earth's surface. Dietz (1961) suggested that seafloor spreading occurred when convecting mantle magma intruded into the seafloor crust at mid-oceanic ridges, causing ACCRETION of new seafloor and carrying older seafloor away in both directions. Tests of the hypothesis followed, relying heavily on sonar maps of the submarine geomorphology and aeromagnetic maps of the seafloor that had been produced to track nuclear submarines. Quickly it was shown, on going perpendicularly away from a mid-oceanic ridge, that the youngest rock ages on oceanic islands increases linearly, that the K-Ar radiometric and fission track (see FISSION TRACK ANALYSIS) ages of seafloor basalts increase linearly, that the seafloor cools, and that the microfossils and tuffs in seafloor sediments record increasing ages – all supporting seafloor spreading. Most critical was the recognition that reversals of the Earth's magnetic field every 10^4 to 10^7 Ma were recorded in new lavas at the mid-oceanic ridges, which then spread away equally in both directions to create linear rift-parallel anomalies of increased and decreased magnetic field intensity. This analysis both confirmed the seafloor spreading hypothesis and provided a way to easily map and date the seafloor, which represents about two-thirds of the Earth's surface (Heirtzler *et al.* 1968). The plate tectonic theory evolved to explain how the Earth's crust is moving everywhere systematically.

The Earth's surface has seven large plates – the Pacific, Eurasian, North American, South American, African, Indo-Australian and Antarctic plates, five medium-sized plates (Figure 124), and an uncertain number of small plates in between. Plates move at rates up to about 20 cm yr^{-1} although the norm is only 5–10 cm yr^{-1}. This motion can be directly measured using very long baseline interferometry, satellite laser ranging, and satellite radiopositioning using the Global Positioning System's satellites. Past rates are measured using the spacing of seafloor magnetic anomalies and palaeomagnetic methods. Further, because crustal plates are fitted around a sphere, each plate actually rotates about its own Euler pole with a varying rate of motion across it. Plate boundaries have three types of ACTIVE MARGINs: rift zones, subduction zones and transform faults. Each may involve oceanic, continental or both types of crust and each combination of boundary and crust has its own typical topographic expression, dynamics and geologic characteristics.

Rift zones and ridges

Rift zones are linear zones of extension where new crust is added to the Earth's surface and where the motion is perpendicularly away from the lineament. Topographically, they have a central depression, typically 20 to 50 km wide and 1½; to 3 km deep, between uplifted plateaux that extend outwards for hundreds of kilometres on either side. Below the depression, convecting upwelling magma splits at the crust–mantle interface, pushing older crust upward and outward on inward-dipping normal faults to form the plateaux. Decreased lithostatic loading in the depression causes melting of the uppermost mantle to form a mafic magma. It intrudes along the rift lineament, particularly along its central axis, to form sheeted gabbroic intrusions at depth and basaltic lavas on the depression's floor, creating a new crust about 5 km thick. Because the crust is thin and the forces tensional, rift zones generate numerous relatively weak, shallow earthquakes.

On the seafloor, the rift zones are called ‘ridges’. They form a great continuous submarine mountain system through the world’s oceans that extends for about 60,000 km (Figure 124), but is exposed in only a few places such as Iceland. Typical mid-oceanic ridge basalts (MORB) are olivine tholeiites with only minor compositional variations. Erosion is minimal in the oceans, so that only a thin veneer of mostly chemical and biochemical sediments precipitate on the seafloor volcanics as siliceous and calcareous oozes. These sediments increase slowly in thickness progressively away from the ridge, typically at a rate of about 1 mm/10^2 yr. In the ridge depression, high heat flow from the Earth causes abundant geothermal activity. The hot hydrothermal fluids can precipitate elegant chimney structures around their vents that are rich in metal sulphides to form ore deposits. Also the fluids chemically alter the basalts and gabbros as they pass through and they create a marine microenvironment with a diverse range of exotic flora and fauna. Heat flow, seismic and gravity studies show that the seafloor crust progressively cools, increases in density and sinks, and thickens by underplating as magma freezes on its lower surface as it is pushed further away from the rift.

The East African rift zone is about 5,000 km long and exemplifies how the continental crust of a CRATON is split (Figure 124). Such rift zones are topographically and seismically similar to oceanic ridges. Because the bounding plateaux are exposed to WEATHERING, minor chemical sediments occur around hot springs only but EROSION produces abundant terrestrial clastic sediments that are deposited in the rift valley and the volcanic plateaux are progressively attacked by surficial weathering away from the valley. Tholeiite basalts sometimes fill and overflow the rift, producing large plateau lava sequences such as the Columbia River or Deccan basalts. Also, alkali basalt volcanism may occur in the adjacent plateaux through partial melting of continental crust, creating intrusive plugs, extrusive cinder cones and spectacular stratovolcanoes such as Mt. Kilimanjaro with its equatorial GLACIER peak. If the rift zone is truly active, the continental crust is entirely split in 10 to 20 Ma, the intervening crust becomes a mid-oceanic ridge, the sea invades as has happened in the Red Sea, and the continental edges become PASSIVE MARGINS.

Subduction zones

Subduction zones are convergent margins where the Earth destroys old crust to make room for the new crust created in rift zones. As two plates converge, one plate bends downward and is driven back into the Earth’s mantle at an angle to be remelted. A subducted slab can reach depth extents of 1,400 km long and reach a depth of 700 km. The plates may converge head on or at a substantial angle, resulting in strong compressive forces that produce numerous powerful earthquakes with foci along the depth extent. In fact, about 90 per cent of all earthquake energy is released in subduction zones so that their locations are well known from seismology studies. Three types of crustal collisions are possible: ocean-to-continent, ocean-to-ocean, and continent-to-continent.

When oceanic crust collides with continental crust, as is happening around most of the Pacific Ocean (Figure 124), the thin (5–10 km) denser (~2.9 g/cc) oceanic plate dives under the thicker (30–50 km) less dense (~2.7 g/cc) continental plate at a 30° to 70° angle. At the surface a trench is formed that may be several thousand kilometres long in plan with typical widths of 50 to 100 km and depths of 7 to 9 km. Thus the epicentres for shallow earthquakes occur at the trench. Most trenches are arcuate with their concave side facing the continent like the Aleutian trench, but some are straight like the Tonga trench. Any sediment from erosion on the exposed continent that reaches the trench’s bottom is subducted down into the Earth, so the trench never fills. However, such sediments do accumulate on the CONTINENTAL SHELF as TURBIDITY CURRENTs carry them into a fore-arc basin between the continental margin and the continent’s shoreline. Commonly the turbidites are compressed, deformed and metamorphosed in the plates’ collision to form a flysch belt along the coastline. Sometimes seafloor sediments and crust are obducted onto the shelf and preserved in the flysch. Inbound of the flysch, an ISLAND ARC forms. Heat, generated mostly by friction in the subduction zone, melts the overlying sialic and simatic rocks of the upper and lower continental crust and the mafic to ultramafic uppermost mantle, causing large diorite to granite intrusions to be emplaced into the old continental crust and mafic to felsic volcanics to be extruded onto it. Thus the world’s most spectacular volcanoes are found mostly around the Pacific Ocean. Where there is a wide continental shelf, this intrusive–extrusive rock complex forms a mountainous island arc with a submarine back-arc basin behind, like the Aleutian arc and Bering Sea or the Japanese arc and Sea of Japan. Alternatively, if the shelf is narrow, the complex forms a range of high coastal mountains. Further compression deforms and stacks the rocks behind the arc into sheets on thrust faults, forming high mountains like the Andes and Cordillera with their photogenic MOUNTAIN GEOMORPHOLOGY.

Collisions between oceanic plates are less common than oceanic–continental collisions but the process is similar except that both crusts are relatively thin and dense. Consequently their subduction zones descend a steeper angle approaching 90° and go deepest into the Earth, producing most of the world's deepest earthquakes. Also their trenches, like the Mariana trench, are deeper and reach depths of about 11 km. Further, only the peaks of the arc volcanoes reach the surface as a chain of basaltic islands, such as the Antilles, on the margin of the non-subducting plate. With so little land above sea level, very little clastic sedimentation occurs, but chemical sediments and CORAL REEFs often form around such islands in tropical climates to form barrier (see BARRIER AND BARRIER ISLAND) reefs and atolls.

Collisions between continental plates create the world's greatest mountains with spectacular TECTONIC GEOMORPHOLOGY, like the 10-km high Himalayan Mountains where the Indian subcontinent rides on the Indo-Australian plate and butts against the Eurasian continent and plate. The mountains are high because both crusts are relatively thick but less dense than the Earth's mantle and so they float to high elevation. Their combined 70-km thick crust sinks deeper into the Earth's mantle because of isostasy. Prior to collision, the continents had oceanic crust between them with either a passive margin like most of the Atlantic coastline or more commonly with an oceanic–continental collisional margin as described above. As the continents collide, some seafloor sediments, volcanics and gabbroic intrusions are often squeezed up and trapped between them where they are complexly deformed, and metamorphosed to serpentinites. Finding serpentinites defines the suture between the plates. On both sides of the suture zone are the deformed and metamorphosed remains of island arcs and back-arc basins first, then very high thrust-faulted mountains of stacked sedimentary sequences, followed by high foothills of folded sedimentary rocks, and then high plateaux such as the Tibetan plateau, with gently deformed strata that are deeply dissected by canyons. The suture zone and thrust-fault belts are tectonically active, producing strong shallow earthquakes mainly. Below the suture, subduction stops on closure. However, it takes tens of millions of years for the relict slab of descending crust to melt so that some minor earthquakes still occur at intermediate depths. Closure also terminates most volcanism because frictional heating ceases in the subduction zone.

Transform faults and triple junctions

J. T. Wilson's (1965) description of a new type of fault, a transform fault (Figure 125), was the key to understanding the dynamics of plate tectonics. The motions of rift and subduction zones were well understood, but the seafloor scarps were thought to be transcurrent faults where the two sides slid in opposite directions to offset a marker. Noting that these scarps cut the ridges at right angles and offset them, Wilson reasoned that the seafloor was moving away from both offset ridge segments in both directions so that the fault's sides were moving: (a) towards each other between the offset ridges as an active plate boundary, and (b) in the same direction and speed outside of the two ridges where they are tectonically

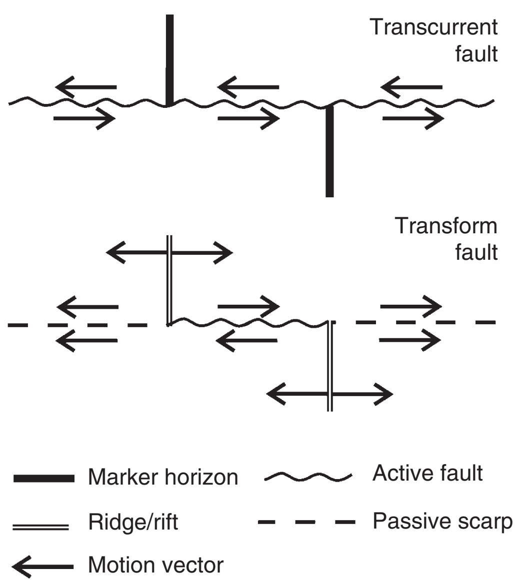

Figure 125 Dynamic comparison of transcurrent and transform faults

passive and inside the plates (Figure 125). He further explained how the transform fault motions interacted with ridge and subduction zone motions to create plate boundaries.

On the seafloor, transform fault lineaments can be over 3,000 km long and 3 km in height, being highest on the side of the closest ridge segment. Excluding very minor amounts of active volcanism, the Earth's crust is neither created nor destroyed. Between the ridges, horizontal shear generates weak to moderate, shallow earthquakes that are rarely destructive. In a few places, such as the San Andreas fault, the active part of a transform fault cuts much thicker continental crust, causing periodic powerful and highly destructive shallow earthquakes as the two sides grind past each other. Minor alkalic volcanics, sag ponds and other geomorphic features may occur along such faults.

Active plate boundaries form triple junctions where three plates meet, separated by three boundaries or arms. Each arm may be a rift, subduction zone or transform fault to make sixteen possible combinations. For example, the Pacific–Cocos–Nazca triple junction has three radiating rifts as arms whereas the Pacific–Cocos–North American triple junction has a rift, a subduction zone and a transform fault as its arms. The geometrically important fact is that the velocity and direction of each plate at a triple junction must form a spherical vector triangle. Thus, if the rotational motions of two plates are known from seafloor magnetic anomalies or palaeomagnetism, the rotational motion and Euler pole of the third plate can be calculated.

Importance of the theory

Understanding plate tectonics has led to a revolution in the Earth sciences. Virtually all of its supporting evidence relates to the last 200 Ma or so but, invoking UNIFORMITARIANISM, it has provided the basis for geographers to understand how the Earth's landscapes developed and for geologists to interpret the preceding 4,000 Ma of the Earth's rock record. For example, the Caledonian and Appalachian mountain belts are now recognized to be the deformation product of three Palaeozoic collisional events with oceans opening and closing in between in WILSON CYCLES. Similarly, geologists and geophysicists are using the theory to fit Precambrian supercontinents together such as Rhodinia, to identify failed rift valleys such as the 1,100 Ma Keweenawan aulacogen, to explain how geologic terranes that formed in radically different tectonic settings are now abutting, and to explain epeirogenic vertical motions in the continental interiors to form basins and plateaux. Finally, knowing how plates have moved over the past 200 Ma is enabling geophysicists to constrain models to investigate the Earth's magnetic field and to understand convective motions in its interior.

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D.T.A. SYMONS

PLOUGHING BLOCK AND BOULDER

Ploughing blocks and boulders are individual boulders that move downslope faster than their surrounding material by processes related to seasonal frost. Their movement ranges from millimetres to a few centimetres a year and is restricted to the annual freeze–thaw cycle (Ballantyne 2001; Berthling *et al*. 2001a). Due to the differential movement, the boulder pushes up a mound against its downslope side while leaving a depression along its upslope track. Ploughing boulders belong to the area of PERIGLACIAL GEOMORPHOLOGY. They are developed on slopes in the warmer part of the periglacial belt, commonly together with solifluction lobes. Only a few detailed process studies exist, but these indicate that boulder movements are caused by the same processes that occur in SOLIFLUCTION. During autumn and winter, the boulders protrude above the snow cover for some time. Combined with differences in thermal conductivity, this causes more intensive heat loss through the boulder than through ground elsewhere. This results in favourable conditions for ice segregation (Ballantyne 2001) and causes FROST HEAVE of the boulders. A heave of up to 7.5 cm has been demonstrated from southern Norway (Berthling *et al*. 2001b). It was shown that boulder heave stopped in midwinter regardless of snow conditions. Depletion of soil moisture might explain this behaviour. During spring and early summer, the boulders melt out of the snow and the soil beneath the boulder starts to melt earlier than the surrounding ground. Consolidation rates of up to 4.2 mm/day through a six-day period were measured by Berthling *et al*. (2001b). If melting of ice is more rapid than the ability of the released water to escape, water is trapped beneath the boulder so that the porewater pressure rises. This causes the soil beneath the boulder to lose strength and downslope deformations may occur. The process is referred to as gelifluction, and is the main cause for ploughing boulder movement. A second process that has been invoked is frost creep. Frost creep results from frost heave normal to the freezing plane and settlement along the vertical. This results in a net downslope movement in the cases where the freezing plane is essentially parallel to the sloping ground surface. The frost creep model in its simple form should be abandoned, as the boulders heave and tilt in directions determined by variations in heat removal, frost susceptibility and soil water content beneath the boulder, not slope. Yet, instability caused by tilting during frost heave might induce some displacements during thaw.

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SEE ALSO: frost heave; periglacial geomorphology; solifluction

IVAR BERTHLING

PLUVIAL LAKE

Bodies of water that accumulate in basins as a result of former greater moisture availability resulting from changes in temperature and/or precipitation. The study of pluvial lakes developed in the second half of the nineteenth century. Jamieson (1863) called attention to the former greater extent of the great saline lakes of Asia: The Caspian, Aral, Balkhash and Lop-Nor and Lartet (1865) pointed to the expansion of the Dead Sea. The term pluvial appears to have first been applied to an expanded lake by Hull (1885), but was originally applied by Tylor (1868) to valley fills in England and France. A major advance in the study of pluvial lakes came in the western USA with the work of Russell (1885) on Lake Lahontan and of Gilbert (1890) on Lake Bonneville. A discussion of these early studies and their bibliographic details is given in Flint (1971: Chapter 2).

The Great Basin of the USA held some eighty pluvial lakes during the Pleistocene, and they occupied an area at least eleven times greater than the area they cover today. Lake Bonneville (Plate 91), was roughly the size of present-day Lake Michigan, about 370 m deep and covered 51,640

Plate 91 A group of high shorelines that developed in the Late Pleistocene around pluvial Lake Bonneville, Utah, USA

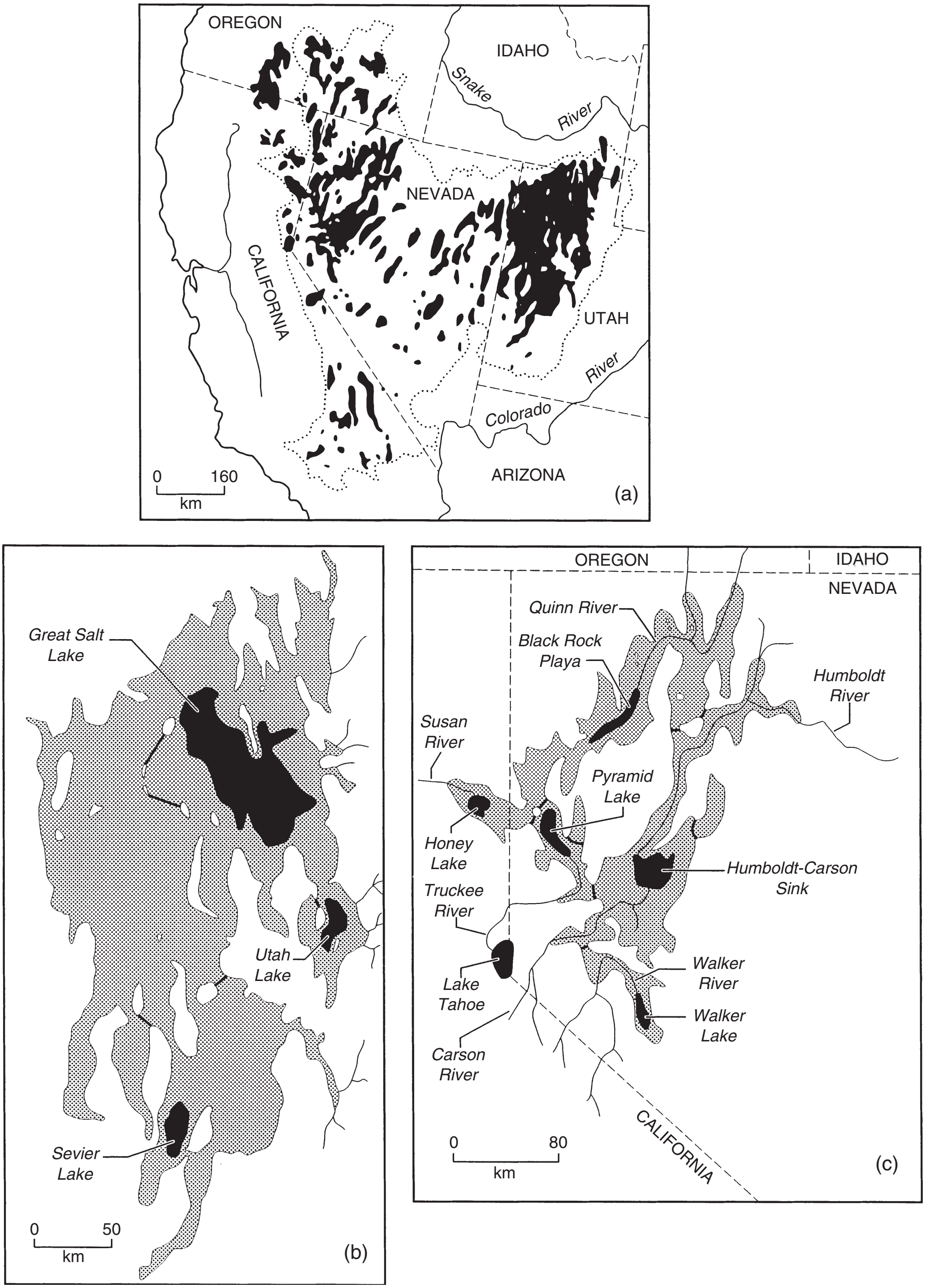

Figure 126 (a) The distribution of Pleistocene pluvial lakes in the southwestern USA; (b) Lake Bonneville; (c) Lake Lahontan km^2. Lake Lahontan was rather more complicated in form, covered 23,000 km^2, and reached a depth of about 280 m at the site of today's Pyramid Lake (Figure 126). It covered an area nearly as great as present-day Lake Erie. River courses became integrated and lakes overflowed from one sub-basin to another. For example, the Mojave River drainage, the largest arid fluvial system in the Mojave Desert, fed at least four basins and their lakes in pluvial times: Lake Mojave (including present-day Soda and Silver lakes), the Cronese basin and the Manix basin (which includes the Afton, Troy, Coyote and Harper sub-basins; Tchakerian and Lancaster 2002). Also important was the Owens Lake–Death Valley system.

A very large amount of work has been done to date and correlate the fluctuations in the levels of the pluvial lakes. Much of the early work is reviewed by Smith and Street-Perrott (1983). They demonstrated that many basins had particularly high stands during the period that spanned the Late Glacial Maximum, between about 25,000 and 10,000 years ago. More recently there have been studies of the longer term evolution of some of the basins, facilitated by the study of sediment cores, as for example from Owens Lake, the Bonneville Basin, Mono Lake, Searles Lake and Death Valley.

The high lake levels during the Last Glacial Maximum may well be the result of a combination of factors, including lower temperatures and evaporation rates, and reduced precipitation levels. Pacific storms associated with the southerly branch of the polar jet stream were deflected southwards compared to the situation today.

Other major pluvial lakes occurred in the Atacama and Altiplano of South America (Lavenu *et al*. 1984). The morphological evidence for high lake stands is impressive and this is particularly true with regard to the presence of algal accumulations at high levels (as much as 100 m) above the present saline crusts of depressions like Uyuni (Rouchy *et al*. 1996). There is a great deal of variability and confusion about climatic trends in the Late Quaternary in this region, not least with respect to the situation at the Late Glacial Maximum and in the mid-Holocene (Placzek *et al*. 2001). Nonetheless, various estimates have been made of the degree of precipitation change that the high lake stands imply. Pluvial Laguna Lejíca, which was 15–25 m higher than today at 13.5 to 11.3 Kyr BP and covered an area of 9–11 km^2 compared to its present extent of 2 km^2, had an annual rainfall of 400–500 mm, whereas today it has only around 200 mm. Pluvial Lake Tauca, which incorporates present Lake Poopo, the Salar de Coipasa and the Salar de Uyuni and which had a high stillstand between 15 and 13.5 Kyr BP, had an annual rainfall of 600 mm compared with 200–400 mm today.

In the Sahara there are huge numbers of pluvial lakes both in the Chotts of the north, in the middle (Petit-Maire *et al*. 1999) and in the south (e.g. Mega-Chad). In the Western Desert there are many closed depressions or playas, relict river systems and abundant evidence of prehistoric human activity (Hoelzmann *et al*. 2001). Playa sediments contained within basins such as Nabta Playa indicate that they once contained substantial bodies of water, which attracted Neolithic settlers. Many of these sediments have now been subjected to radiocarbon dating and they indicate the ubiquity of an early to mid-Holocene pluvial phase, which has often been termed the Neolithic pluvial. A large lake formed in the far north-west of Sudan, and this has been called 'The West Nubian Palaeolake' (Hoelzmann *et al*. 2001). It was especially extensive between 9,500 and 4,000 years BP, and may have covered as much as 7,000 km^2. If it was indeed that big, then a large amount of precipitation would have been needed to maintain it – possibly as much as 900 mm compared to the less than 15 mm it receives today.

In the Kalahari of southern Africa, Lake Palaeo-Makgadikgadi encompassed a substantial part of the Okavango Delta, parts of the Chobe–Zambezi confluence, the Caprivi Strip, and the Ngami, Mababe and Makgadikgadi basins. At its greatest extent it was over 50 m deep and covered 120,000 km^2. This is vastly greater than the present area of Lake Victoria (68,800 km^2) and makes Palaeo-Makgadikgadi second in size in Africa to Lake Chad at its Quaternary maximum. Dating it, however, is problematic (Thomas and Shaw 1991) as is its source of water. Some of the water may have been derived when the now DRY VALLEYs of the Kalahari (the *mekgacha*) were active and much could have been derived from the Angolan Highlands via the Okavango. However, tectonic changes may also have played a role and led to major inputs from the Zambezi.

In the Middle East expanded lakes occurred in the currently arid Rub-Al-Khali and also in Anatolia (Roberts 1983). In Central Asia the

Aral–Caspian system was hugely expanded. At several times during the late Pleistocene (Late Valdai) the level of the lake rose to around 0 m (present global sea level) compared to −27 m today and it inundated a huge area, particularly to its north. In the early Valdai glaciation it was even more extensive, rising to about +50 m above sea level, linking up to the Aral, extending some 1,300 km up the Volga River from its present mouth and covering an area in excess of 1.1 million km^2 (compared to 400,000 km^2 today). At its highest it may have overflowed through the Manych depression into the Black Sea. In general, transgressions have been associated with warming and large-scale influxes of meltwater (Mamedov 1997), but they are also a feature of the glacial phases when there was also a decrease in evaporation and a blocking of ground water by permafrost. Regressions occurred during interglacials and so, for example, in the Early Holocene the level of the Caspian dropped to −50 to −60 m below sea level.

Large pluvial lakes also occur in the drylands and highlands of China and Tibet and levels appear to have been high from 40,000 to 25,000 BP (Li and Zhu 2001). Similarly the interior basins of Australia, including Lake Eyre, have shown major expansion and contractions, with a tendency for high stands in interglacials (Harrison and Dodson 1993).

As can be seen from these regional examples, pluvial lakes are widespread (even in hyper-arid areas), reached enormous dimensions, and had different histories in different areas. Pluvials were not in phase in all regions and in both hemispheres (Spaulding 1991). In general, however, dry conditions during and just after the Late Glacial Maximum and humid conditions during part of the Early to Mid Holocene appear to have been characteristic of tropical deserts, though not of the south-west USA.

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A.S. GOUDIE

POINT BAR

Point bars are a form of river bar. They are located along the convex banks of river bends. They typically have an arcuate shape that reflects the radius of curvature of the bend. The cross-sectional slope of the bar is inclined towards the centre of the channel, reflecting the asymmetrical channel geometry at the bend apex. Textural attributes of the bar reflect patterns of secondary helical flow over the bar surface as the thalweg shifts to the outside of the bend at high flow stage.

Point bars are most commonly found along meandering rivers where there are clear genetic links between instream processes that form and maintain pool–riffle sequences, the channel morphology that results, the formation of point bars on the insides of bends, and resulting channel planform attributes (the meandering behaviour of the channel).

At bankfull stages, helical flow in bends carries sediment up the convex slope of point bars, while the concave bank is scoured. Sand or gravel bedload material is moved by traction towards the inner sides of channel bends via this helical flow. In this lateral accretion process, bedload materials are deposited on point bar surfaces. Lateral accretion deposits are detectable in the sedimentology of point bars by their oblique structures dipping towards the channel. Differing patterns of sedimentation are imposed by the radius of bend curvature (bend tightness) as well as the flow regime and sediment load.

Grain size typically fines down-bar (around the bend) and laterally (away from the channel). This produces a longitudinal 'around the bend' set of sedimentary structures comprising bedload material at the head of the bar, where the thalweg is aligned adjacent to the convex bank (at the entrance to the bend). As the thalweg moves away from the bend down-bar, lower energy suspended load materials are deposited. A mix of bedload and suspended load is generally evident at the tail of the bar. The most recently accumulated deposits are laid down as bar platform deposits at the bend apex. Typically these unit bar forms are largely unvegetated.

In many instances, point bars are compound features. These bank attached compound point bars comprise a mosaic of geomorphic units. In gravel situations, the bar platform is a relatively flat, coarse feature atop which a range of features are deposited or scoured. In the centre of the point bar is a gravel lobe. This likely represents the position of the shear zone during high flow stage. At high flow stage, compound point bars are dissected by chute channels, often with associated ramp deposits (McGowen and Garner 1970). Flow around vegetation produces a series of depositional ridges that have distinct grain size distributions (Brierley and Cunial 1998). The bar apex is a shallower feature inclined towards the channel.

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SEE ALSO: bar, river; channel, alluvial; fluvial geomorphology

KIRSTIE FRYIRS

POLJE

A form of large, flat-floored closed depression formed in KARST regions. They are a distinctive feature of limestone geomorphology. The term comes from the Slav word for field. Poljes are often covered with alluvium, are subject to periodical flooding, and they may have sinks, called *ponors*, into which streams may disappear. There is much debate as to the origin of poljes. They are probably polygenetic, with solution and tectonics playing important roles.

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A.S. GOUDIE

POOL AND RIFFLE

Pools and riffles are one of the most common and recognizable bedform sequences in channels. In the most basic sense, pools are classified as deep areas with low velocities at low stage, while riffles exhibit higher water-surface slopes and faster velocities. Pools and riffles exist in both alluvial and bedrock channels, but are best developed in gravel-bed substrates and meandering channels. Pools and riffles occur in moderate gradient channels with a transition to step-pool morphologies at higher gradients. These undulations in bed topography provide the primary framework for aquatic habitat in channels, and are of great interest because of their importance for macro invertebrates and fish species. The features also create tremendous form drag and flow resistance that may help rivers achieve equilibrium.

Because basic flow characteristics in pools and riffles change with stage (Richards 1976), limitations in the definition of pools and riffles exist. The zero-crossing method, where pools and riffles are recognized as residuals above a calculated mean bed profile, and spectral analysis provide robust methods for identifying the bedforms provided minimum size criteria are specified (Carling and Orr 2000). The residual-depth criteria or control-point method is used to define a pool based on the idea that riffles pond water in pools if flow ceases (Lisle and Hilton 1992). The residual pool is easily recognized in the field as the area that would be inundated, but it is still necessary to use minimum depth and width criteria to avoid subdividing a channel into extreme numbers of small morphologic units. Complications can arise because the residual depth of a pool is also influenced by sediment deposition, which can fill much of the pool at low flow. However, this fine sediment infilling can be used to estimate the sediment supply in a particular channel based on the idea that the residual volume of a pool prior to low-flow deposition is represented by the elevation of the coarse substrate below the overlying fines (Lisle and Hilton 1992). This assessment of pool sedimentation is particularly useful to evaluate land-use impacts on channels. It is also worth recognizing an important distinction between different types of pools where pools formed by some obstruction to flow are called forced pools and the remaining pools are termed free-formed pools (Montgomery *et al.* 1995).

The depth of pools and height of riffles is clearly maintained by some process, but the nature of the process is still debated. One idea that receives considerable attention is the velocity-reversal hypothesis. Keller (1971) used near-bed velocity measurements to show that velocities in pools are initially lower than those in riffles but increase at a faster rate and may exceed riffle velocities near bankfull stage. The water-surface elevations also change so that water-surface slopes equalized at high stage and may be steeper over riffles above bankfull level. The idea is closely related to the concept of two-phased bedload transport where low-flow deposition occurs in pools and high-flow deposition occurs in riffles (Jackson and Beschta 1982). Although the velocity-reversal hypothesis is often cited, it is frequently criticized based on continuity of mass concerns. However, recirculating eddies can form in some channels, increase velocities in pools and maintain continuity (Thompson *et al.* 1999). Meanwhile, the constrictions create backwater and locally elevated water-surface slopes in pools that can exceed water-surface slopes in riffles at bankfull stage. Evidence for recirculating-eddy enhanced velocity reversals exists, but flow routing of sediment around pools may dominate in other locations (Brooker *et al.* 2001). The combination of sedimentological properties and turbulence characteristics also is used to explain pool formation and maintenance. According to this idea, an obstacle to flow temporarily creates turbulence fluctuations, the perturbations to the flow generate the pool-riffle morphology, and differences in turbulence intensities and sediment characteristics between pools and riffles help to maintain disparities in sediment movement along the sequence (Clifford 1993). Much of the remaining work on pool formation and maintenance focuses on meandering channels and draws links between meandering and pool formation because pools tend to form at bends. Yang (1971) used this linkage and the idea that channels would adjust to minimize unit stream power along a channel in a theoretical approach with an equalization of water-surface slopes over pools and riffles at high stage. He concluded that the resulting formation process was a combination of dispersion and sorting of sediments. Studies also draw links between helical flow development and pool and riffle formation, but these processes are so clearly linked it is difficult to determine a casual relationship between them.

Pools and riffles exert an important influence on sediment sorting along a channel, especially the downstream end of pools, which create an uphill climb for particles moving downstream (Thompson *et al.* 1999). Sediment size, packing density and relative protrusion can all differ between pools and riffles. However, the combination of low-flow and high-flow sediment deposition can create a large variability in the size of bed sediments within a small area, and make it difficult to recognize distinct differences between pools and riffles (Richards 1976). Although channel-bed sediments in pools are often reported to be smaller than those in riffles (Clifford 1993), the opposite trend is reported in sediment supply-limited channels (Thompson *et al.* 1999). The disagreement in the general sorting trend probably results from two-phase bedload transport. During low flows, fine sediments can cover coarse substrate in pools along channels with high sediment loads, while supply-limited channels generally preserve the sediment-sorting patterns established at high flow.

Another fundamental characteristic of pools and riffles is the distance between successive pools or riffles, a measure termed the pool and riffle spacing. Values between five and seven average bankfull widths are often reported for pool and riffle spacing (Keller and Melhorn 1978), and this spacing has been attributed to reach-scale influences related to meander wavelengths in sinuous channels (Carling and Orr 2000). For example, variations in bed topography follow second-order autoregressive models as a result of a combination of periodic and random effects that may be related to meander wavelength (Richards 1976). Variation in spacing occurs because the channel-bed slope exerts a control on average spacing between pools (Wohl *et al*. 1993). Average spacing also varies due to the influence of obstructions to flow and variations in how pools are defined (Montgomery *et al*. 1995). In channels dominated by forced pools, local-scaling effects related to recirculating eddies behind randomly spaced channel constrictions can build morphologies with spacing values that agree with published values for a range of channel conditions (Thompson 2001). Therefore, it is unclear if reach-length or local-scaling effects create the semi-rhythmic spacing reported in natural channels.

Channel slope, channel-bed resistance and drainage area influence pool and riffle dimensions. Pool length and depth both tend to decrease with increased channel-bed slope (Wohl *et al*. 1993), and riffles become smaller in deeper water (Carling and Orr 2000). Given the fact that stream power increases with slope, the inverse relation between pool size and slope reflect changes in channel-bed resistance with more resistant beds associated with higher slopes. Pools also increase in size on larger channels, presumably because of the simultaneous increase in stream power with an increase in discharge on these larger systems. The relative magnitude of the bed undulations also tends to decrease with increased sediment supply because pools begin to fill and lose their distinct characteristics (Lisle and Hilton 1992).

As demonstrated by past research, pools and riffles will continue to be a central focus of research in fluvial geomorphology because of their important influence on both the physical and biological characteristics of natural channels.

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SEE ALSO: bedform; meandering; point bar; rapids; step-pool system

DOUGLAS M. THOMPSON

PORE-WATER PRESSURE

REGOLITH and highly fractured rock at Earth's surface (here termed soil) contain voids (pores) that are variously wetted or filled with water (pore water). Forces acting on pore water establish gradients of fluid potential, the work required to move a unit quantity of fluid from a datum to a specified position, and pore-water flows in response to these gradients. The concept of hydraulic head usefully describes pore-water potential. Total hydraulic head, or potential per unit weight of fluid, is usefully described in terms of gravitational, pressure and kinetic energy potential. The total hydraulic head (h) for an incompressible fluid (fluid having a constant density; ρ_w for water) is given by (Hubbert 1940):

$$h = \xi + \frac{p}{\rho_w g} + \frac{u^2}{2g}$$

where ξ is the gravitational, or elevation, potential; $p/\rho_w g$ is the pressure potential, in which p is the gauge pressure of the water relative to atmospheric pressure and g is gravitational acceleration in the co-ordinate direction; and $u^2/2g$ is the kinetic energy, or velocity, potential, where u is water velocity. Flow velocity in soil is usually very small, so calculations of hydraulic head in soils typically neglect velocity head. Pore-water pressure therefore constitutes one of two dominant components of the fluid potential in many soils.

Pore-water pressure is isotropic, but it varies with position relative to the water table (the depth horizon where pore-water pressure is atmospheric, which defines the zero-pressure datum) and with the proportion of soil weight carried by intergranular contacts. Below the water table, pore-water pressure is greater than atmospheric and positive; above the water table pore-water pressure is less than atmospheric and negative owing to tensional capillary forces exerted on pore water (e.g. Remson and Randolph 1962). If intergranular contacts carry all of the soil weight and water statically fills pore space, then the hydrostatic pore-water pressure (p_h) at a depth z normal to the soil surface (Figure 127) is given by

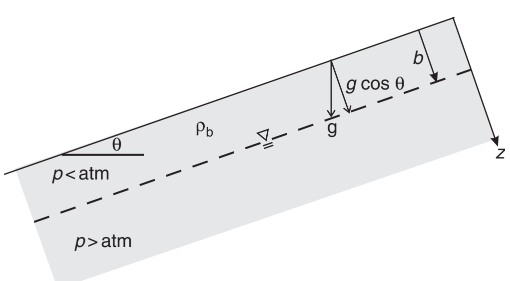

Figure 127 Schematic profile and definition of geometric parameters in an infinite slope having a water table at depth. Above the water table pore-water pressure is less than atmospheric; below it is greater than atmospheric

$$p_h = \rho_w g \cos\theta\,(z - b)$$

where b is depth to the water table and θ the slope of the soil surface. Pore-water pressure can exceed or fall short of hydrostatic under hydrodynamic conditions or if a soil collapses or dilates under load. If a soil collapses, it compacts. Below the water table this will cause a transient increase in pore-water pressure, the duration and magnitude of which are governed mainly by the rate of collapse and the permeability of the soil. The gauge pressure (p) at depth z can then be written as $p = p_h + p_e$, where p_e is a nonequilibrium pressure in excess of hydrostatic. If collapse thoroughly disrupts intergranular contacts, then the pore fluid may bear the entire weight of the solid grains, and the soil will liquefy. In that case, gauge pressure can be written as

$$p = \rho_w g \cos\theta(z - b) + \rho_b g \cos\theta\, b + (\rho_s - \rho_w)(1 - \phi)g\cos\theta(z - b)$$

where ρ_b is the moist bulk density of soil above the water table, ρ_s is grain density and φ is soil porosity. When a soil liquefies, the excess water pressure equals the sum of the unit weight of soil above the water table and the buoyant unit weight of the soil below the water table:

$$p_e = \rho_b g\cos\theta b + (\rho_s - \rho_w)(1 - \phi)g\cos\theta(z - b)$$

In unsaturated soil, such as occurs above the water table or (occasionally) when a saturated soil dilates (expands) under load, water does not fill pore space completely, and pore-water pressure locally is less than atmospheric. In that case capillary and electrostatic forces cause water to adhere to solid particles. As soil water content decreases, tensional forces increase and negative pore-water pressure bonds solid particles, increasing soil strength. The magnitude of negative pore-water pressure depends on soil texture and physical properties as well as on water content. Fine soils have a broader pore-size distribution and larger particle-surface area than do coarse soils. As a result fine soils have a greater range of negative-pressure potential because they hold more water than coarse soils, and the water bonds more tightly to particle surfaces. Piezometers are used to measure positive pore-water pressure; tensiometers commonly are used to measure negative pore-water pressure (e.g. Reeve 1986).

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SEE ALSO: debris flow; effective stress; landslide; liquefaction; mass movement; piezometric; undrained loading

JON J. MAJOR

POSTGLACIAL TRANSGRESSION

The postglacial transgression (see TRANSGRESSION) has had many names, depending from the area and the time period studied. The name of Flandrian stage was first proposed in the late nineteenth century to indicate the 'Campinian sands' of Flanders and of the Anvers' Campine. Dubois (1924) included in the Flandrian stage all sediments, marine and continental, that characterize the displacement of the shoreline from the sea-level minimum, corresponding to the last glacial maximum, to the present situation.

In the British Isles, on the other hand, the deposits called Flandrian correspond more or less to those of the Holocene period, i.e. to the last 10 ka (Shotton 1973). An Irish variant for the Flandrian is the Littletonian, the base of which is placed near the base of a peat dated about 10,130 BP in a core, or at the maximum of *Juniperus* pollen in another core.

In the Mediterranean, Blanc (1936) described in the Versilia Plain the stratigraphy of deposits, *c.*90 m deep, covering all epiglacial and postglacial times, that he considered equivalent to the Flandrian transgression. The term Versilian was subsequently applied to other Mediterranean deposits by several authors, following the chronostratigraphic meaning proposed initially by Dubois (1924).

In Japan the postglacial transgression has received several local names: Yurakucho transgression, after the name of a marine Holocene bed in the Tokyo area, Numa transgression, after the raised coral bed of the Numa Terrace in the Boso Peninsula (Naruse and Ota 1984), or Umeda transgression, corresponding to deposits found in the Kinki area. In uplifted Japanese regions, the higher part of the marine deposits are also called Jomon transgression, after the age of the older stage of the Japanese Neolithic culture (between *c.*9,400 and 2,800 BP) (Takai *et al.* 1963).

In west Africa, the Nouakchottian episode, defined by Elouard (1966), consists of shell deposits, a few decimetres to over 2 m thick, which are found along most of the coast of Sénégal and Mauritania, at elevations close or slightly higher than the present sea level (between −2 m and +2 or +3 m). These RAISED BEACH deposits date from 5,500 to 1,700 BP and follow a shoreline that formed several gulfs. In the Ndrhamcha Sebkha area, where the coastline is now rectilinear, the Nouakchottian gulf extended into the continent about 90 km. Other similar transgression stories, each with a different name, could be added from other coastal regions of the world.

The use of many names to identify the same phenomenon in various areas, in order to attempt correlations, may have been useful in the past, when precise dating tools were rare or unavailable, but seems not to be justified today. Radiochronology has shown that the last maximal glacial peak occurred about 18,000 radiocarbon years BP, i.e. after calibration, *c.*21 ka ago (Bard *et al.* 1990). A more general use of the terms 'postglacial transgression' (for the last 21 ka) or 'Holocene transgression' (for the last 10 ka) would certainly contribute to clarify and unify the international terminology.

A marine transgression may occur, however, not only with a rising sea level, but even with a falling sea level if sediment supply is depleted and erosion can occur; conversely, a regression of the sea often results from a sea-level fall, but may also occur with a rising sea level in the case of high sediment supply and coastal progradation. Transgression–regression sequences, i.e. lithostratigraphic evidence of marine deposits inter-fingered with freshwater or terrestrial sediments, usually correspond to major changes in sea level. However, when the inter-fingered layers are not continuous in space and time, interpretation should be careful, especially in the case of Late Holocene sediments deposited near river mouths or near plate boundaries where tectonic displacements may have taken place.

In postglacial times, especially in the mid- to late Holocene, interpretation of transgression

–regression sequences has been reported from several coastal localities, again with various names. This was the case, e.g. on the southern coasts of the North Sea, where within the upper part of the Holocene transgression, three Dunkerquian transgressions were distinguished above Calais deposits (Tavernier and Moorman 1954). For each transgression a former sea-level position was deduced from the present elevation of the deposits. Correlations between the Flemish Dunkerquian stratigraphic sequences and other local deposits were subsequently attempted by various workers in France, the Netherlands and Germany, giving rise to the reconstruction of sea-level histories showing oscillations of varying amplitudes. Nevertheless, some fifty years later, the precise amount of these Dunkerquian sea-level oscillations, as well as their existence, remains to be demonstrated.

As SEA LEVEL is concerned, the last deglaciation seems to have occurred mainly in two eustatic (see EUSTASY) steps, with a first warming period that peaked at the Bölling (about 13–12 ka BP) and a second warming period after about 11.6 ka BP, separated by a temporary cooling (Younger Dryas). The melting histories of the various ICE SHEETS were non-synchronous and the last deglaciation ended in each place when the former ice sheet had completely melted. This seems to have occurred around 10 ka BP in Scotland and for the Cordilleran ice sheet, close to 9 ka BP in the Russian Arctic, around 7.5 ka BP in Scandinavia and around 6 ka BP for the Laurentide ice sheet. In Antarctica and Greenland, only the outer part of the ice domes melted.

The melting of the ice sheets caused considerable glacio-isostatic (see ISOSTASY) vertical movements of the Earth's crust: mainly uplift in unloaded areas, and subsidence in wide peripheral areas around the former ice sheets. The load of melted water on the ocean floor caused the latter to subside (hydro-isostasy). This is expected to have caused a flow of deep material from beneath the oceans to beneath the continents, with possible reactivation of seaward flexuring at the continental edges. Part of the above isostatic movements were elastic, i.e. contemporaneous to loading and unloading. Part continued, at gradually decreasing rates, for several thousand years after loading or unloading had stopped, because of the viscosity of the Earth's material. Part of such isostatic vertical displacements is still going on.

When the subsidence in areas peripheral to former ice sheets took place under the sea, it increased locally the volume of the oceanic container. This caused hydrostatic imbalance, with an indraught of water from other areas, decreasing the sea-level rise or even producing a slight sea-level fall in oceanic regions far away from the influence of former ice sheets. It is generally towards 7 to 6 ka BP, with the ending of ice melting, that slight emergence of isostatic or tectonic origin started to occur in many areas. Combination of the above processes produced a variety of relative sea-level and postglacial transgression histories along the world's coastlines. According to Mörner (1996), during the last 5 ka relative sea-level changes have been affected also by the redistribution of water masses due to changes in oceanic circulation systems.

In former ice-sheet areas, where marks of past shorelines on ice vanished with ice flowing and melting, local sea-level histories can be reconstructed only after deglaciation. They usually show continuing regression/sea-level fall, though at decreasing rates, up to the present.

In peripheral areas to the former ice sheets, the maximum postglacial sea-level peak has not yet been reached and the postglacial transgression is more or less still going on. Finally, in areas remote from the ice sheets and in most uplifting regions, a sea-level maximum, now emerged at variable elevations, has generally been reached towards the mid-Holocene.

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P.A. PIRAZZOLI

POT-HOLE

Pot-holes are vertical, circular and cylindrical erosion features in consolidated rock of various lithologies. They are common in fluvial, fluviglacial and shore environments. Their sizes (diameter and depth) range from a few dm to a few m; but mega pot-holes many metres deep and wide also occur in formerly glaciated areas. Pot-holes are produced by abrasion, CAVITATION, dissolution and/or corrosion. A tool (pebble, sand) is necessary for abrasion whereas cavitation is a mechanical process of wearing created by a turbulent flow. Dissolution is active in carbonate rocks whereas corrosion, a more complex process, is manifested in most rock types, particularly in warmer climates. Many potholes have a complex origin. Shallow cavities made by cavitation or dissolution are often subsequently eroded through abrasion. Coastal pot-holes are less common than fluvial and fluviglacial. A few anthropogenic pot-holes on shore platforms have been reported in Brittany. The use of the term pothole for kettle and moulin should be avoided.

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JEAN CLAUDE DIONNE

PRESSURE MELTING POINT

The melting point is the temperature at which a solid changes into a liquid. The application of pressure to a solid depresses the melting point, and the melting point under a given pressure is referred to as the pressure melting point. For ice, the lapse rate of melting point with pressure is $-0.072\,°\mathrm{C\,MPa^{-1}}$, the effect of which, for example, is to depress the melting point to $-1.28\,°\mathrm{C}$ under 2,000 m thickness of ice.

The pressure melting point is significant in glacial geomorphology because of its effect on the interaction of glacier ice with its substrate. Ice exists on the surface of the Earth at temperatures very close to, and sometimes at, its melting point. The application of pressure to Earth surface ice, as a result of either the hydrostatic pressure of ice overburden or the interaction of moving ice with undulations in the substrate, can lead to the depression of the melting point sufficient to allow melting. The process of pressure melting on the upstream over-pressured side of bedrock bumps, and refreezing of water on the downstream under-pressured side, plays a major role in glacier motion, and is also significant in the entrainment and transport of subglacially eroded debris. This REGELATION process allows material generated by subglacial abrasion and quarrying to be entrained in re-frozen layers, and is one process by which debris-laden ice can be added to the basal layer of a glacier or ice sheet.

WENDY LAWSON

PRESSURE RELEASE

Many rocky outcrops show sets of horizontal or curvilinear fractures (SHEETING or EXFOLIATION structures) that are roughly parallel to the topographical surface. They are known with different names, some of them equivalent, such as pressure release, relief of load or offloading. It is obvious that all rock fractures are an expression of erosional offloading because only through the release of vertical and/or lateral pressure can the closed discontinuities become opened. But the gist of the pressure release concept is that rocks which cool and solidify deep in the Earth's crust (e.g. magmatic rocks), do so under conditions of high lithostatic pressure, i.e. loading by overlying materials (either rocks, sediments or even water or ice). So, many people suppose that when the rock outcrops in the Earth's surface it suffers a pressure release and this causes the development of stress and subsequent fractures parallel to the land surface. That is why the form of the land surface in broad terms could determine the geometry of the so-called pressure release fractures. According to this the fractures so generated would be secondary features (i.e. developed after the topography). But that is not always true because it is generally accepted that many plutons have been emplaced at shallow depths and the related structure was generated by the stresses imposed on magmas during injection or emplacement and, hence, so was the shape of the original pluton. Moreover, the so-called pressure release structures (i.e. sheet jointing) are well developed in rocks such as sedimentary and volcanic, which have never been emplaced, and even in granites, the magnetic orientation contemporaneous with the emplacement is clearly discordant to the pressure release structures. The pressure release theory may be questioned on several other grounds. Unloading appears to be mechanically incapable of producing fractures because if expansive stress developed during erosion, it would be accommodated along pre-existing lines of weakness and does not need to generate new ones, namely the sheet fractures. Another reason is that in fact several morphological and structural features developed on and in granitic rocks are incompatible with the tensional or expansive conditions implied by offloading. It is the case of structural domes, wedges and overthrusting associated with sheeting and is impossible to explain in terms of an extensional regime. Furthermore, evidence of dislocation and mylonitization along sheet fractures suggests that they are true tensional faults. Thus the pressure release structures may be better interpreted as primary features of the rock and accordingly the joints (see JOINTNG) were first developed in the bedrock and the shape of the land surface is a response to this previous internal structure.

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JUAN RAMON VIDAL-ROMANI

PRIOR STREAM

Prior streams are Late Quaternary PALAEOCHANNELS of the semi-arid Riverine Plain in southeastern Australia. These ancient rivers were first described in the scientific literature by Butler (1958), who was given the task of producing soil maps in a region set aside for expanded irrigated agriculture in the period following the Second World War.

The 77,000 km^2 Riverine Plain consists of the coalescing floodplains of westward-flowing rivers of the southern Murray–Darling system. Despite its exceedingly subdued topography and low surface elevation (the great majority is less than 100 m above sea level), the Plain displays a complex pattern of sediments, soils and micro-topography. At first, the apparently featureless nature of the landscape frustrated Butler's attempts to make sense of his field observations. However, with the aid of aerial photographs, it became clear that well-drained sandy linear depressions that stood a little above the adjacent plain marked the locations of ancient aggraded palaeochannels that Butler called prior streams.

Soil variation on the Plain was controlled by proximity to a prior stream. Well-drained calcareous soils on prior stream levees graded laterally into heavy clays on the distal floodplain. Beneath the prior stream channels were thick beds of pebbly sand. As Butler mapped the regional soil landscape in more detail he discovered that the prior streams formed a complex distributary pattern (Plate 92) that petered out to the

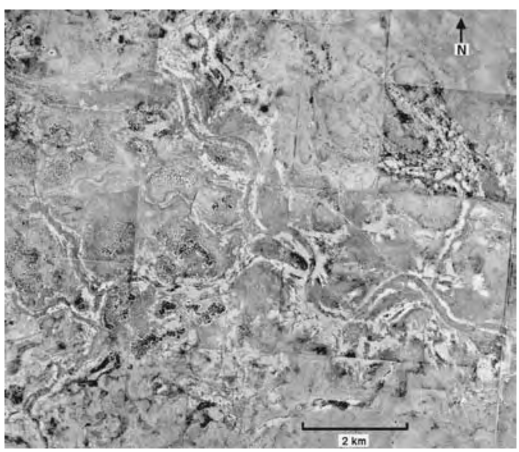

Plate 92 Air photograph mosaic of distributary prior stream channels on the Riverine Plain, Australia

west. Because the channels often intersected one another it was clear that there had been more than one period of prior stream activity. Butler invoked a cyclic model to interpret the different prior stream phases. Channel incision was thought to occur during more humid conditions when an absence of deposition permitted the development of well-organized soil profiles. The stable surfaces on which soils developed were called *groundsurfaces*. Channel aggradation occurred during more arid conditions when copious amounts of bedload sediment from upland catchment regions resulted in channel aggradation and extensive deposition.

Phases of prior stream activity

The oldest groundsurface described by Butler was the Katandra. It was thought to represent a long period of soil formation under more humid conditions than exist at present. A switch to more arid conditions resulted in a new phase of fluvial deposition (Quiamong) that progressively buried the Katandra surface. As aridity intensified vegetation breakdown in the region to the west led to widespread clay deflation by westerly winds and the deposition of an extensive blanket of pelletal clay (Widgelli PARNA) which mantled the earlier Quiamong deposits. As the peak of aridity passed parna deposition waned and renewed deposition by Mayrung prior streams occurred. This final phase of prior stream deposition was followed by a long humid phase when stream incision occurred and well-developed soils formed across the surface of the Plain. Soils of the Mayrung groundsurface developed on fluvial and aeolian parent materials of Quiamong, Widgelli and Mayrung age. A generalized summary of Butler's stratigraphic units is shown in Figure 128.

The modern channels of the Riverine Plain developed in the post-Mayrung period. They occupy narrow floodplain trenches incised two to three metres below the Mayrung surface and are characterized by high sinuosity, low width to depth ratio and a dominance of suspended sediment. According to Butler, these younger Coonambidgal deposits display very weak soil organization.

Post Butler

Not all workers agreed with Butler's interpretation of the prior stream deposits. The geomorphologist Langford-Smith (1960) argued that large meander wavelengths of the prior stream channels demanded greater discharges associated with late glacial pluvial, rather than arid, conditions. However, the absence of absolute dates on the prior streams (they all appeared to be beyond the radiocarbon limit of about 30,000 years) precluded any secure correlation with the glacial and interglacial episodes of the Late Quaternary.

Butler's early ideas were extensively revised during the latter part of the twentieth century. In the 1960s, Pels (1971) concluded that Butler's youngest stratigraphic unit, the Coonambidgal, was more complex than previously supposed. The early Coonambidgal phase was characterized by distinctive ancestral rivers that post-dated the prior streams and were the immediate precursors

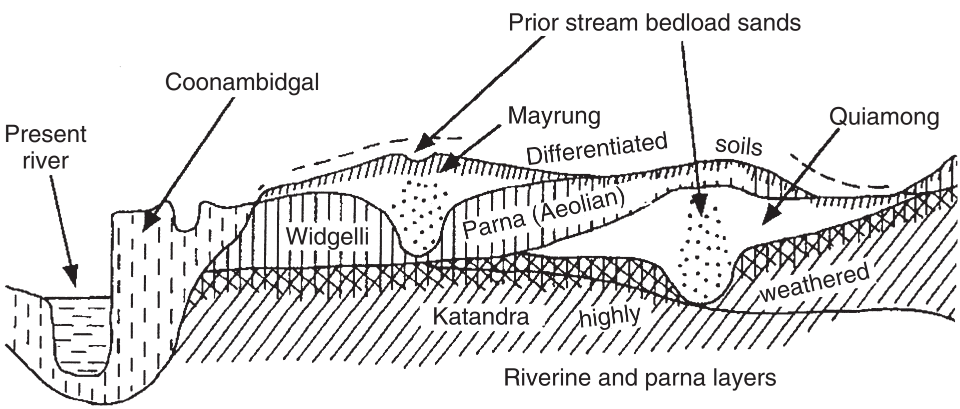

Figure 128 Generalized cross section of Riverine Plain showing Butler's (1958) prior stream sediments and soils (groundsurfaces)

of the modern drainage. The ancestral rivers were deep, sinuous, without levees and dominated by suspended load. They were much larger than the present rivers and maintained their courses across the Riverine Plain. Pels' model of Riverine Plain ancestral rivers and prior streams gained wide acceptance in the 1970s.

However, Bowler (1978), questioned the classification of ancient channels into two exclusive sequential types. He noted that both prior and ancestral attributes sometimes occur in different reaches of the same palaeochannel. In addition, Bowler found that some ancestral channels carried appreciable quantities of bedload, were bordered by sand dunes and mantled by well-developed soils similar to those of Butler's Mayrung groundsurface. Bowler concluded that Pels' separation of palaeochannels into two genetically different categories was unjustified.

Despite Bowler's misgivings, further progress on the nature and chronology of fluvial deposition awaited the development of thermoluminescence dating (Page *et al*. 1996). In brief, it was shown that four phases of palaeochannel activity occurred between approximately 100,000 and 12,000 years ago. Channel activity was characterized by alternations between sinuous, laterally migrating, mixed-load and straighter, vertically aggrading, bedload channel modes. Although the laterally migrating and vertically aggrading channel modes respectively approximate ancestral and prior streams, the sequence of channel activity was more complex than envisaged by Pels.

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KEN PAGE

PROGLACIAL LANDFORM

The literal meaning of 'proglacial' is 'in front of the glacier'; it is the area that receives the products of glaciation. The proglacial environment is complicated, especially where warmer glaciers produce more meltwater. It can include terrestrial environments, streams, lakes and the ocean. The deposits include moraines, large outwash fans, deltas, marine fans and thick packages of sediment deposited in the marine realm. Other than where streams rework and erode previously deposited material, this is a depositional environment. The proglacial zone moves with the ice edge. As the ice advances, the zone of proglacial deposition moves forward as well. As the ice retreats, the proglacial setting follows the retreating margin 'backwards' and former subglacially deposited sediment-landform assemblages (see SUBGLACIAL GEOMORPHOLOGY) will be partially eroded, redeposited and/or buried.

The use of the term 'proglacial' has not been consistent in the literature. It is clear that there is a transition between ice-marginal and proglacial fluvial or lacustrine/marine processes. Some have suggested a further transition between proglacial fluvial processes and paraglacial processes, defined as any non-glacial processes conditioned by glaciation (Ryder, 1971). On this definition PARAGLACIAL processes strictly subsume proglacial fluvial processes. Thus the terms 'ice-marginal', 'proglacial' and 'paraglacial' have been used in overlapping senses and there is no universally agreed set of definitions. Table 37 classifies proglacial environments according to distance from the ice margin and lists the associated landforms. As most of these landforms are covered by separate entries (see cross references in the fourth column of the table) proglacial landform assemblages in the following will be viewed at the larger scale of ice sheet systems, mountain valley systems and subaquatic landsystems. A detailed review of these three landsystems can be found in Benn and Evans (1998).

Ice sheet systems

The processes and patterns of proglacial deposition in front of ice sheets and ice caps are strongly influenced by glacier thermal regime. Temperate glacier margins are wet-based for at least part of the year. Meltstreams are the main agents of sediment transport and deposition. Ice-marginal forms are produced by alternating glacifluvial

Table 37 Classification of proglacial landforms according to different land-forming environments

Environment	Process	Landform	See also:
Terrestrial ice-marginal	Meltwater erosion	Ice-marginal meltwater channels	Meltwater and meltwater channel Urstromtäler
	Mass movement/ meltwater deposition	Ice-marginal ramps and fans Dump and push moraines Recessional moraines	Moraine Glacitectonics Kame Kettle and kettle-hole
	Glacitectonics	Composite ridges and thrust block moraines Hill-hole pairs and cupola hills	Ice stagnantion topography
	Meltwater deposition	Kame and kettle topography	
Subaquatic ice-marginal	Mass movement/ meltwater deposition	Morainal banks De Geer moraines	Glacilacustrine Glacimarine Moraine
	Meltwater deposition	Grounding-line fans Ice-contact (kame-) deltas	
	Debris flows	Grunding-line wedge	
Transitional from ice-marginal to fluvial	Meltwater erosion	Scabland topography Spillways	Meltwater and meltwater channel Outburst flood
	Meltwater deposition	Outwash plain (sandur) Outwash fan Valley train Pitted outwash Kettle hole/pond	Glacifluvial Kettle and kettle-hole
Transitional from ice-marginal to lacustrine and marine	Meltwater deposition/mass movement	Deltas	Glacideltaic Glacilacustrine Glacimarine
	Deposition from suspension settling and iceberg activity	Cyclopels, cyclopsams, varves Dropstone mud and diamicton Iceberg dump mounds Iceberg scour marks	

and gravitational processes, locally modified by glacitectonic deformation. In addition, extensive proglacial rivers may rework glacigenic sediments. In some cases, virtually all the evidence of a temperate ice lobe is in the form of glacifluvial sediments. At temperate glacier margins with moderate debris cover deposition typically produces small dump moraines, or push and squeeze moraines derived from sediment exposed on the glacier foreland. During deglaciation suites of recessional moraines are commonly formed. The areas between recessional moraines often exhibit well-preserved subglacial landforms. At temperate glacier margins covered by considerable quantities of glacifluvial sediment uneven ablation of sediment-covered ice can lead to the development of a karst-like topography with a relative relief of up to several tens of metres. Sediment deposited in supraglacial outwash fans and lakes produces a complex assemblage of landforms including ESKER systems, kame ridges and plateaux, and pitted outwash. Distinctive spatial associations of glacifluvial landforms deposited during ice wastage can be recognized: proglacial outwash passes upvalley into kame and kettle topography. Kame and kettle topography is locally refashioned into suites of river terraces by proglacial and postglacial streams.

Subpolar margins are characterized by cold-based conditions near the snout and an upglacier wet-based zone. Subpolar margins are commonly affected by glacitectonic processes. Meltwater is available on the glacier surface but not on the bed. It can have some impact on ice-marginal landforms, but in general these are only locally reworked into outwash deposits. Where thick accumulations of unconsolidated sediments are present, terminal moraines in the form of composite ridges are constructed by proglacial tectonics. Where proglacial thrusting does not occur, ice margin positions are recorded by frontal aprons built up from fallen debris. Upvalley end moraines or frontal aprons pass into ice-cored moraines of chaotic hummocky or transverse ridges and/or kames. In the cases of totally cold-based glaciers and where bedrock is close to the surface the amount of available debris reaches a minimum. Former glacier margins in such areas are often marked by lateral meltwater channels cut into bedrock.

Entirely cold-based polar-continental glacier margins leave very little imprint on the landscape. Features that appear to be terminal and hummocky moraines often constitute a thin veneer of debris overlying buried ice.

The margins of surging glaciers typically have a very high debris content. Surging glaciers also tend to be associated with widespread subglacial and proglacial glacitectonic deformation. Near the margin there is often extensive tectonic thrusting, particularly if the substratum has been weakened by high pore-water pressures. In addition, large discharges of meltwater and sediment associated with glacier surges are responsible for major changes in deposition rates in proglacial lakes and sandar. Sharp (1988) described the geomorphic effects of a glacier surge cycle. During the advance (surge) phase, fluted tills and thrust moraines are formed. At the termination of the surge, crevasse-fill ridges form at the bed, then hummocky moraine and outwash are deposited during glacier recession.

Under circumstances where glacially dammed lakes are breached, exceptionally high magnitude discharges are generated and the proglacial environment will be characterized by extensive erosional landforms as well as outwash deposits. The Channeled Scablands region of Washington State for example derives its name from the dramatic erosional forms generated by the draining of glacial lake Missoula.

Mountain valley systems

The majority of the debris transported by valley glaciers is derived from mass wasting of valley walls. Valley glaciers in high relief settings typically have extensive covers of supraglacial debris. Following ice retreat, high-relief mountain environments are subject to paraglacial reworking of ice-marginal sediments and landforms.

The margins of valley glaciers in mountain areas are commonly delimited by latero-frontal dump and push moraines. Meltwater deposition produces ice-marginal ramps and fans. In low relief mountain areas (e.g. the Scottish Highlands or the Norwegian mountains) Neoglacial end moraines are typically 2–5 m high. In debris-rich high relief settings the latero-frontal moraines can be much higher, so that repeated glacier advances may terminate at the same location and contribute to moraine-building, resulting in large landforms which can exceed 100 m in height.

Glacier retreat in mountain environments is normally recorded by recessional moraines. However, debris-mantled glaciers tend to remain at the limits imposed by their latero-frontal moraines until advance or retreat is triggered by significant climatic change. The landform record of a retreating debris-mantled glacier often consists of major moraine complexes, separated by extensive tracts of hummocky moraine deposited during episodes of ice-margin wastage and stagnation.

Glacial lakes are common features in mountainous environments. In low relief mountains they are mainly ice dammed as a result of the blocking of side or trunk valleys by expanded glacier tongues. Good examples of landform and sediment assemblages formed in Pleistocene lakes are found in the Highlands of Scotland and the water levels of former Glen Roy lake, for instance, are recorded by very prominent shorelines known locally as the 'Parallel Roads'. In high-relief mountain environments proglacial lakes dammed by moraines or by rockfalls, landslides and debris flow fans are more important than ice-dammed lakes. Some of the modern proglacial lakes impounded by latero-frontal moraines in the Andes and the Himalaya present a high risk of outburst floods to downvalley settlements.

Glacifluvial deposits are commonly well preserved in low to moderate-relief glaciated valleys. The focusing of meltwater flow by valley sides results in the erosion of gorges or the deposition of ribbon-like valley trains along valley axes. Staircases of terraces occur along the floors of many glaciated valleys, and the highest members may show signs of ice-marginal kame deposition. In valleys of high mountain environments mass movement features, lacustrine and fluvial sediment accumulations, and river terraces may be much more widespread than glacigenic landforms soon after their deglaciation. Paraglacial reworking of glacial landforms and sediments is less effective where glaciers advanced from high-relief mountainous regions to the foothills, and in such settings the preservation potential of the substantial ice-marginal landforms is greater. The margins of the Pleistocene piedmont glaciers in the northern foothills of the European Alps are delimited by high semicircular terminal moraines surrounding excavational basins. The ice proximal flanks of the moraines are characterized by kame and kettle topography and the central parts of the basins are either occupied by lakes or exhibit well-preserved DRUMLIN fields. The outer flanks of the moraines are skirted by large outwash fans, in which flights of several terraces were incised by postglacial rivers. For this spatial arrangement of landforms the term 'glacial sequence' was coined by Penck and Brückner (1909).

Subaquatic systems

Given the extent of water-terminating glacier margins today and during the past, the subaquatic proglacial environment is an important one. More than 90 per cent of the Antarctic ice sheet margins terminate in the sea. The Pleistocene northern hemisphere ice sheets were bordered in many places by lakes hundreds of kilometres across, ponded between the ice and topographic barriers, or by epicontinental seas. In Europe, the largest proglacial lake was the Baltic ice lake which, around 10,500 years BP, stretched for some 1,200 km along the southern margin of the Scandinavian ice sheet. The most extensive of the North American lakes, inundating 2,000,000 km^2, has been named proglacial Lake Agassiz (Teller 1995).

The type of sediment-landform association deposited adjacent to the glacier grounding line depends on ice velocity and calving rate, sediment supply, input of meltwater from the ice, and water depth and salinity.

Sediment supply and subglacial discharges of meltwater are highest at temperate glaciers with a tidewater front. Large amounts of coarse debris are deposited in morainal banks and grounding-line fans, and fine-grained sediments are carried away in turbid plumes to form a distal zone of laminated mud deposits. At the grounding line of temperate glaciers ending as an ice shelf meltwater-related processes tend to be less important and grounding-line fans are less common. Grounding-line deposits pass into drapes of dropstone muds, released by meltout of sediments embedded in the basal zone of the ice shelf and further out into dropstone muds derived from icebergs. Iceberg-rafted debris is generally more important in the vicinity of ice shelves where icebergs are often trapped close to the ice margin by sea ice, whereas in the forefront of grounded temperate ice margins sea ice does not restrict iceberg drift.

In contrast to wet-based ice bodies, all glaciers and ice sheets, which are frozen to their beds, provide little debris and little meltwater. Grounding lines below ice shelves are associated with grounding-line wedges, composed of mass flow deposits. In the case of ice margins ending with a tidewater front hardly any sediment will be released to the lacustrine/marine environment and proglacial landforms are rare.

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CHRISTINE EMBLETON-HAMANN

PROTALUS RAMPART

A ridge, series of ridges or a ramp of debris formed at the downslope margin of a perennial or semi-permanent snowbed, which is located typically near the base of a steep bedrock slope in a periglacial environment. Observations of active examples indicate that constituent sediments range from diamicton to accumulations of coarse rock fragments. Roundness of rock fragments can vary from subangular to very angular, depending on the source of sediment supply. In planform, ramparts range from curved, to sinuous and complex. Typically, they have thicknesses (measured perpendicular to the slope) of up to 10 m. Examples on relatively steep slopes tend to have short proximal (i.e. adjacent to the snowbed) and long distal slopes.

The term 'pronival rampart' was preferred by Shakesby *et al*. (1995) on the basis that all examples lie at the foot of a snowbed (as 'pronival' indicates) but not all lie at the foot of a TALUS slope (as 'protalus' indicates). Until the early 1980s, there had been few observations of active forms. Suggested origins were based almost entirely on circular reasoning linking logical but hypothetical processes to supposed fossil 'ramparts', which might easily have been mistaken for other landforms (e.g. ROCK GLACIER, LANDSLIDE, MORAINE, avalanche impact ridge). It was reasoned that ramparts were formed entirely by coarse rockfall debris rolling, bouncing or sliding down a snowbed surface with very little if any fine debris reaching the rampart (White 1981). Any fine sediment in the ramparts, it was suggested, had been derived by *in situ* weathering or by the impacts of transported rock fragments. During the mid- to late 1980s, other processes (AVALANCHES and DEBRIS FLOWS) were found to be supplying fine as well as coarse material across snowbed surfaces to actively forming ramparts (Ono and Watanabe 1986; Ballantyne 1987). During the 1990s, the range of processes was expanded to include those operating beneath snowbeds, both as regards sediment supply (snowmelt, debris flow, SOLIFLUCTION) and modification of pre-existing sediment (bulldozing by a moving snowbed) (Shakesby *et al*. 1995, 1999). Because of confusion with other upland depositional landforms, there have been a number of attempts to identify diagnostic criteria, which have included morphological and sedimentological characteristics. In particular, attention has focused on the distinction between ramparts and moraines formed by small GLACIERS. Since, however, many 'rampart' characteristics have been based on (1) conjectural fossil examples, and (2) assumed formation at the bases of static snowbeds (although a rampart origin by a mobile snowbed has been demonstrated (Shakesby *et al*. 1999)), such diagnostic criteria must be viewed with extreme caution.

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SEE ALSO: nivation; periglacial geomorphology

RICHARD A. SHAKESBY

PSEUDOKARST

A term first employed by von Knebel in 1906 (Bates and Jackson 1980), which has been widely used to describe topography, landform assemblages or features developed on non-carbonate rocks which exhibit a morphology similar to those characteristic of carbonate KARST terrain.

Such a lithology based definition excludes landforms in non-carbonate rocks from genuine karst. This classification is still in use by some geomorphologists and speleologists. However a more recent, all-encompassing definition of karst is now becoming increasingly widely accepted (Jennings 1983). Jennings's designation is less restrictive and he argued that karstic processes and landforms may be found on any rock type where the 'process of solution is critical, although not necessarily dominant'. Pseudokarst landforms should therefore be considered as those that morphologically resemble karst, but have formed through processes that are not dominated by solutional weathering or solution-induced subsidence and collapse.

Pseudokarst includes landforms morphologically similar to those commonly associated with carbonate or GYPSUM KARST landscapes, and include subterranean drainage, CAVES, DOLINES, BLIND VALLEYS, grikes, SPELEOTHEMS and surface KARREN.

Examples of pseudokarst fulfilling the conditions of Jennings definition, that is, where solution is not a critical formative process, include (1) caves in glaciers, or topographic depressions in permafrost regions (thermokarst), caused by a change in phase (i.e. from solid to liquid water) rather than dissolution (Otvos 1976); (2) VOLCANIC KARST, comprising tunnels within lava, formed where molten lava continued to flow inside an already partially solidified lava bed (i.e. caves are formed at the same time as the host rock) (Anderson 1930), and also depressions associated with the mechanical collapse of such caves; and (3) caverns and karst-like features caused by predominantly mechanical erosion of rock by animals such as abrasion caused by molluscs in the tidal zone of limestone outcrops on tropical and temperate coasts (Sunamura 1992), or moving water, wind or ice. Some workers also class as pseudokarst depressions and pipes (see PIPE AND PIPING) formed in soils or other unconsolidated sediments by the mechanical erosion of unconsolidated material (piping) (Otvos 1976) as, for example, often found within loess deposits.

In view of its original definition and long-term usage, the term pseudokarst can also be widely found in the literature referring to any karst-like features in rocks other than limestone (or gypsum) (including rocks such as basalt, granite or diorite) regardless of their mode of formation. Examples of these so-called pseudokarst features include basins, runnels, caves, underground drainage and even small speleothems. Provided these features can be ascribed to a range of physical or chemical weathering and erosive processes that do not rely on solution to any significant extent, the use of the term pseudokarst is appropriate.

The term pseudokarst has, however, also often been applied to landforms on rocks of relatively low solubility such as quartzites or highly siliceous sandstones, which consist almost entirely of silica (SiO_2) (e.g. Pouyllau and Seurin 1985). Such usage has been based on the widely held but incorrect assumption that quartzose rocks are practically immune to chemical weathering (Tricart 1972). This belief is based on the fact that the equilibrium solubility of many carbonate rocks ranges between 250 and 350 $mg\,l^{-1}$ at normal temperatures, whilst under the same conditions the solubility of crystalline silica (quartz) does not exceed 15 $mg\,l^{-1}$, and even that of amorphous silica is less than half that of many carbonates. Quartzose rocks were thus generally considered not to develop solutional and therefore 'genuine' karst, but rather pseudokarst (e.g. Pouyllau and Seurin 1985).

However, during the past few decades features of considerable dimensions and striking morphological similarity to dissolutional karst have been identified in quartzose rocks in Africa, South America and Australia. For example, in Africa solutional landforms and caves are found in the quartz sandstones and quartzites of Tchad, Nigeria, South Africa and the Transvaal; the great South American quartzite landscapes of Brazil and the Venezuelan Roraima display numerous large and small, remarkably carbonate-like, surface forms, silica speleothems and many cave systems with lengths exceeding 2.5 km and depths of 350 m; and the quartz sandstones of the Arnhem Land and Kimberley regions of northern Australia and even the Sydney region of south-eastern Australia displays many caves, tower karst, smaller surface karren and speleothems (see Wray 1997 for a detailed review).

A range of studies carried out in these highly quartzose regions has now either argued or directly shown that the prime process leading to these 'pseudokarst' features is the direct dissolution of silica. Where quartz grains are held together by amorphous silica cement, the dissolution of this comparatively soluble material (up to about 150 $mg\,l^{-1}$) may isolate individual quartz grains from the parent rock (arenization) (Jennings 1983), which may then be removed by flowing water. However, arenization also occurs in rocks with very little amorphous cement when individual quartz grains and crystalline overgrowths are dissolved despite their low solubility (see especially Jennings 1983 for northern Australia; Wray 1997 for south-eastern Australia; and Chalcraft and Pye 1984 for South America). In a study investigating cave passages in the well-developed quartzite karst in Venezuela, Doerr (1999) has even argued that, under specific conditions, such karstforms may develop largely through dissolution, with arenization playing only a minor role.

A range of workers conclude that in many areas with quartzose rocks, dissolution is the key process in the formation of karst-like features, and argue that genuine karst may develop in highly siliceous rocks, where very long periods of weathering offset slow rates of dissolution (e.g. Jennings 1983; Chalcraft and Pye 1984; Wray 1997; Doerr 1999). Following the earlier urgings of Jennings (1983), Wray (1997) argued in a wide-ranging and comprehensive analysis of the worldwide karst-like features in quartzites and quartz sandstones, that in these features, the critical role of solution clearly identifies these forms as true karst (i.e. quartzite or sandstone karst) and not pseudokarst.

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SEE ALSO: biokarst; chemical weathering

STEFAN H. DOERR AND ROBERT WRAY

PULL-APART AND PIGGY-BACK BASIN

Pull-apart and piggy-back sedimentary basins are typically associated with convergent plate tectonic settings (see PLATE TECTONICS). Pull-apart basins are topographic lows developed by rifting along strike-slip faults in areas of transtension (i.e. areas subjected to both transform and extension tectonics). The term 'pull-apart' was first used by Burchfiel and Stewart in 1966 to describe features in the Death Valley region of the USA and was later used by Crowell in 1974 to describe features along the San Andreas fault. Pull-apart basins have also been referred to as 'rhombochasm' and 'rhombograben' for the largest features (several kilometres by tens of kilometres in dimensions) and 'sag pond' for the smallest features (a scale of tens to hundreds of metres) (Seyfot 1987). The areas of transtension that develop pull-apart basins are typically associated with either (1) bends in the fault system (known as releasing bends) or (2) fault off-sets. These bends or off-sets need to step over to the left for a left lateral fault system or right for a right lateral fault system in order to generate the required transtension for basin development. The resulting basin is bounded on two sides by the strike-slip faults (which also have a significant normal component to the fault movement) and on the other two sides, approximately perpendicular to the main strike-slip faults, by normal faults. With continued extension of the basin the floor can be stretched and thinned to the extent that volcanism may occur and thus may cover the floor of the basin. Sedimentary fill of the basins may be developed in one main depocentre (area of maximum subsidence) in the central part of the basin or two depocentres, each adjacent to the bounding normal faults (Deng *et al.* 1986). Modelling by these authors suggests that the number and position of the depocentres is dependent on the geometry of the basin which in turn is dependent on three main factors (1) separation between the overlapping lateral fault strands; (2) degree of overlap between the main lateral faultstrands; and (3) the depth to the basement. Basins elongated parallel to the main lateral faults (overlap is more than separation) tend to have two depocentres, whereas 'shorter' basins where the separation is more than the overlap tend to have one depocentre. In most cases the depth of the basins typically tends to be greater than typical rift basins developed in divergent plate tectonic settings, and tends to be dominated by alluvial or lacustrine sedimentary fill.

Piggy-back basins, in contrast, are typically associated with thrusted terrain where basin development is complicated by deformation of earlier basin deposits by more recent thrusting. The term 'piggy-back basin' was first used by Ori and Friend in 1984 to describe minor sedimentary basins that rest on moving thrust sheets. Such basins have also been termed 'thrust-sheet-top basins' (Ori and Friend 1984)', 'satellite basins' (Ricci-Lucchi 1986) and 'detached basins' (Steidmann and Schmitt 1988). These basins are typically a few tens of kilometres across and are physically separated from the foredeep (the basin in front of all the active thrusts). Classic examples of piggy-back basins are found throughout the Alpine mountain chains of Europe. The basin fill comprises sediment sources from all basin margins, with a dominant provenance from the uplifted ramp of the older thrust behind the basins. Sedimentary environments range from coarse submarine fan and fan delta to fluvial deposits. Fluvial systems typically comprise a transverse drainage from the thrust ramps on both sides of the basin and a longitudinal drainage which enters the basin from the topographic lows that develop above lateral fault terminations (Miall 1999).

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SEE ALSO: fault and fault scarp; plate tectonics; rift valley and rifting; tectonic geomorphology

ANNE E. MATHER

PUNCTUATED AGGRADATION

The theory that the long-term aggradation of sediment (through geological time) has been via episodic SEDIMENTATION. This is in contrast with the traditional concept of UNIFORMITARIANISM and the continual and gradual build-up of sediments through time. Early studies such as that by Barrell (1917) provided the initial challenge to the long-held paradigm of gradual aggradation. The theory of punctuated aggradation began to gather momentum once more in the early 1980s. Ager (1980: 43) fuelled the debate by referring to sediment stratigraphy as having 'more gaps than record', and argued that the large disparities between modern sediment deposition (for a specific environment) and ancient calculated deposition was a result of the episodic nature of aggradation. The theory of punctuated aggradation treats each bedding plane as a pause in sedimentation, whereas continual aggradation considers bedding planes as merely signifying a change in diagenesis or texture, and treats the formation as the basic stratigraphic unit, each one a product of a particular environment.

The term punctuated aggradational cycle (or PAC) was coined by Goodwin and Anderson (1985), within their hypothesis for episodic stratigraphic accumulation. The hypothesis argues that, allowing minor exceptions, the stratigraphic record consists of thin (1–5 m thick), basin-wide, shallowing-upward cycles. These are sharply defined by surfaces produced by geologically instantaneous relative BASE-LEVEL rises (termed punctuation events). Deposition occurs during intervening periods of base-level stability. A host of depositional environments can be included in the PAC hypothesis (e.g. fluvial, deltaic, shelf, slope, etc.), as PACs are assumed to exist in all depositional environments influenced by rapid base-level rises.

The PAC hypothesis proposes that allogenic processes such as sea-level change are responsible for changes in the stratigraphic record, rather than autogenic processes (e.g. channel migration, etc.) that are held as responsible in continuous aggradation. Autogenic processes are not dismissed entirely, but are treated as localized stratigraphic influences, superimposed on the allogenic processes. The bounding surfaces between the PACs are often traceable laterally for vast distances since they are formed by large-scale allogenic processes. This allows them to be accurate stratigraphic markers in the field. Base-level rise during a punctuation event can be rapid (reaching 1 m per 100 years) whereas stratigraphical analysis indicates that the recurrence of such punctuation events can be as frequent as 50,000 years, thus reflecting the rapidity of the base-level rise. Thickness of PACs, though generally thin, varies considerably though long-term aggradation rates remain similar. Goodwin and Anderson suggest that the most likely mechanisms responsible for PACs would include episodic crustal movement, episodic movement of the geoidal surface and global eustatic sea-level changes.

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STEVE WARD

PYROCLASTIC FLOW DEPOSIT

Pyroclastic flow deposits are the products of fragmental material transported laterally by gas-charged, concentrated flows (sometimes called NUÉES ARDENTES). Pyroclastic flows are generated in many ways, with a spectrum from the 'passive' collapse of oversteepened lava-flow or dome margins, through the gravitational collapse of high eruption columns, to powerful overpressured blast-like events. In contrast to gravity-controlled lava flows (see LAVA LANDFORM) and water-charged LAHARS, pyroclastic flows may possess considerable momentum and cross substantial obstacles (sometimes >1-km high mountains). Pyroclastic flow deposits are so diverse that they are here described in terms of five spectra.

The first spectrum is in densities of the juvenile (newly erupted) component, which reflect the relative importance of expansion of dissolved volatiles in frothing and fragmenting the magma. Densities are higher (1.0–2.7 $\mathrm{Mg\,m^{-3}}$) in many small-volume deposits, particularly those associated with composite VOLCANOes and/or the collapse of lava domes, where the magma is fragmented by external means such as crushing or shattering by interaction with water. In larger deposits, clast densities reduce to $<1\,\mathrm{Mg\,m^{-3}}$ (i.e. pumice), commensurate with an increasing role for expansion of dissolved volatiles. Simultaneously, contents of fine ash (<1/16 mm) increase; deposits with dense juvenile clasts have low contents (typically <2–5 per cent), those containing pumice have higher contents (>10–15 wt per cent). Dense clast-rich deposits often contain abundant large (dm–m-sized) juvenile clasts, and are labelled as, e.g. 'block-and-ash flow deposits' or 'dome-collapse avalanche deposits'. Deposits where the juvenile component is pumice are termed 'ignimbrites' or 'ash-flow tuffs'; collectively they represent by far the greatest volume of pyroclastic flow deposits worldwide.

The second spectrum is size. Distances travelled range from a few hundred metres in lava-collapse flows, to >150 km for prehistoric large ignimbrites. Areas range from a few thousand square metres to >30,000 $\mathrm{km^2}$. Volumes range from about 1,000 $\mathrm{m^3}$ for individual dome-collapse events to >1,000 $\mathrm{km^3}$ for large ignimbrites. Observed pyroclastic flow eruptions generated only relatively small examples, with distances travelled up to 30–40 km, areas up to 400 $\mathrm{km^2}$, and volumes up to ~15 $\mathrm{km^3}$. Small pyroclastic flow deposits (<1 $\mathrm{km^3}$) are generated from vents on composite volcanoes or from collapse of lava flows/domes. Intermediate-sized deposits (up to a few tens of $\mathrm{km^3}$) can be generated from composite volcanoes or CALDERA volcanoes, often associated with caldera collapse. Larger deposits are associated with eruptions of gas-rich, evolved magmas (particularly rhyolite) from caldera volcanoes.

The third spectrum is deposit morphology. Individual, small-volume pyroclastic flows form tongue-like deposits, often with surface ridging, marginal levees and lobate flow fronts akin to those developed on DEBRIS FLOWS. However, most deposits form during many (tens to hundreds of) individual flow events, and so the gross deposit morphology then reflects the energetics of flow emplacement and deposit volume. The energetics are represented by the 'aspect ratio', which is the ratio of the average deposit thickness to the diameter of a circle with the same area as the deposit. Sluggishly emplaced deposits have a high aspect ratio (as high as 1 : 200), that is, the material is relatively thick for its extent. Energetically emplaced deposits have low aspect ratios (> 1 : 10,000), that is, the material is very widespread for a given volume of material. The volume of the pyroclastic deposit coupled with its aspect ratio then yields three major morphologies: landscape-mantling, landscape-modifying, and landscape-forming (Figure 129). The largest deposits can create wholly new land surfaces over areas of > 1,000 km^2, forming fan- or pediment-like surfaces around the source volcano.

The fourth spectrum is in the internal structure of the deposits. Single pyroclastic flows generate single flow units, that may be composed of a number of layers and facies that in turn reflect the mechanics of flow emplacement. Deposits of multiple flows should, in principle, show multiple flow units, but the clarity with which flow-unit boundaries can be discerned within such deposits is very variable. Thick stacks of ignimbrite may show no stratification, or only vague bedding or fluctuations in grain size, to suggest that they are the product of multiple flows. Grading structures within individual flow units vary widely also, and can reflect both migration of coarse clasts (regardless of density) under shearing forces, and

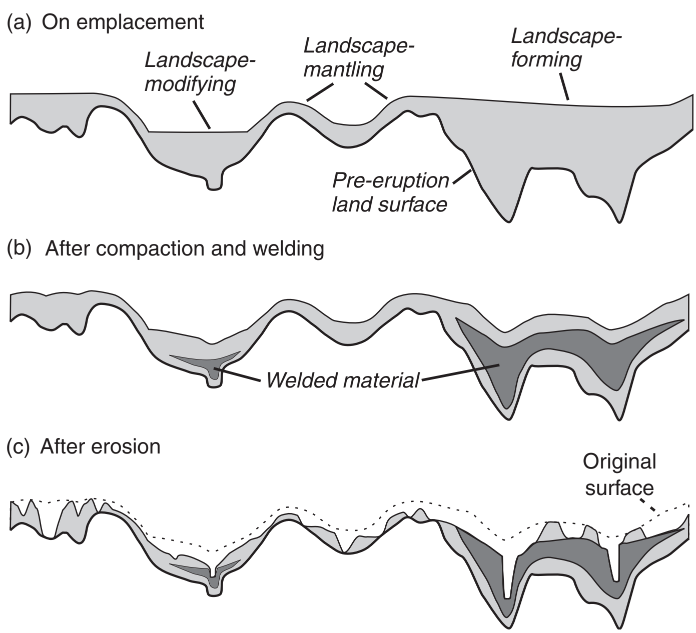

Figure 129 Schematic diagram to illustrate the morphologies of pyroclastic flow deposits. (a) On immediate emplacement, showing how the pre-eruptive landscape may be mantled, modified, or buried, depending on the thickness of the deposit, the topographic relief and the energetics of emplacement. (b) After consolidation and welding; note how compaction is greatest where the deposits are thickest, thus new valleys are generated along the line of pre-eruptive buried valleys. (c) After erosion; valleys are re-cut along their old courses. Non-welded deposits are preferentially removed, but summit heights may still be concordant, reflecting the original surface

flotation/sinking of lighter/denser coarse clasts, respectively, under buoyancy forces.

The fifth spectrum is in the lithologies of the deposits. Pyroclastic flows are efficient conservators of heat, and so many deposits are emplaced at temperatures above those at which the juvenile material can flow plastically (e.g. >550–$600\,^{\circ}C$ for rhyolitic pumice). The combination of retained heat and load stresses imposed by overlying deposits causes the juvenile fragments to adhere and flatten (weld) to form a coherent rock. At its most extreme, welding can eliminate all initial pore space and the rock may be so hot as to continue to flow plastically as a kind of lava flow. Welding can only occur as long as the juvenile phase is glassy, but in most welded deposits the glass has subsequently devitrified. In addition, gases released from the juvenile material can cause further crystallization and vapour-phase alteration of the deposit, either along discrete pathways ('fossil fumaroles') or pervasively through the porous rock mass. Non-welded deposits show little or no JOINTING, but welding (and any other causes of induration) is generally accompanied by formation of jointing in the rock mass. The orientation and spacing of the joints can vary, but columnar joints, spaced at decimetres to metres apart, are characteristic of the interior of thick ignimbrites. Closer to the base, top or sides of the deposits, or in places where local fluxes of hot gases have occurred, the jointing can be more closely spaced and fan-like in disposition.

The morphologies of freshly emplaced pyroclastic flow deposits (Figure 129) are generally very rapidly modified by erosion, as loose pyroclastic-flow material is readily eroded, generating syn- and post-eruptive debris flows, lahars and HYPERCONCENTRATED FLOWS. Incision by streams often occurs so rapidly that interaction may occur between water and the still-hot interior of the deposits, leading to 'rootless' phreatic explosions. In non-welded deposits, incision rates of metres to tens of metres per rain event are known. Incision tends to recur along the lines of the pre-eruption valleys; the greatest thicknesses of deposits (and hence the greatest compaction) occur there and so the pre-eruptive topography is mirrored in subdued fashion on the surface of the deposits, controlling the paths of re-established streams. Erosion slows considerably when hard (welded) material is reached, or the non-welded deposits are stabilized by regrowth of vegetation.

Landscape morphologies seen in areas covered by pyroclastic flow deposits reflect a complex interplay between the initial depositional morphology, the presence or absence of welding or induration to create hard rock, and the local climate. A characteristic feature in dissected large ignimbrites is a concordance of ridge or summit heights, defining a surface parallel to the original deposit surface. Slopes in non-welded deposits are typically at or close to the angle of rest, except along streams or river where undercutting leads to vertical cliffs. Slopes in welded deposits are often cliffed, as the removal of material is controlled by vertical jointing that allows toppling of columnar masses as they are undermined by erosion.

Although pyroclastic flow deposits are volumetrically important in many volcanic terrains, the enormous variety of characteristics these deposits can display, and the hazards associated with flow emplacement, mean that there is still much to be discovered about the processes and products of pyroclastic flows.

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Q

QUICK FLOW

Hydrologists generally separate streamflow into two operationally defined components: event flow, considered to be the direct response to a given water-input event (also called direct runoff, storm runoff or stormflow), and base flow, which is water that enters from persistent, slowly varying sources and maintains streamflow between water-input events (derived largely from groundwater circulation). Quick flow is simply another term for event flow. The mechanisms involved may be one, or a combination, of Hortonian overland flow, saturation overland flow, and near-stream subsurface storm flow via groundwater mounding. In the latter case, at least some of the water identified as quick flow is 'old water' that entered the basin in a previous event. Quick flow can also be 'delayed', which involves storm runoff from distal sources via predominantly subsurface routes.

SEE ALSO: runoff generation

MICHAEL SLATTERY

QUICKCLAY

The quickclays (quick clays, quick-clays, Swedish: *kvicklera*) are clay-sized postglacial marine sediments of very high sensitivity (see SENSITIVE CLAY). The term relates to the old Nordic *qveck*, meaning living. They are found in Norway, Sweden and Canada, and to a much lesser extent in Alaska, Finland and Russia, and they have been defined as having a sensitivity of greater than 50. The original definition was: a clay whose consistency changed by remoulding from a solid to a viscous fluid. Very high sensitivity values have been found – up to 200 for the Champlain clays of east Canada. The literature is dispersed; there are reviews by Bentley and Smalley (1984), Cabrera and Smalley (1973), Maerz and Smalley (1985), McKay (1979, 1982), Brand and Brenner (1981) and Locat (1995). The high sensitivity value means that the clays lose most of their strength on remoulding, and this can lead to catastrophic landslides, which progress rapidly as flowslides. Soderblom (1974) proposed that two types of quickclays should be recognized: rapid quickclays and slow quickclays. The rapid materials lose their strength very quickly on reworking; but the slow materials require the input of a fairly large amount of energy before they convert to a liquid. The strength parameters of the remoulded clays can be difficult to measure.

The classic quickclay explanation by I.Th. Rosenqvist (1953) depended on postglacial uplift, and leaching. The clay material was deposited in shallow salty seas in immediate postglacial times. As postglacial uplift occurred these deposits became dry land and were exposed to rainfall and groundwater flow. This had the effect of leaching out the salts and changing the electrochemical environment of the soil particles. The loss of the soil cations meant that the system became more metastable and responded to stress via soil structure collapse, LIQUEFACTION and flowsliding. The Rosenqvist theory appeared to work for the rapid Scandinavian clays, but not to be so suitable for the slower Canadian clays.

As mineralogical analysis became more sophisticated it became apparent that in many quickclays the actual clay mineral content was quite low and that they were perhaps better described as very fine silts. This fitted in rather well with their observed distribution on the fringes of glaciated regions. Glacial action could provide the very fine primary mineral material required to form the quickclay deposits. In fact the geomorphological observations led to a new approach to quickclays which has become known as the inactive-particle, short-range bond theory. This requires that the quickclay systems be cohesive (by virtue of the small particle size) but not plastic (because of the predominance of primary mineral particles, e.g. quartz, and the shortage of clay mineral particles). The fine blade-shaped primary mineral particles sediment in the shallow sea as Rosenqvist required, and form an open rigid structure; but the interparticle bonding is not the long-range clay mineral-type bonding but rather a short-range contact bond, enhanced by cementation.

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SEE ALSO: liquefaction; sensitive clay

IAN SMALLEY

QUICKSAND

Quicksand requires a flow of water. As the water flows through sands and silts and loses pressure its energy is transferred to the particles that it is flowing past, which in turn creates a drag effect on the particles. If the drag effect is in the same direction as the force of gravity, then the effective pressure is increased and the system is stable. In fact the soil/sediment tends to become denser. Conversely, if the water flows towards the surface, then the drag effect works against gravity, and reduces the effective pressure between the particles. If the velocity of the upward flow is sufficient it can buoy up the particles so that the effective pressure is reduced to zero. This represents a critical condition where the weight of the submerged soils is balanced by the upward-acting seepage force. This critical condition sometimes occurs in sands and silts. If the upward velocity of flow increases beyond the critical hydraulic gradient a quick condition develops.

Quicksands, if subjected to deformation or disturbance, can undergo a spontaneous loss of strength, which causes them to flow like viscous liquids. Karl Terzaghi, in 1925, explained the quicksand phenomenon as follows: first, the sand or silt concerned must be saturated and loosely packed. Second, on disturbance the constituent grains become more closely packed, which leads to an increase in pore-water pressure, reducing the forces acting between the grains. This brings about a reduction in strength. If the pore water can escape very rapidly the loss in strength is momentary. The third condition is that the pore water cannot escape readily. This occurs if the sand or silt has a low permeability or the seepage path is long, or both.

Casagrande, in 1936, demonstrated that a critical packing porosity existed above which a quick condition could be developed. He proposed that many coarse-grained sands, even when loosely packed, have porosities just about equal to the critical condition, while medium- and fine-grained sands, especially if uniformly graded (a narrow range of particle size), exist well above the critical porosity when loosely packed. Thus fine sands (say 60–150 μm) tend to be potentially more unstable than coarse-grained sands. The finer sands tend to have lower permeabilities.

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SEE ALSO: liquefaction; quickclay

IAN SMALLEY

R

RAINDROP IMPACT, SPLASH AND WASH

One of the most important driving forces in soil and hillslope EROSION is the kinetic energy of raindrops striking the soil surface. Raindrop impact contributes to soil erosion directly by splashing particles downslope, by entraining particles in OVERLAND FLOW which is below the threshold conditions necessary to pick up material. It can also affect erosion indirectly by disrupting soil aggregates, increasing ERODIBILITY, and by beating the surface into an almost impermeable seal or crust (see CRUSTING OF SOIL), which reduces infiltration and increases runoff (see RUNOFF GENERATION) discharge during rainstorms.

The kinetic energy of a moving object is expressed by $0.5\,MV^2$ where M = mass of the object and V = velocity. In the case of raindrops, the velocity is the terminal velocity which, in still air, reaches values around $9\,m\,s^{-1}$, for drops of 5 mm diameter (Laws 1941). During rainstorms, this value can be significantly affected by near-ground turbulence and wind. Raindrop mass is an even more critical control on the kinetic energy of raindrop impact. Raindrop size varies greatly from minute droplets a few microns in diameter, to an upper limit around 6.5 mm. As raindrop mass is directly proportional to diameter, there is a huge difference in the kinetic energy expended by small and large drops as they strike the surface. Comprehensive understanding of the relationship between raindrop impact and rainstorm characteristics is limited by the scarcity of accurate drop size measurements, particularly during rainstorms of very high intensity. However, Hudson (1981), amongst many others, has shown that rainstorms typically have a normally distributed spectrum of drop sizes, which can be expressed by the median drop diameter. This ranges from around 1.8 mm for a rainstorm of $12.7\,mm\,h^{-1}$ intensity to about 2.3 mm for a rainstorm of $65–115\,mm\,h^{-1}$ intensity. Information about characteristics of very high intensity rainfall is limited, because the most intense storms are usually of very limited duration and extent. It was thought that intensities above $150\,mm\,h^{-1}$ are very rare and largely limited to the tropics, but recent observations suggest that intensities as high as $400\,mm\,h^{-1}$ are by no means uncommon, particularly for very short periods, particularly at the beginning of thunderstorms.

Although information about raindrop size and rainfall intensities is still deficient, it is clear that there are major systematic differences between different types of rainfall and different parts of the world, which are reflected in the kinetic energy expended and the capacity of raindrop impact to generate erosion. The highest energy expenditure is certainly associated with the large drops and high intensities of severe thunderstorms or orographic rainfall and so the highest annual rainfall erosivities (see EROSIVITY) occur in areas like Assam or Hawaii, where such rainfall is combined with high annual totals. By comparison, the predominantly frontal rainfall of temperate areas produces very low kinetic energy, though occasional severe storms can, of course, cause much damage. The effect of raindrop impact is, however, strongly affected by vegetation. A dense vegetation cover can absorb virtually all the kinetic energy of raindrops, almost eliminating erosional hazard. However, although it takes about 30 m fall for drops to achieve full terminal velocity, they can achieve 60–70 per cent with a fall of some 3 m. Unless there is dense vegetation near or on the surface, raindrops can therefore regain much of their kinetic energy before hitting the surface. As a result, trees are not usually effective in controlling soil erosion in the absence of ground cover.

Raindrop impact affects the soil surface in several different ways. It may cause crusting by compacting the surface, increasing soil density and shear strength. It may also disrupt unstable soil aggregates, producing small fragments which can wash into pores and cracks, effectively sealing the surface. The resulting thin seal (often <1 mm in thickness) can make the soil surface almost entirely impermeable. The effectiveness of raindrop impact in causing compaction, disruption, crusting and sealing depends on rainfall characteristics, soil properties and on soil moisture content. Aggregate disruption by SLAKING is most effective on dry soils, while compaction is most effective on wet clay soils where cohesion drops close to zero. Although bursts of extremely high intensity rainfall, which cause most disruption, are usually very short-lived, they can strongly influence the subsequent effectiveness of erosional processes. This is particularly true in the case of intense summer thunderstorms where initial very high intensity rainfall often falls on a dry surface. These bursts usually last only a few minutes, but by initiating sealing, can result in almost instantaneous overland flow.

Raindrop impact may be entirely absorbed by soil and vegetation, but in intense storms there is usually sufficient energy available to generate some erosional processes as well. The exact processes depend on the balance between the amount of water (rainfall) arriving at the surface and the soil infiltration capacity. This will determine whether all the water can infiltrate or whether excess will be available to generate surface ponding and overland flow. Where no excess occurs, wash erosion processes are absent, but splash erosion can occur. On dry soils raindrop impact can produce miniature surface craters, but usually does not move soil particles. As the water content increases, however, soil strength drops rapidly and the surface can become fluidized. Raindrop impact is converted to an upward force which can entrain soil particles and transport them in a parabola away from the point of impact. The distance of movement depends on the mass of the particle, but is rarely more than 0.6 m above the surface, or more than 2 m in a horizontal direction, unless splash is carried by a strong wind. On a horizontal surface (in the absence of wind), movement is not significant, because the ultimate effect of many raindrops striking the surface is abundant movement, but no net transport in any direction. When the surface slopes, however, this changes as up to 60 per cent of entrained material is deposited downslope from the original impact point, so significant net transport can occur.

The relative vulnerability of soil particles and aggregates to entrainment by splash is an important component of soil ERODIBILITY. Poesen and Savat (1981) have shown in laboratory experiments that the relationship between particle size and the threshold impact energy necessary to cause entrainment is quite similar to the Hjulstrom Curve for flowing water. Entrainment of particles with diameters around 0.125 mm typically requires the lowest impact energy. Splash erosion on most slopes during most storms is therefore a selective process, which ultimately transforms the surface material, producing an *erosional lag deposit* which progressively protects the underlying soil from entrainment.

Pure splash erosion (in which material is both entrained and transported by splash) is comparatively rare, but De Ploey and Savat (1968), who originally identified the influence of slope gradient on the balance of upslope and downslope deposition of splashed material, also described the evolution of sandy hillslopes near Kinshasa, Congo, which is almost exclusively controlled by splash. Elsewhere the effects of splash erosion are subtle and often indistinguishable, but where parts of very erodible surfaces are protected by stones or bits of vegetation, the effect of splash is easily seen by the occurrence of miniature Earth pillars or hoodoos.

Splash erosion can occur without any surface water layer, but in the intense rainfall conditions which produce most splash, such a water layer usually forms quite swiftly. Initially this concentrates in micro-depressions, but ultimately it increases sufficiently in depth to overtop roughness elements and generate overland flow. Before reaching this point, however, it starts to influence the splash process. Initially, except on sandy soils, the water layer actually increases splash transport, up to a critical depth which, laboratory experiments suggest, ranges from about the diameter of the raindrops (Palmer 1963) to about one-fifth of that value (Torri *et al.* 1987). As drop size varies greatly in any rainstorm, the precise result is a very complex mixture of processes on the surface. Eventually, however, the increasingly deep water layer protects parts of the surface from splash erosion. As the first areas protected are microtopographic depressions, the overall effect of continued splash erosion is diffusion of soil particles from higher points to these depressions, progressively reducing the amplitude of the microtopography. Another important effect is the increasing heterogeneity of soil infiltration characteristics, as the *structural* crusts which form on the high points typically have infiltration capacities up to six times higher than the *depositional* crusts which form in depressions (Boiffin and Monnier 1985).

The interaction of spatially varied rainfall, splash and microtopography produces complex, heterogeneous conditions on most hillslopes, particularly with regard to transition from splash-dominated areas to those dominated by overland flow and wash processes. On simple, idealized, homogeneous hillslopes, it is possible to distinguish an upper splash-dominated zone from a lower wash-dominated zone, and finally, from a zone in which concentrated RILL erosion occurs. In practice, the boundaries between these zones are highly irregular and dynamic. However, a transition does occur downslope as surface water deepens progressively, ultimately protecting the surface from raindrop impact. The first stages of overland flow are, however, typically very shallow. Conceptually, on very smooth surface there may actually be a thin, continuous sheet of water, but in practice as most surfaces are quite irregular, this is very rare. The initial flow usually consists of irregular, tortuous concentrations in depressions, which vary significantly in depth and width, and are separated by microtopographic protuberances. Numerous field and laboratory studies have shown that flows of this sort are typically laminar or transitional, with Reynolds numbers often well below 2,500, and relatively smooth, with Froude numbers well below 1. The flows, whether as a sheet or as more or less concentrated streams, are slow and pulsatory, and typically do not exert sufficient shear stress to entrain soil particles. However, as the Hjulstrom curve shows, flow velocities necessary to transport fine silts and clays are significantly lower than those required for entrainment. In these circumstances, raindrop impact and splash are still important, as they may be able to entrain material which can then be transported by flow. Such flows are usually referred to as *rain-impacted flows*, and the erosional process as *rainflow* or *rainwash* erosion (De Ploey 1971). The particle transport distance and the effectiveness of rainflow erosion are governed largely by particle density and settling velocity (Kinnell 2001). Significant transport is typically limited to shallow flows no more than about 1.5 times the average raindrop diameter (Kinnell 1991). Because surfaces are irregular, and flow often discontinuous, the transport distance is frequently very short, resulting in small patches of sediment deposition on the hillslope. Nevertheless, in many areas, rainflow is the most effective and frequent erosional process on upper slopes and interrill areas and can ultimately result in highly significant movement of soil to the base of the slope. This is particularly true where loose soil aggregates are of low density or are water-repellent. In some cases, the patches of sediment deposited on the slope by intermittent flows progressively join to form quite extensive *sedimentary* or *depositional seals*. These are usually highly impermeable, and become preferred locations for runoff generation and wash erosion during subsequent rainstorms (Bryan *et al*. 1978).

Once overland flow is sufficiently deep to protect the surface from raindrop impact, rainflow erosion gives way to *wash erosion*. Surface irregularities ensure that most hillslopes will have patches of wash erosion intermixed with splash and rainflow. Once the surface is fully protected, the only force which can cause entrainment is the bed shear stress exerted by flow. Transport will then occur only if shear stress exceeds the threshold necessary to move the most erodible particle. This critical value depends on soil properties, but Moore and Burch (1986) found that it was equivalent to a unit stream power of $0.002\,\mathrm{m\,s^{-1}}$ for many soils. Once unit stream power exceeds values of $0.01\,\mathrm{m\,s^{-1}}$, transport increases rapidly, and wash erosion tends to be replaced by concentrated rill erosion.

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RORKE BRYAN

RAINFALL SIMULATION

The purpose of a rainfall simulator is to deliver rainfall to the soil surface in a controlled manner with realistic simulation of rainfall intensity and drop-size distribution. Rainfall simulators have been used widely over the past few decades, both in the field and the laboratory. Various factors influence the method of rainfall generation including the purpose of the experiment, the soil surface area to be studied, the drop-size distribution of the simulated rainfall, the need to reproduce realistic terminal velocities, and the need for precise replication of rainfall characteristics between experiments.

Broadly, rainfall simulators fall into three categories (Foster *et al*. 2000): sprays, rotating sprays and drip-screens. Because they eject raindrops relatively high above the ground surface, spray systems are capable of achieving rainfall delivery at terminal velocities approaching that of natural rainfall. However, rainfall intensities can be hard to control, because of variation in pumping rates, and rainfall intensity usually decreases with distance from the rotating nozzle. To overcome this latter problem, multiple rotating nozzles are employed, with the overlap distance between the nozzles being determined by the area over which the simulation is to be performed (Foster *et al*. 2000). Drip systems, using hypodermic needles or drop formers, are usually used over small surface areas (typically $< 1\,m^2$). They are less likely to achieve realistic terminal velocities, because of the difficulty of raising the drip screen high enough, but give much better control of rainfall intensity. Intensities as low as $3\,mm\,hr^{-1}$ can be maintained, and replication between experimental runs is good (Bowyer-Bower and Burt 1989).

Despite the widespread use of rainfall simulators in geomorphological research, until recently there has been little co-ordinated effort to collate all the available information regarding the design and purpose of such simulators, or to discuss future developments relating to the use of this technique. To this end, the British Geomorphological Research Group established a Rainfall Simulation Working Group in 1995 to address these issues. The work resulted in a special issue of *Earth Surface Processes and Landforms* (Volume 25, Number 7, 2000) and creation of a website: http://www.geog.le.ac.uk/bgrg/index.html which includes a database of simulators and a lengthy reference list. Lascelles *et al*. (2000) make the point that when rainfall simulation is used for explicitly spatial studies, some prior analysis of the simulator's inherent variability is vital.

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TIM BURT

RAISED BEACH

A raised beach is a relict depositional landform comprising mostly wave-transported sedimentary material and preserved above and landward of the active shoreline. First described by Jamieson (1908), raised beaches can form along marine coasts or lake shorelines and are well recognized as indicators of a fall in relative sea (or lake) level. In certain situations, multiple raised beaches may form adjacent to one another, producing a BEACH RIDGE plain, or strandplain (Otvos 2000). Raised beaches are distinguished here from raised marine terraces on the basis that the former are solely the product of physical depositional mechanisms, whereas the latter have a broader genesis that may incorporate depositional, erosional and/or biogenic (i.e. reefal) processes.

The elevated position of a raised beach relative to active shoreline processes may be the product of one or more of the following mechanisms: (1) tectonic uplift associated with plate-margin convergence (e.g. New Zealand east coast; Garrick 1979); (2) isostatic rebound related to ice-unloading of a land mass (e.g. mainland Scotland; Smith *et al.* 2000); (2) depositional regression involving delivery of sediment to a shoreline at a rate sufficient to allow formation and stranding of successive beaches (e.g. east coast of Australia; Thom 1984), and; (3) forced regression whereby eustatic sea-level fall leads to abandonment of a shoreline (e.g. southern Australia coast; Murray-Wallace and Belperio 1991). In the case of depositional and forced regression, the beach remains at its original elevation, as is the case for many shoreline deposits formed during the Last Interglacial sea level highstand *c.*125 ka BP. Thus the word 'raised' is applied to all stranded fossil beaches regardless of whether the associated landmass has undergone uplift or remained stable.

Clear identification of a raised beach deposit requires satisfying a range of criteria related to the morphology and sedimentology of that deposit. Doing so allows separation from similar coastal depositional landforms such as cheniers (see CHENIER RIDGE) and linear dune ridges. For ice-free coasts, Tanner (1995) identifies four depositional processes that lead to beach ridge formation: wave-swash action, settling lag, storm surge and aeolian action. Along coasts that experience annual freeze-over of the sea (or lake) surface, ice-push is an additional mechanism for beach ridge formation. Each of these five physical mechanisms produces a shoreline deposit with different morphology and sedimentology, as described below.

The most common form of raised beach is produced by wave-swash processes on sandy to gravelly shores. Onshore transport and sorting of sediment across a beach face produces a berm that accretes to maximum wave run-up under spring tidal conditions. Subtle variations in berm morphology exist, ranging from a linear, convex-up ridge with low-angle cross-bedding to a gently landward-sloping uniform surface with continuous subhorizontal bedding. Given alongshore variations in wave energy, both forms may be present along different parts of a shoreline at one time. Consequently, it is possible to find equally variable morphology and internal structure within a raised beach.

Formation of a beach ridge by settling-lag processes is comparatively rare, developing under fetch-limited shallow water conditions such as a small lagoon or pond. Deposition occurs by sand settling out of the water column to produce a low subaqueous flat-topped ridge or bar with discontinuous horizontal bedding. Because wave action is minimal, sediments are not as well sorted as on a swash-formed beach ridge and cross-bedding is characteristically absent. Preservation of a settling-lag ridge as a raised beach typically requires a relatively rapid and permanent lowering of relative sea (or lake) level.

Storm surge is known to result in deposits at elevations above mean high water spring tidal level, either at the beach–dune interface or as a strandline feature on supratidal flats to landward of the fair-weather beach. Grain size is more varied than for fair-weather swash deposits, incorporating the largest materials available. Sedimentary structures reflect higher wave energy, ranging from complex trough cross-bedded sands to imbricated cobbles. The distinction is drawn here between these truly raised storm deposits and an overwash (see OVERWASHING) fan that is also a product of storm surge and typically located on the lagoon side of a low-lying coastal barrier, but is not raised above the elevation range of active sedimentary processes. The role of storms as an agent in the formation of raised beaches is debated in the literature (Tanner 1995), with some authors arguing for storms as an agent of net beach erosion rather than deposition. Documented instances of storm ridge formation (e.g. hurricane ridges, Florida; Tanner

1995), record these as ephemeral features, lasting only until the next storm. Good examples of multiple raised storm beaches exist along the Ross Sea coast of Antarctica where glacio-isostasy during the Holocene has driven coastal uplift (e.g. Hall and Denton 1999).

Aeolian action may contribute to the formation of a raised beach to the extent that wind-blown sand is placed directly on top of a swash-built or settling-lag initiated ridge. A raised beach with aeolian decoration is characterized by an irregular hummocky morphology with low to high-angle cross-bedding that is multidirectional and discontinuous. If vegetated, the internal structure may be weakly bedded to massive in the root zone. Relict dune ridges that are oriented parallel to a shoreline but are solely the product of aeolian processes are excluded from the range of raised beach forms.

Ice-push may also lead to formation of a beach ridge along shorelines that undergo annual freezing of the sea (or lake) surface. An ice-push ridge is typically a discontinuous accumulation of poorly sorted sand- to boulder-sized sediment that forms along the margins of winter sea-ice sheets. Ridge height is a function of the available sediment size, with boulder ridges attaining elevations of ~5 m. Due to the lack of grain sorting, the internal structure of ice-push ridges is characteristically massive. Summer wave-reworking may produce some subsequent sorting of sediment and generation of low-angle cross-bedding of the sand fraction. However, these features are mostly ephemeral, being reworked by the next ice-push.

A raised beach can be used as a proxy for palaeo-sea (or lake) level, providing the range of diagnostic physical sedimentary structures and texture noted above are preserved in the deposits. In particular, a distinction between wave-formed and aeolian sedimentary units is necessary. Thus, a vertical transition from subhorizontal or low-angle cross-bedding in medium to coarse-grained sand to high-angle cross bedding in fine to medium sand, or massive rooted structure would allow this distinction between beach berm and foredune to be drawn. Where multiple raised beaches are preserved on a strandplain, mapping of the beach-foredune contact along a dip-oriented profile can provide for reconstruction of sea (or lake) level change. Examples of this application of the raised beach sedimentary record range from decadal scale fluctuations in shoreline position along Lake Michigan (Thompson and Baedke 1995) to inferred sea level fall along the New Zealand north-east coast toward the close of the Last Interglacial period (Nichol 2002).

Where material suitable for reliable age-dating is incorporated into a raised beach deposit, it is possible to construct a chronology of formation. This is particularly useful for calculating rates of isostatic uplift (e.g. Smith *et al*. 2000), or for estimating rates of shoreline progradation in relation to local sea level and sediment supply (e.g. Tanner 1993). Traditionally, chronological analysis of raised beaches has applied radiocarbon dating to the remains of marine organisms such as shallow water molluscs (Taylor and Stone 1996). Difficulties arise with this method, however, if the material used for dating is not *in situ*. Most of the organic material incorporated in a raised beach is typically reworked from offshore environments and may therefore be considerably older than the enclosing beach sediment. Alternative dating techniques, such as optical dating of beach and dune sands, offer a more reliable avenue for establishing a detailed and accurate chronology of raised beaches, thereby enhancing their utility as a landform that can be used as an indicator of regional geomorphological and geological processes.

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SEE ALSO: beach ridge; chenier ridge; sea level; strandflat

SCOTT NICHOL

RAMP, COASTAL

The term 'ramp' has been used by some workers to refer to gently sloping SHORE PLATFORMS, particularly to those in the north Atlantic, in order to distinguish them from the subhorizontal platforms which are more common in Australasia. Generally, however, the term is either used for sections of higher gradient at the rear of gently sloping shore platforms, or for steeply sloping rock surfaces (commonly 4° to 10°) that occupy the entire intertidal zone and may extend to elevations that are well above the high tidal level. Both types of ramp have been reported most frequently from the swell wave environments of Australasia and elsewhere around the Pacific, and less frequently from the storm wave environments of the mid-latitudes of the northern hemisphere. It has been suggested that ramp occurrence and morphology are related to the strength and frequency of the swash generated by storm waves, to waves of translation that sweep across the platforms, and to the presence of abrasive material at the cliff foot. In northeastern England, ABRASION accomplishes rapid erosion, ranging up to 30 mm yr^{-1}, on the steeply sloping ramp where there is a sand and pebble beach, whereas dessication of the shale is dominant on the more gently sloping platform (Robinson 1977). In some places the occurrence of ramps appears to reflect variations in rock structure and lithology. The presence of thick shale beds and other weak material near the high tidal level seems to be particularly suitable for the development of prominent ramps in eastern Canada and in northeastern England, and this is supported by mathematical modelling, which suggests that ramps are most common where rapid erosion produces wide intertidal platforms. Where contemporary rates of erosion are low, however, as in northwestern Spain, ramps extending up to several metres above the modern high tidal level are the result of higher SEA LEVEL during the last interglacial (see ICE AGES) (Trenhaile *et al.* 1999). In southern Australia, sloping ramps, which extend up to more than 10 m above present sea level, are probably polygenic, having developed under rising and falling sea level during the Cenozoic Era (Young and Bryant 1993).

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ALAN TRENHAILE

RAPIDS

Rapids in bedrock channels are not technically defined in fluvial literature, but imply steep reaches with rough water and very variable depth between lower gradient pools (Leopold 1969). Their origin is attributed to the erratic and episodic supply of boulders into the channel, both debris flows from tributaries and rock avalanches and rock fall from the valley sides (Howard and Dolan 1981; Webb *et al.* 1984). Subsequent accelerated flow through the constriction redistributes boulders downstream and partly reshapes the channel bed into quasi-stable form of boulder-strewn bars (Graf 1979; Kieffer 1987).

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KEITH J. TINKLER

RASA AND CONSTRUCTED RASA

The term rasas, of Spanish derivation, refers to old and perched littoral levelling surfaces or planation surfaces. Their width can reach several kilometres. The erosion surfaces are bordered inland by steep relief and by cliffs towards the sea. They were described for the first time by Hernandez-Pacheco (1950) on the Cantabrian Coast, northern Spain. Guilcher (1974) made a remarkable synthesis. These forms were also observed in Galicia (Nonn 1966), northern Chile (Paskoff 1970), southern Morocco, Brittany and Cornwall in England (Guilcher 1974), and Sardinia (Ozer 1986). Guilcher distinguished three types of rasas. The first one was described above, the second is more complex and is constituted by a succession of levellings arranged in stairs, and the third is when the passage towards the inland is gradual.

Plate 93 Rasa: Coast of Gallura (north Sardinia)

Plate 94 Constructed rasa: Coast of Anglona (north Sardinia). Accumulation of aeolianites on the terrace of the last interglacial sea level

Many of these rasas are covered by marine deposits (sand and rounded pebbles). These sediments were brought at a later date, during tertiary transgressions which only slightly retouched these levelling surfaces. Evidence of this process is found through ancient reefs in Brittany (Guilcher 1974), northern Sardinia (Ozer 1986) and south of Tangier, Morocco (Ozer, 2001 observation).

However, a convergence of shapes can exist, which is then called constructed rasas. This is a littoral aeolian accumulation, generally indurated (aeolianites), often mixed with local deposits of torrential origin. These accumulations are cut again in a shelf shape, slightly sloping towards the sea subsequent to runoff erosion.

The most spectacular constructed rasas are developed on slopes preceded by a well-developed continental shelf exposed to dominant winds. During Quaternary regressions, winds transported abandoned sands from the continental shelf until the first relief was formed by ancient cliffs which developed during the Quaternary transgressions. These deposits, essentially aeolian, became consolidated and were later shaped into cliffs by the current sea level. They are bounded inland by strong relief which is a previous Quaternary dead cliff.

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ANDRÉ OZER

RATES OF OPERATION

Rates of operation of geomorphic processes are determined in a number of different ways depending on the time and space scales of interest, and whether one is interested in rates of operation of individual processes or in the aggregate rates resulting from all processes combined.

The current dynamic tectonic conditions need to be considered alongside of the overall denudation rates in order to place measurement programmes conducted at site or watershed scale into proper perspective (Brunsden 1990). Brunsden notes that with respect to the major geotectonic provinces the Cenozoic orogenic regions, and especially the subduction areas on plate margins, can experience greater than 20 $\mathrm{mm\,yr^{-1}}$ of vertical movement at the same time as the overall denudation rate rarely exceeds 1 $\mathrm{mm\,yr^{-1}}$. At the other extreme, shields, platforms, cratonic regions and intracratonic basins experience less than 1 mm/1,000 yrs of vertical movement; nevertheless, overall denudation rates scarcely exceed 1 mm/10,000 yrs. Superimposed on these orogenic and epeirogenic movements are the isostatic readjustments which occur in regions recently emerged from under thick ice sheet cover. In the cratonic regions of the Baltic Sea and Hudson Bay, rates of isostatic readjustment were as high as 1–10 m/100 yrs at the close of the Wisconsinan glaciation and remain as high as 10 $\mathrm{mm\,yr^{-1}}$ in the Gulf of Bothnia. The implication drawn from these data corresponds closely with that of Schumm (1963) namely that 'the style and location of landform change is determined by the type, location and rate of tectonic movements and their associated stress fields over the relevant time and space framework of the landform assemblage' (Brunsden 1990: 3). There is general agreement on the order of magnitude of these rates at global to regional scale; the extent to which they are relevant to site and watershed scale is open to debate.

Average rates of operation of processes can obscure the fact that many processes are episodic and that land surfaces may evolve in a series of step jumps, with periods of relative stability followed by brief periods of rapid erosion or accelerated uplift. Variations in rates of change through time are further complicated by variations in space. Even within geotectonic provinces, spatial variations can be large.

Fundamentally, landscape stability and rates of change depend upon the ratio of resistances to change to the forces promoting change. Where these forces are in balance, little change occurs; where resistance exceeds the forces of denudation, weathering processes permit the deepening of soil profiles. This condition is called transport-limited. Where the forces of denudation are greater than the landscape resistance, erosion removes soil and weathering products as quickly as they are formed. This condition is called weathering-limited.

It is apparent that erosion rates will depend in large measure on the availability of transportable soil and sediment. As soil can only accumulate to considerable depths under stable conditions and eroded sediment can only accumulate at regional scale under conditions of continental-scale glaciation, the most extreme erosion rates occur when there is a marked change from one set of processes to another. When a threshold between one set of processes and another is crossed, extremely high rates of denudation may occur. The time period over which these accelerated rates can last is limited by the supply of readily eroded soil and sediment. Paraglacial geomorphology is one striking example of accelerated erosion and sedimentation following threshold exceedance. Landscape sensitivity to change is therefore as effective in controlling the short-term denudation rate as is the energy of the processes of erosion and transport.

In a brief historical sketch of the development of interest in rates of operation of geomorphic processes, Archibald Geikie and Charles Darwin are two of the early researchers who attempted to determine the rates of operation of individual processes. Geikie estimated the rate of rock weathering by measuring changes on dated tombstones in Edinburgh churchyards and Darwin estimated the rate of soil movement on slopes caused by worm casting. The first spatially representative estimates of the overall rate of ground loss derived from a summary of river sediment loads in the United States. Early twentieth-century estimates of the rate of cliff retreat in Germany on sandstones and in Brazil on granites under rainforest found that the rates in Brazil were an order of magnitude greater than those in Germany. Seasonal rates of movement of stones on talus in the Alps and longer term integrations of postglacial creep of till (135 m in 30,000 years) and Lester King's estimates of the rate of retreat of the Drakensberg scarp in South Africa (240 km in 150 million years) were some of the few quantitative rates of erosion estimated before the 1950s. No one seems to have correlated these data as they were simply too scattered and lacking in formal methodology. One notable exception was the US Soil Conservation data. The first systematic programme to measure soil erosion came about in the United States during the 1930s when one of the New Deal programes of President Roosevelt, intended to stem the growth of unemployment, resulted in the construction of tens of thousands of small dams by the US Soil Conservation Service. Large data sets of volumes of sediment delivered to small reservoirs thereby became available. Accelerated erosion plots usually included an adjacent control plot to demonstrate the negative effects of poor land use practices. From a strictly geomorphic perspective, the control plot data gave indications of spatial variability of surface wash rates, but integrated analyses were not published until the late 1940s. One of the important theoretical contributions from the US Soil Conservation data was the formulation of the dynamic concept of sediment sources. There was a recognition of the difference between sediment sources and sediment delivery at the outlet of each basin and the sediment delivery ratio became a useful tool to determine sediment storage. The 1950s were a decade of pioneering studies on rates of geomorphic process, all the way from sediment budgeting (Jackli 1957; Rapp 1960; Leopold *et al*. 1966) to surface wash (Schumm 1956) and a variety of creep processes (Jahn 1961).

By 1983, Saunders and Young summarized (somewhat uncritically) literally thousands of reported data on rates of process operation. Data are no longer the problem but standardized data, both in terms of methods of collection and units of measurement, remain a serious problem. With respect to endogenic processes, England and Molnar (1990) summarized the major difficulties. Many reports of surface uplift in mountain ranges are based on mistaking exhumation of rocks or uplift of rocks for surface uplift and provide no information whatsoever on the rates of surface uplift. Some observations provide reliable measures of the uplift of rocks but, because erosion rates may be high, the mean surface elevation may be decreasing while the rocks are uplifting.

Standardization of data

How does one compare (a) the linear downslope movement of the uppermost layer of the regolith with (b) the volumetric downslope movement of the whole regolith with (c) the slope retreat or ground loss perpendicular to the ground surface with (d) the mass of sediment transported past a control section with (e) the bedrock mass uplifted above the geoid surface? These are all common ways of reporting the results of contemporary process measurements. Caine (1976) stated the problem coherently. Not only is there a problem of the use of disparate units and dimensions, but there is a need to define hillslope erosion and river channel erosion in terms that are mutually compatible, and storage effects within river systems should also be accounted for. His solution is the calculation of a unit of geomorphic work which incorporates the product of the mass of sediment, the change in elevation and the gravitational acceleration. The approach is logically compelling but has not been widely adopted.

An alternative solution has been to convert all data to a linear measure of denudation distributed evenly across the basin. The Bubnoff unit (Fischer 1969), which is equivalent to 1 mm of denudation per 1,000 years, has also encountered some resistance, partly on account of the somewhat arbitrary specific gravity and packing corrections that have to be made, but also because of the impression created of even denudation across a highly spatially variable surface. It seems fair to say that the prevailing attitude is to maintain different units of measurement for slope, channel and basin data.

Equilibria between hillslope erosion and sediment yield

A number of studies have engaged the question of the quantitative balance between hillslope erosion or contemporary uplift and sediment yield. Here we consider just two examples of apparent balance between measured rates. Adams (1980) examined the Southern Alps of New Zealand and compared rates of crustal shortening, tectonic uplift, river sediment and dissolved load, and offshore deposition. In billions of kg yr^{-1}, the rates were respectively of the order of 700, 600, 700 and 580. Data on crustal shortening derived from geophysical estimates of the rate of convergent plate motion across the Indian–Pacific plate boundary, amounted to about 22 mm yr^{-1}. This process would lead to a build-up of crustal lithosphere. Data on tectonic uplift were calculated by converting the shortening to uplift along the Alpine Fault. Data on river loads were taken from water analyses (dissolved load), estimates from formulae and field measurement (bedload) and monitored data supplemented by runoff vs sediment concentration relations (suspended load). The average amount removed was adequate (on an annual basis) to balance the build-up effect from tectonic uplift. Finally, data on offshore deposition showed similar order of magnitude effects, thereby removing sediment to the east and west to the converging plate margins. The model described by Adams is a steady-state mountain range with rapid uplift being balanced by rapid erosion. The details are contentious, but the example is instructive in that it demonstrates the extensive data demands placed on such an interpretation. The author is fully cognizant of the errors inherent in the calculations. He confirms his findings in an interesting appeal to the shapes of New Zealand's mountains. The Southern Alps are spiky mountains (suggesting a steady-state condition) whereas immediately adjacent, in Otago, the mountains are flat-topped and are the remains of a pre-uplift surface of low relief.

Reneau and Dietrich (1991) examined a part of the southern Oregon Coast Range and compared data on bedrock exfoliation rates, thicknesses and dates of accumulations of colluvial fill in topographic hollows and the size of the contributing source area with monitored suspended and dissolved load data from the region. The novelty of this approach derives from some premises with respect to the effectiveness of topographic hollows in trapping colluvium and the ability to satisfactorily date the colluvial fill at up to five stratigraphic levels. If it be admitted that colluvial transport rates down the axis of a hollow are dependent on gradient and are constant in the part being evaluated, then net deposition is entirely due to colluvium added from the adjacent side slope. Calculations of volumetric colluvial transport rates into each hollow involved using measures of local topographic convergence, average soil density and the mass depositional rate of colluvium. Calculated average erosion rates from dated hollows were equivalent to about 70 Bubnoffs (mm/1,000 yrs); calculated exfoliation rates were equivalent to about 90 B and calculated denudation rates varied from 50–80 B. Again, the authors carefully identify error bars on their data but conclude that because hillslope and basin-wide erosion rates are so similar hillslope sediment production and stream sediment yield in the Oregon Coast Range are roughly in balance. Net changes in sediment storage downstream are necessarily also minor. Again it should be noted that the data needs are onerous and creative field measurement programmes are necessary.

By contrast with rates of geomorphic process in apparently steady-state environments, relatively few measurement programmes on rates of bedrock incision have been reported.

Whipple *et al.* (2000) took advantage of the diversion of the upper Ukak River in Alaska by an ash flow in 1912 to measure rates of incision along a newly formed bedrock channel. Although the minimum rates of incision are high (10–100 $\mathrm{mm\,yr^{-1}}$), they are within the range of previously published estimates (e.g. Stock and Montgomery 1999). In this branch of process geomorphology there are substantially more modelled rates of operation of process than confirmed field data.

There remains considerable ambiguity over the significance of measured rates of erosion at site scale and over short periods of time vis-à-vis the evolving shape of the landscape. During the 1960s there was optimism that measured rates might be extrapolated from site scale and from short-term measurements to larger landscapes and longer term rates. Such expectations have been shown to be naive and derived in part from assumptions about equilibrium, a balanced condition and the ignoring of contingent environmental constraints. Perhaps the central question now being engaged is that of how to link measurements of rates of operation of geomorphic process at one scale (whether temporal or spatial) to another scale. The information is urgently required in the context of concerns about global environmental change (at what scales are the effects of human activity clearly differentiable from the effects of climate change?) and also in the context of a better understanding of Earth history.

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SEE ALSO: Bubnoff unit; chemical denudation; denudation

OLAV SLAYMAKER

REDUCTION

Reduction is the gain of a negative electron so an element becomes less positively charged, for example ferric iron, Fe^{3+} (or Iron III) becomes reduced to ferrous iron Fe^{2+} or Iron II. This process commonly occurs in the absence of oxygen but can equally occur when iron is in an acid solution. The latter process accounts for the solubilization and loss of iron oxides in the upper parts of soil profiles where there is, in fact, oxygen available but where organic acids derived from the decomposition of plant material acidifies the soil.

In geomorphology, the focus is on the mobilization of iron under reducing conditions and its transport in anoxic/acid waters, often in deep ground water, and the redeposition of iron oxides in oxic conditions. Retallack (1992) proposes that a study of fossil soils shows how the Earth's atmosphere evolved with the gradual increase in oxygen due to the rise of the plants. Around 1,000 million years BP virtually all palaeosols contained oxidized iron, but palaeosols with reduced iron present occur before that date and 3,000 million years ago there were very few paleosols with oxidized iron present.

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STEVE TRUDGILL

REEF

Reefs can broadly be classified as spatially heterogeneous, three-dimensional structures which have morphological form that is different from that of the underlying substrata. Historically and currently the term reef has been used to classify a whole host of organic and inorganic structures including stone reefs, OYSTER REEFs, CORAL REEFs, ATOLLs, SERPULID REEFs, algal reefs and artificial reefs. Due to the range of disparate features being classified as reefs, there has been much debate in the literature over what does and does not constitute a reef.

Although many reef specialists, both biologists and geologists, have argued that reefs must be of biogenic origin to be classified as reefs, numerous applications of the term reef have been applied to inorganic structures. In the late nineteenth and early twentieth centuries, emminent natural scientists and geologists referred to inorganic structures, such as beachrock or the curious bar at Pernambuco, Brazil, as stone reefs (e.g. Branner 1905). More recently, there has been a resurgence of the use of the term reef for inorganic structures, as artificial reefs. In many coastal environments, artificial reefs are being built for a range of purposes including as offshore coastal defences, such as those at Sea Palling, Norfolk, England; or as subtidal structures designed to enhance biodiversity in inshore waters. From a geomorphological perspective, both organic and inorganic reef structures influence geomorphological processes and the morphology of coastal environments. As such, both organic and inorganic reef structures are classified as reefs for geomorphological research purposes.

Smaller reefs have often been termed bioherms and biostromes: bioherms are reef-like, mound-like or lens-like features of purely organic origin which are found embedded in rocks of different lithologies, while biostromes are organic layers, which are thinner and less developed structures than bioherms, such as oyster reefs (Cummings 1932).

Importantly, Cummings was one of the first authors to stipulate that reefs are organic forms which can be produced by several different species and they exhibit a variety of forms, ranging from reefs to bioherms and biostromes, where corals are only one type of reef form. Although this subdivision of reefs into more specialized categories had many merits, modern authors still preferentially use the term reef.

Reefs are found in temperate to tropical marine ecosystems, with the most prominent reef types, corals and atolls, being found in tropical and subtropical zones. Algal reefs and bioherms are commonly found in more moderate climatic zones, such as the Mediterranean, and include corniches, trottoirs and mini-atolls built primarily by calcarous algae, vermetids and serpulids. In temperate regions, reefs are often more like bioherms or biostromes in structure and include reef communities such as *Sabellaria*, oyster or skeletal carbonate reefs. Temperate reefs are found in the eulittoral to pelgaic zones and typically develop on a firm substrata. Reefs can enhance the growth and persistence of other species, by providing sheltered habitat or by providing a fixed substrata upon which cryptic communities can colonize and they can also influence sediment dynamics by trapping and storing sediment.

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LARISSA NAYLOR

REGELATION

Means to 'freeze again'. In the glacial context it refers to those processes which permit a glacier to slide over a rough bed by means of melting on the upglacier side of an obstacle and to refreeze on the downglacier side. Regelation occurs because the greatest resistance to glacier movement is on the upstream side of an obstacle. This results in locally high pressures and a consequent lowering of the pressure melting point. Thus melting of ice occurs immediately upglacier of the obstacle, and the resulting meltwater migrates to the lower pressure zone on the downglacier side of the obstacle. There it refreezes because the pressure melting point is higher. It is through this mechanism that the ice in effect overcomes the obstacle by temporarily turning to water and back again. It is, therefore, an important process in glacier sliding and has been confirmed by direct observation in subglacial cavities.

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A.S. GOUDIE

REGOLITH

The term was coined by Merrill (1897) to describe an 'incoherent mass of varying thickness composed of materials essentially the same as make up the rocks themselves, but in greatly varying conditions of mechanical aggradation and chemical combination'. He went on to point out regolith may be formed *in situ* or from sediments transported from another source.

Merrill derived the word from the Greek regos ($\rho\eta\gamma o\sigma$) meaning blanket or cover and lithos ($\lambda\iota\theta o\sigma$) meaning rock or stone. Jackson (1997) defines regolith as a term for 'the layer or mantle of fragmental and unconsolidated material, whether residual or transported and highly varied in character, that nearly everywhere forms the surface of the land and overlies the bedrock'. Another more simple definition is everything that lies between fresh rock and fresh air.

Regolith is restricted to terrestrial environments and is generally considered to comprise mechanically and chemically weathered rock debris whether *in situ* or transported. It includes rock weathered to varying degrees, sediments of colluvial, alluvial, aeolian, marginal marine and glacial origin as well as volcanic ash and lag gravels, pisolites and sand. It ranges from soft and loose to consolidated and/or cemented and very hard.

Regolith is the earth material usually called 'soil' by many scientists and engineers. Engineers tend to call any earth materials that can be moved with a bulldozer or mechanical digger soil. Forensic geologists call their sampling media soil. Agricultural scientists on the other hand think of soil as a growing medium from crops and pasture. To a regolith scientist soil is a part of the regolith at the uppermost part of the whole body of unconsolidated material they call regolith. Regolith also may contain buried soils that formed during periods when little accretion occurred in an accretionary (sedimentary or volcanic) landscape.

Both Merrill and Jackson consider regolith to be unconsolidated, but when duricrusts are considered this concept falls down. A silcrete for example is as tough a rock as one can find, but it is considered by most to be part of the regolith. Equally in many parts of the world lava flows are encapsulated by regolith (Figure 130). Does this mean that the lavas are part of the regolith or are there two different regolith units above and below the lava? In Figure 130 it is clear that at section A this is the case, but laterally in the section there is only one regolith, part a lateral equivalent of the one below the lava and one laterally equivalent to that above.

This dilemma raises the issue of regolith stratigraphy and dating. Within transported parts of regolith it is possible to apply the principles of lithostratigraphy remembering that this provides little in the way of chronological control on regolith materials. The lithostratigraphy in section A (Figure 130) is very different from that in section B. The age of weathering in the regolith unit below the lava may very well be very different from that above it. The age of weathering in section B will be complex because this section has been exposed to weathering for a longer time than the upper regolith unit in section A and probably has a complex weathering profile carrying components of pre- and post-lava weathering. Moreover because weathering occurs continuously, albeit at different rates, weathering overprints on regolith materials cannot be used for correlation unless dating of weathering demonstrates equivalence. Without dates on regolith materials this dilemma cannot be resolved except in a relative sense, and even then with some difficulty.

The age of regolith does not form part of its definition, but many would consider that Palaeozoic (or even older) materials now at the surface are not regolith but exhumed surfaces on which some regolith is preserved. In many parts of the world ancient regolith exists at or near the modern surface. In some cases it is unlikely that these surfaces and materials have ever been buried (Craig and Brown 1984). In other cases they were buried and have since been exhumed (Lidmar-Bergström 1995). Carboniferous weathering profiles have been dated by palaeomagnetic methods within 1–2 m of the surface in central

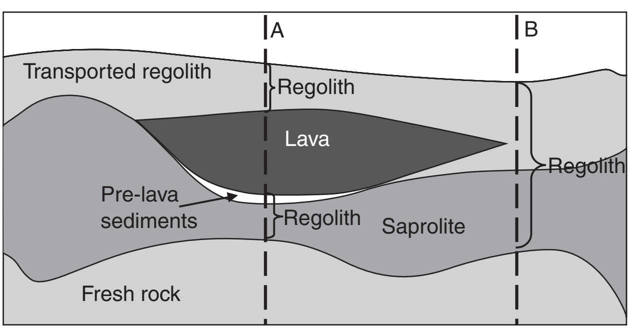

Figure 130 Is it logical to include the lava in with regolith or define it as detached and possibly part of the fresh rock even though it may be a different age and composition?

New South Wales, Australia (Pillans *et al*. 1999), but it has been suggested this profile is exhumed several times, 3.5 km during the Permo-Carboniferous and another 2.5 km during the Triassic to early (O'Sullivan *et al*. 2000). Most regolith however is very much younger than the Palaeozoic and it is still forming across the Earth's surface.

In situ regolith generally forms a weathering profile and these profiles often have a characteristic sequence of materials developed in them. Taylor and Eggleton (2001) provide detailed descriptions and interpretations of weathering profiles. Essentially the sequence is:

- soil
- ferruginous and/or aluminous lag
- collapsed saprolite (may be mottled by ferric oxihydroxides)
- saprolite mottled by ferric oxihydroxides
- bleached saprolite (composed of kaolinite and/or quartz grading downward into more complex clay minerals and quartz ± other primary minerals)
- saprock
- weathering front
- fresh rock.

Weathering profiles of this type are often considered to be the norm and if the upper parts of the profile (e.g. the lag and/or collapsed saprolite) are not present it is often inferred that there has been erosion. This is a misguided inference as there is often no evidence to suggest that the profile was completely developed or that it ever had all those components. Such inferences can lead to erroneous conclusions regarding landscape evolution and the formation of various regolith materials.

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GRAHAM TAYLOR

REJUVENATION

Rejuvenation stems from *juvenis*, Latin for young. Thus rejuvenation is to make young again. The term has been applied to individual landforms such as a hillslope or a river channel, but it is most commonly and more appropriately applied in the context of the entire landscape. The term enjoys wide usage among physical geographers and historically based geomorphologists. Its origin and usage in geomorphology can be traced to the interpretation of several lengthy philosophical discourses in the late nineteenth century when some of the major paradigms of long-term landscape evolution were first established (Davis 1889, 1899).

The geographic cycle of Davis (1899) (see CYCLE OF EROSION) continues to influence modern thoughts on long-term landscape evolution. Davisian theory explains landscapes and their constituent landforms primarily in the context of the amount of time that they have been subjected to the forces of erosion. Landscapes are viewed as being born from impulsive rock uplift above sea level. This uplift is followed by a protracted period of erosion that lowers the mean elevation of the landscape by first incising deep, narrow valleys, then widening the valley bottoms and rounding the hillslopes, finally leading to the decline of interfluves to the point that the entire landscape has been reduced to a flat plain or PENEPLAIN. During the valley incision stage, the landscape is traditionally described as youthful, in the valley widening and hillslope rounding phase, the landscape is thought of as mature, and as a peneplain, the landscape is thought of as old. Davis (1899), as well as the subsequent generation of geomorphic thought, recognized that in reality, the geographic cycle almost never proceeded to completion creating a widespread peneplain. Rather, tectonism was understood to be frequent enough such that landscapes in various stages of maturity or old age were uplifted, increasing mean elevation, causing renewed stream incision, and effectively making the landscape appear young again. Such active tectonics has the effect of rejuvenating the landscape.

Rejuvenation is a useful concept when viewing landscape evolution over long (10^6–10^7 yrs) timescales, especially when the flux of sediment that is eroded from those landscapes is considered (Schumm and Rea 1995). Long-term sediment yield from landscape erosion tends to follow a decaying exponential relationship that records an initial, large erosion response in concert with the rock uplift, followed by a long period of time where the rate of erosion decreases as mean elevation and mean slope are reduced (Ahnert 1970; Pazzaglia and Brandon 1996). Impulsive increases in sediment yield over these timescales are probably correctly interpreted as some major change in the erosion processes and rates operating on a rejuvenated landscape imposed by renewed rock uplift, a change in climate, or both.

Unfortunately, use of the term rejuvenate has been extended to explain the forms and changes in individual components of a landscape over shorter timescales (10^0–10^5 yrs), but its applicability in this context is probably not correct. For example in the strict Davisian interpretation, a meandering river channel flowing in a wide river valley is a mature or even old landform whereas a steep river channel flowing in a narrow valley is a youthful landform. Individual landforms such as river channels are much better explained as a DYNAMIC EQUILIBRIUM expression between driving and resisting forces where form and process are mutually dependent. The meandering channel speaks more to the fact that the river has a stable discharge, primarily fine-grain size, gentle slopes, and stable, vegetated channel banks rather than its age in the geographic cycle. In fact, active meander channels in bedrock are known to exist in even the most rapidly uplifting landscapes such as Taiwan where there is no evidence that they have been superimposed or inherited from earlier forms (Hovius and Stark 2001; Hartshorn *et al.* 2002). Similarly, steep, narrow river valleys are common on the great ESCARPMENTs of the southern continents which are known to be among the most slowly eroding and changing landscapes on the planet (Bierman and Caffee 2001). The term rejuvenation is improperly used in these cases of attempting to explain relative landform age or changes in the landscape through an investigation of forms only, without consideration of process or tectonic setting.

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FRANK J. PAZZAGLIA

RELAXATION TIME

Geomorphological change may be envisaged as a set of responses to the varying frequencies and magnitudes of formative events at all scales (Graf 1977; Brunsden and Thornes 1979). The concept of LANDSCAPE SENSITIVITY to changes in the operation of controlling processes suggests three divisions of time (Brunsden 1980, 1990; Figure 131): the time taken to react to an impulse of change (lag or *reaction time*); the time taken to attain the characteristic state (*relaxation time*); and the time over which the form exists (*characteristic time* or *landform lifetime*) (McSaveney and Griffiths 1988). Relaxation time is an important measure because landforms can only reach a slowly changing (stable?) state if the interval between form-changing events is greater than the sum of reaction and relaxation times. If the interval is shorter then the landforms will be in a state of constant readjustment and strong flux. This state may be called *transient*.

A further application of the idea of relaxation is to define the term as '*recovery*'. After a severe or land forming event the more 'normal ' frequent events will seek to erase the landform or to modify the form until it is compatible with them.

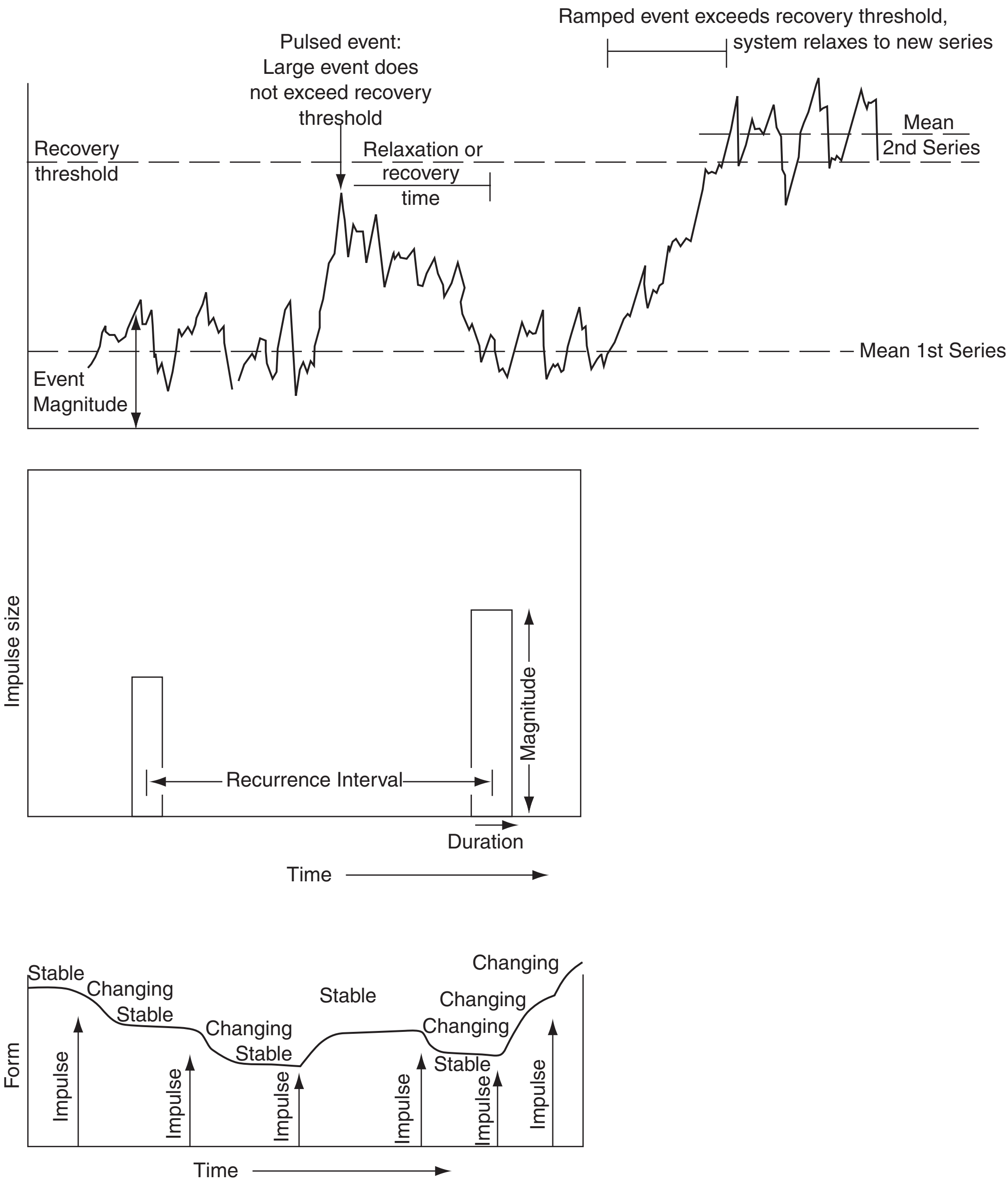

Figure 131 A schematic representation of the concepts of *reaction* (lag) time, *relaxation* (recovery, healing, form adjusting) time and *characteristic form* (form constant?) time

The process can also be an attempt to 'heal' the scars and to return the landscape to its former state. Crozier (1986; see also Crozier *et al*. 1990) regards this as a process of 'ripening' in which the landscape is again prepared for another effective event. This idea is usually applied to soil erosion and mass movement on hillslopes where hollows produced by these processes are weathered and infilled until critical depth is reached and failure can again take place (Deitrich and Dorn 1984, Deitrich *et al*. 1992).

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DENYS BRUNSDEN

RELIEF

Relief may be defined most generally as the elevation difference over a predetermined area or inferred length scale. This simple definition allows for specifying a number of particular types of relief. The relief of a mountain range, for example, may be considered as the difference in elevation between the highest peak and the base of the range front. Alternatively, the relief of a mountain range can refer to the absolute height of the highest peak, with the implicit reference to sea level as a datum. When defined over much shorter length scales, relief can be defined as the range in elevation spanned by a particular hillslope, ridge to valley transect, or physiographic feature such as an escarpment. Relief may also simply refer to topography in general, or more specifically to the collective elevations or their inequalities of a land surface. In other words, the term relief has a variety of possible meanings depending upon the context within which it is used. The most common use of the term, however, generally refers either to topography itself or to the elevation difference between the highest and lowest points within an area of interest.

Differences in elevation that produce relief arise from the interaction of spatial variations in rock uplift and erosion. Volcanic and tectonic processes that raise rocks above sea level are ultimately responsible for elevating mountain ranges, although normal faulting also may produce local relief in extensional settings. Erosional processes may limit the total relief maintained by rock uplift but also cut valleys and produce relief over shorter length scales. Fluvial and glacial processes that incise the landscape produce relief, whereas mass-wasting processes (such as soil creep and many types of landsliding) tend to reduce relief. The overall relief of a mountain range ultimately depends on the balance between uplift and erosion, unless accumulation of crustal material exceeds the mechanical limit supportable by crustal strength, leading to the growth of a high plateau.

Several kinds of relief can be used to describe different aspects of a drainage basin. Fluvial relief represents the elevation drop measured down the longitudinal profile of a river network, as given by the elevation difference between the channel head and the basin outlet. This portion of the total basin relief may be influenced by changes in fluvial processes and rates of river incision. Hillslope relief defined by the elevation difference from the channel head to the drainage divide at the head of the basin represents that portion of the total relief within a drainage basin beyond the immediate influence of fluvial processes and which is instead controlled by hillslope processes. The geophysical relief of a drainage basin has been described as the local elevation difference between the ridgetop and valley bottom, which consists of both hillslope and fluvial relief. In addition to these specific types of relief, local relief may be defined by the elevation difference between the highest and lowest point on the topography measured over an area of predetermined size or a proscribed length scale.

Local relief is inherently scale dependent. The larger the length scale over which it is measured, the larger the relief. Generally, local relief increases as a non-linear function of the diameter of the area over which it is measured, with an exponent <1 and typically about 0.7 to 0.8. In addition, mean local relief is strongly correlated with mean local slope. But in comparison to mean slopes, which have a strong grid-size dependence, mean local relief is less grid-size dependent when calculated from digital elevation models.

Fundamental relationships between relief and erosion rates have been posited since early workers argued that greater relief and steeper slopes lead to faster erosion. In one of the first modern studies of the influence of relief on erosion rates, Schumm (1963) reported a linear relation between erosion rate and drainage basin relief (the height above sea level of the highest point in the basin) for large North American drainage basins. Ahnert (1970) subsequently reported that erosion rates increase linearly with mean local relief (the difference in elevation measured over a specified length scale) for mid-latitude drainage basins. Later studies bolstered Ahnert's relation with data from other regions and showed that local relief and runoff are dominant controls on erosion rate for major world drainage basins (e.g. Summerfield and Hulton 1994). Different relations between erosion rates and mean elevation characterize tectonically active and inactive mountain ranges (Pinet and Souriau 1988), and Montgomery and Brandon (2002) recently reported evidence for a strongly non-linear relation between long-term erosion rates and mean local relief.

Until relatively recently, the relief of bedrock hillslopes was thought not to be strength limited because of the great cohesive strength of intact rock. But the development of discontinuities in rock strength at the scale of an entire hillslope, valley side, or mountain can limit relief development through large-scale bedrock landsliding (Schmidt and Montgomery 1995). The catastrophic 1991 failure of the crest of Mt Cook – the highest point in New Zealand – illustrates how bedrock landsliding can limit relief in steep, highly dissected terrain. Arguing for the generality of strength-limited hillslope relief, Burbank and others (1996) demonstrated that the gorge of the Indus River had strong gradients in incision rate through a region where mean hillslope gradients are independent of the local river incision rate. Hence, they concluded that the development of strength-limited hillslopes allowed bedrock landsliding to efficiently adjust slope profiles such that ridgetop lowering keeps pace with rapid bedrock river incision. This emerging view of the role of relief on erosion rates holds that in steep tectonically active regions erosion rates adjust to high rates of rock uplift primarily through changes in the frequency of landsliding rather than increased hillslope steepness or increased relief (Montgomery and Brandon 2002). In contrast, in lower gradient landscapes the steepness of hillslopes, and therefore local relief, may respond to changes in the controls on landscape-scale erosion rates.

Climate setting and variability constrain the total relief of mountain ranges. Highly orographic rainfall variability can either limit or increase the fluvial relief depending upon the nature of the feedback operating in a specific mountain range (Roe *et al*. 2002). Enhanced erosion by glaciers and periglacial processes can preclude development of relief substantially above the perennial snowline (Brozovic *et al*. 1997). The role of erosion in reducing mass accumulation in mountain ranges is perhaps best illustrated by the exceptional cases where lack of rainfall allows mass accumulation to engage the mechanical limit to crustal thickening and results in development of high plateaux like the Altiplano and Tibet. The position of Earth's high plateaux in the dry latitudes suggests that plateau formation reflects the coincidence of high rates of tectonically driven mass convergence and low rates of erosion due to an arid climate.

A new view of the coupling and feedback among climate, erosion and tectonic processes is coalescing from recent studies focused on their interactions. Geologists are recognizing that spatial gradients in the climate forcing that drives erosion can influence the development and evolution of geologic structures. Development of mountain ranges strongly influences patterns of precipitation and numerical simulations of evolving and steady-state orogens show that both the topography and the resulting metamorphic gradients exposed at the surface reflect the influence of spatial variability in erosion (Willett 1999). Gradients in climate and tectonic forcing strongly influence erosional intensity, and this interaction in turn governs the development and evolution of topography. Hence, the development of relief is strongly coupled to large-scale feedback involving the interplay of climate, erosion and tectonics. Whereas the geographical distribution of plate tectonic environments has changed over geologic time, the global pattern of climate variability exhibits robust latitudinal patterns characterized by abundant and intense rainfall in the equatorial tropics, a low latitude belt of deserts and stronger glacial influences toward the poles. In this context, the feedback between climate, tectonics and erosion implies large-scale climatic controls on the global distribution of topography; high plateaux are likely to form astride the desert latitudes, whereas high mountains are unlikely to form in the equatorial or polar regions where erosion rates are high due to either intense rainfall or glacial processes.

Substantial debate has centred on the relation of climate change to the relief of mountainous topography. An increase in the absolute relief of a mountain range, and consequent increase in the area of alpine environments, can increase rates of weathering through rapid mechanical breakdown of fresh rock by periglacial and glacial processes. Hence, increased relief in alpine areas potentially could influence global carbon cycles and large-scale climate. Wager (1933) noted the proximity of high Himalayan peaks to deep valleys and proposed that isostatic rebound in response to valley incision was responsible for elevating Himalayan peaks above the Tibetan Plateau. Molnar and England (1990) proposed that much of the evidence for substantial late Cenozoic uplift of mountain ranges may simply represent the effect of climatic deterioration on increased erosional exhumation of rocks or the uplift of mountain peaks in response to deepening and enlargement of valleys. Analyses of valley geometry show that such an effect could account for up to about a quarter of the elevation of mountain peaks, although the potential magnitude of such an effect depends on the strength of the crust and the nature of erosional processes (Gilchrist *et al*. 1994; Montgomery 1994). However, recent studies have concluded that there is minimal potential for valley incision to substantially influence local relief in tectonically active mountain ranges (Whipple *et al*. 1999; Montgomery and Brandon 2002).

In summary, relief is a simple concept with many variants of meaning that depend on the specific context in which it is used. Nonetheless, an understanding of the controls on relief generation is central to understanding the linkages between geomorphic processes, tectonics and climate that together shape the Earth's surface.

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DAVID R. MONTGOMERY

RELIEF GENERATION

Almost everywhere landforms are composed of elements evolved under different climates, i.e. shaped by different exogenetic forces. The relevant form assemblage is called relief generation. They are the constituents of CLIMATO-GENETIC GEOMORPHOLOGY. This concept has no relation to the stages of the Davisian CYCLE OF EROSION and its DENUDATION CHRONOLOGY, which are based on tectonics.

In central Europe, rumpfflächen (etchplains, plains cutting rocks of different hardness; see ETCHING, ETCHPLAIN AND ETCHPLANATION) came into existence in Tertiary time as is concluded from the form, deposits and relics of tropical weathering. Into these plains valleys are incised, starting with broad terraces of Pliocene/lower Pleistocene age with almost pure quartz gravels, sometimes with a few pisoliths. Obviously they are the eroded and transported result of tropical weathering. The flight of terraces in the middle and lower part of the valleys is of periglacial origin as is proved by pebbles from different rocks, syngenetic permafrost features, and LOESS or dunes (only Würm) on top. A solifluction cover on almost every slope, as strata in which the recent soils are developed, shows the overall small amount of Holocene erosion. This applies too for fluvial processes as the floodplain is only about 3 m below the Würm terrace and the incision was mainly in the late Würm, or early Holocene. The term relief generations was introduced by Büdel in 1955 in a paper about the Hoggar in the central Sahara. Here the rumpfflächen carry red loam, relics of tropical soils, under dated basalt flows. A loams terrace in the valleys does not correspond with the recent processes, and has very old artefacts on top. The recent river bed consists of sand.

These examples show the main methods to distinguish relief generations: (1) separation of landforms of different origin as the younger ones are nested or incised into the older; rarely the younger form is on top of the older ones as for instance dunes; (2) observation of the recent processes thus delineating the recent forms; (3) search for weathering relics and/or correlated sediments giving an indication of a different climate and linking these to the older relief generation. If a complete picture is derived by observations in the field, perhaps added to by laboratory analysis, one then tries to compare the forms of the relief generation with similar forms in a different climatic zone. Absolute datings are helpful as they provide an age, which, via the geologic timescale, gives an idea of paleoecological conditions.

The basis for the comparison is 'Klimatische Geomorphologie', which is quite different from CLIMATIC GEOMORPHOLOGY, which studies the landform assemblage and the relative importance of the recent processes in morphoclimatic zones. A regular assemblage of landforms presents the chance to classify relic forms that are only sporadically preserved and to search for additional forms. For example, if one observes overdeepening and glacial striations in rocks, a wall of mixed deposits in a certain position, then it is most likely a moraine. This can be backed up by looking for drumlins or other features of glacial erosion. The concept of relief generations is broader than investigating palaeoforms, for it asks for form assemblages and their relief forming mechanisms. After all these investigations one might ask for palaeoclimatic data from e.g. palaeobotany for comparison. It has almost never been tried to fix recent climatic data to the boundaries of regions with different relief generations.

The relief forming processes of different climates in Tertiary to recent times might also be seen at the small scale. Blocks in blockfields in the Harz Mts have a red rind from weathering in the Tertiary red loam. This is topped by a small white rind. On several blocks a triangular or square piece has been split off by frost weathering. Here the edges of the block have a white rind only. On the rim of the blockfield further blocks are uncovered by recent wash as most probably happened during warm periods of the Pleistocene, too. Thus these fields may be called 'Mehrzeitformen' (multitude forms of different climates). Ayers Rock has weathering forms in the hard rock, which are nested and which were formed in different climates.

For methodological discussions relief generations should be the basis for the distinction of relief elements of different sensitivity to recent geomorphological processes. Elements of older relief generations are more stable than younger ones, and they may be eroded mainly by valley incision. The strength and place of occurence of thresholds can be explained by relief generations, e.g. the edge of an old plain. The elements of older relief generations are certainly not in equilibrium with recent processes. They have not been formed by them nor are they considerably changed by them. Equilibrium is almost never provable even for the recent generation for two reasons: the influence of older generations on the discharge and water movement paths, and the different resistance of existing forms. The ergodic principle can only be applied to relief forms of one generation, preferably those which change fast. Thus the ergodic principle is limited in time.

Further reading

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HANNA BREMER

REMOTE SENSING IN GEOMORPHOLOGY

Remote sensing is the acquisition of information about an object without physical contact. In geomorphology, remote sensing often implies the collection of information from aerial platforms (e.g. airplanes, balloons or kites) or from spacecraft orbiting the Earth. The term remote sensing is credited to Evelyn L. Pruitt and her staff in the United States Office of Naval Research. It was coined during the early 1960s in recognition that instruments other than cameras and regions of the electromagnetic spectrum outside those visible to the human eye and to which photographic film is sensitive were increasingly being used to image the Earth. The current American Society of Photogrammetry and Remote Sensing's (ASPRS) definition of photogrammetry and remote sensing reads 'the art, science, and technology of obtaining reliable information about physical objects and the environment through the process of recording, measuring, and interpreting imagery and digital representations of energy patterns derived from noncontact sensor systems' (Colwell 1997: 3).

While remote sensing will not replace the traditional geomorphic field study, the value of remote sensing to provide a synoptic overview of a landscape cannot be overlooked. Historically, the use of remote sensing in geomorphology has been mainly interpretive, enabling geomorphologists to develop a 'mental picture' of the landscape and as a map-making aid (Hayden *et al*. 1986). However, the use of remote sensing for quantitative geomorphic study is growing rapidly.

Remote sensing provides unique global views at different spatial scales and in different regions of the electromagnetic spectrum. These global views are extremely useful for the subdiscipline of megageomorphology, which emphasizes the study of planetary surfaces at large scales (Baker 1986). The global views provided by remote sensing are not static, but are being continuously refreshed. This repeated global monitoring captures geomorphic events that might otherwise go unnoticed. For example, in 1983 astronauts aboard the STS-8 Space Shuttle mission photographed a dust storm over northwestern Argentina that was transporting material from exposed salt flats on the Puna Plateau of the South American Andes eastward toward the Argentine Pampas. These remote observations helped confirm the source of the Pampas loess and demonstrated that silt accumulation on the Pampas is an ongoing geomorphic process (Hayden *et al*. 1986). Remote sensing can also enable geomorphic study of areas that are inaccessible to field-based investigations.

Remote sensing provides a unique historical archive of geomorphic change. Aerial photography suitable for geomorphic analysis began to be collected as early as the 1920s. The satellite image archive suitable for geomorphic analysis began with the launch of the first of the Earth Resources Technology Satellite satellites ERTS-1, later renamed to Landsat-1, on 23 July 1972.

The history of remote sensing as a tool for geomorphic analysis is intimately tied to advances in photography and the acquisition of photographs from aerial platforms. Photography was born in 1839 when the photographic processes developed by Joseph Nicephore Niepce, Louis Jacques Mande Daguerre and William Henry Fox Talbot were publicly disclosed. One year later the use of photography to aid in the development of topographic maps was advocated by François Arago, Director of the Paris Observatory (Fischer 1975: 27). The first aerial photograph was taken by Gaspard Felix Tournachon, also known as Nadar, from a balloon outside Paris, France in 1858. However, recognition of photography's value to geomorphology also arose in part from terrestrial photographs of the landscape taken during the latter half of the nineteenth century. In 1890, the Geological Society of America formed a Committee on Photographs and its first report described photogeology (Fischer 1975: 34) which as a science took shape in the 1920s and 1930s. The basis of photogeologic analysis rests on the simple notion that landforms developed under similar geologic and geomorphic processes will appear similar in remotely sensed images (Way and Everett 1997: 117). By the early 1940s, photo interpreters were able to recognize the distinguishing surface features of approximately thirty-five major landforms and realized that aerial photographs provided important information on the origin, composition and history of landforms (Colwell 1997: 26). These landforms, and other features, can be identified in remote sensing images based on their location, size, shape, tone and/or colour, shadow, texture, pattern, height/depth as well as site characteristics and associations among features in the landscape (Jenson 2000: 121–132). The first systematic, although low resolution, observation of Earth from satellite began in 1960 from TIROS I, the world's first meteorological satellite. Since then, remote sensing of the Earth has expanded significantly as spaceborne imaging systems have grown in number and in sophistication.

Successful application of remote sensing images to geomorphic study requires careful matching of an instrument or image's spatial, temporal and spectral characteristics with the requirements of the geomorphic study at hand. A sensor's spatial resolution is the smallest angular or linear separation that it can resolve. The resolution of digital images acquired by non-photographic instruments is often described by the length of one dimension (in metres) of the individual elements (pixels) that comprise the two-dimensional image. In determining the required spatial resolution required for a particular application, a useful rule of thumb is that for an instrument to detect a feature of a certain size, its spatial resolution should be at most one-half of the feature's smallest dimension.

The temporal resolution of a remote sensing system refers to how frequently it can image a certain area. Most orbital sensors have fixed repeat cycles which control how often an area is imaged. They typically range from less than one day to one or two weeks. Aircraft overflights or manned space missions usually acquire images much more infrequently and at much more irregular intervals.

Various wavelength regions of the electromagnetic spectrum provide quite different information about the chemical, physical and biological properties of a landscape. The two most commonly used regions of the electromagnetic spectrum for geomorphic study are the optical, where the propagating electromagnetic energy has wavelengths of 0.3 to 14 micrometres (μm, 1 μm = 10^{-6} m), and the microwave with millimetre to metre wavelengths.

The optical region, which historically has been the most widely used, can be divided into two subregions, a reflected optical region (0.3–3.0 μm) and a thermal infrared region (3.0–14.0 μm). Remote sensing instruments operating in these two wavelength regions are typically passive; the energy supplying the signal to the sensor comes from an external source. In the reflected optical region, solar energy reflected off the landscape provides the signal while in the thermal infrared region, energy emitted directly by the Earth itself as a function of its surface temperature is the primary energy source. These two spectral regions contain numerous atmospheric windows or wavelength intervals in which the atmosphere is fairly transparent to solar energy or emitted energy making remote observations possible. While the atmosphere may be fairly transparent in these windows, for some geomorphic applications correction for atmospheric effects still may be important. Cloud cover can also obscure the surface over all wavelengths in the optical.

The reflected optical wavelengths are the most commonly used in terrestrial remote sensing. The reflected optical is typically subdivided into three spectral regions: the visible (0.4–0.7 μm) to which the human eye is sensitive, the near infrared (0.7–1.1 μm) and mid or short-wave infrared (1.1–2.5 μm). As the name suggests, reflected optical images are formed from the energy reflected from the surface towards the sensor. The reflectance of surface materials varies as a function of wavelength making it often possible to discriminate between different surface materials based solely on their reflectance. While two materials may appear similar at one wavelength they may be quite easy to distinguish at another.

Remote sensing instruments can also provide a valuable three-dimensional view of a landscape through the use of stereopairs, which are two remote sensing images that when viewed together add the illusion of relief to a landscape. Stereoscopic measurements can be used to make topographic maps or digital representations of topography known as digital elevation models (DEMs). Stereopairs also are stellar in their support of one of the original stated purposes of photogeology which was 'to provide better illustrations for teaching geology' (Fischer 1975: 34). Two excellent modern examples of the educational value of stereopairs and satellite images for teaching about the Earth's landforms are the *Atlas of Landforms* (Curran *et al.* 1984) and *Geomorphology from Space* (Short and Blair 1986).

The microwave region (wavelengths ranging from one mm to one m) of the electromagnetic spectrum is also important for geomorphic remote sensing. Synthetic Aperture Radars (SARs) are active remote sensing instruments that illuminate the ground with their own electromagnetic signal and then record the amount of energy that is scattered from the target back to the sending antenna. Therefore, SAR images are sometimes referred to as backscatter images. SARs offer advantages over optical sensors as they can penetrate through clouds and obtain images at night making them ideal for studying cloudy regions and for capturing short-lived dynamical geomorphic processes like flooding. SAR images capture quite different characteristics of the landscape than do optical sensors. The backscatter signal received at a SAR antenna is affected by the roughness of the surface and the moisture present in the soil and in vegetation. This makes SAR useful for assessing such important landscape properties as soil moisture, melting conditions on the surface of glaciers, the biomass of plant communities and flooding or inundation.

One unique feature of SAR that has proved valuable for geomorphic research is the ability of SAR to penetrate into dry materials, such as sand or dry snow. The longer the SAR wavelength the deeper into the subsurface the microwave energy can penetrate. A classic geomorphic study demonstrating SAR's ability to provide subsurface information was the identification of an extensive drainage network under the Selima Sand Sheet covering portions of western Egypt and eastern Sudan not easily visible from field observation or optical sensors (Hayden *et al.* 1986).

In geomorphology, remote sensing is not limited to the collection of images of terrestrial surfaces from the air or space. Ground-based remote sensing techniques including ground penetrating radar (GPR) and seismic reflection profiling provide detailed two or three-dimensional images of the near subsurface useful for studying the internal structure of landforms and glaciers and for environmental site analysis. Since the 1960s, multibeam acoustical sounding instruments, often known as side scanning SONARs (sound, navigation and ranging), have been widely used in marine geomorphology and bathymetric mapping. Remotely operated vehicles (ROVs) are providing fascinating views of otherwise unseen marine environments. Remote sensing of landscapes is not limited to Earth. By necessity, planetary geologists have made extensive use of remote sensing to study other planets and even asteroids in our solar system (Hayden *et al.* 1986).

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ANDREW KLEIN

REPOSE, ANGLE OF

The maximum angle at which a mass of debris under given conditions will remain stable. The angle of repose generally varies between 25° and 40°. For instance, the angle of repose for sand is between 30° and 35°, whereas for scree it is between 32° and 36°. The exact angle of repose depends upon slope conditions such as the size, shape, roughness and degree of interlocking, sorting, the height of fall, and density of the individual sediment grains. Also, the length of slope and the pore-water pressure of the sediment are important, as increased water content enhances structural integrity of the sediment due to surface tension between grains. A general understanding of the angle of repose is known, though studies concerning factors influencing the angle of repose have produced diverse results.

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STEVE WARD

RESIDUAL STRENGTH

Also termed ultimate strength. It refers to the minimum remaining degree of strength (i.e. resistance to movement) in a soil or rock after loss of strength following significant displacement (relative to the material but typically >1 m). The term is thus linked with slope movements, and is extremely important in slope stability analysis in order to gauge the strength of a pre-existing active slope. Residual strength in sands is typically the same as the critical shear strength (a steady state subsequent to shearing in which the effective stresses remain constant and no volume changes occur), whereas materials with high levels of clay provide a residual value of about half the critical shear strength. Soils high in platy clay materials cause a considerable reduction in strength (from peak to residual) as they tend to align themselves parallel to the direction of displacement following movement.

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STEVE WARD

REYNOLDS NUMBER

A dimensionless number used in fluid dynamics to determine the transition from laminar to turbulent flow through a pipe, developed by Osborne Reynolds in 1883. The parameter is based upon the fact that the ratio of kinetic energy to energy transferred by viscous forces was correlated with turbulent flow. The number is defined by the equation $Re = V L / v$, where Re is the Reynolds number, V is the velocity, L is the length, and v is the kinematic viscosity (viscosity/density). When this ratio is less that 1,000 laminar flow will be observed, whereas high Reynolds numbers represent turbulent flow. However, the actual definition of a high Reynolds number is determined by the shape of the system. Reynolds also studied the effects of flow resistance in pipes, demonstrating that the friction coefficient is a unique function of the Reynolds number at various surface roughnesses.

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SEE ALSO: boundary layer

STEVE WARD

RIA

A coastal inlet resulting from the drowning of a former river valley system or estuary. Rias are formations originating during the postglacial glacioeustatic transgression of the seas across the continental shelf during the Flandrian TRANSGRESSION, following the melting of the ice sheets and glaciers. This resulted in the development of an extremely irregular, indented coastline, where only the pre-existing hill peaks remained above sea level. Rias are non-glaciated, having been formed originally by subaerial erosion, and are characteristically long, narrow, often funnel-shaped inlets, whose depth and width uniformly decreases inland. They are also shorter and shallower than a fjord. The term ria originates from the type locations of Galicia and Asturias, north-west Spain, where a series of long mountainous-sided estuaries exists, once drowned by postglacial eustatic sea-level rise. Other examples include the south-west of Ireland (Kerry and Bantry Bay). A less restricted use of the term ria exists, pertaining to any broad estuarine river mouth, including fjords. However, the original application of the term is preferred in geomorphology.

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STEVE WARD

RICHTER DENUDATION SLOPE

A hillslope type, common in extreme environments such as alpine and polar regions, that develops through cliff retreat and forming a uniform (rectilinear) slope at the angle of rest of the accumulating talus. Richter denudation slopes were first noted by E. Richter in 1900, following studies in the Alps. The formation of such slopes was later expanded on by Bakker and Le Heux (1952), who modelled Richter slope formation. Richter denudation slopes develop essentially by rock fall, where the resulting talus is moved gradually by rolling and sliding, and forming a thin veneer over the basal slope. The cliff retreats steadily, often cutting across bedrock, while the basal slope may either accumulate talus, thus raising the foot of the slope, or be removed by weathering or abrasion of sliding talus. The free face will eventually be eliminated resulting in a smooth hillslope of uniform gradient. Examples of such slopes are common in the Transantarctic Mountians and Koettlitz Valley, Antarctica.

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STEVE WARD

RIDGE AND RUNNEL TOPOGRAPHY

Ridge and runnel topography comprises a series of alternating intertidal bars and troughs and is typically found on sandy beaches in fetch-limited, macrotidal coastal environments. The number of ridges (and runnels) is 3–6, the height of the ridges ranges from 0.5 to 1 m and the distance between the ridges varies from 50 to 100 m. The intertidal gradient of ridge and runnel beaches is approximately 0.015, but the seaward slope of the ridges is significantly steeper and may be up to 0.05. Storm wave conditions result in a flattening, or even destruction, of the ridge morphology. Calm wave conditions, on the other hand, induce ridge build-up and promote onshore migration of the ridges. Ridge and runnel topography is relatively stable and rates of onshore bar migration rarely exceed 1 m per tide.

It was previously thought that the ridges develop as swash bars during stationary tide conditions. However, some doubt has been cast on a swash origin of the ridges and it seems more likely that the ridges are breaker bars. Whatever its origin, ridge and runnel topography is subjected to a range of hydrodynamic processes over a tidal cycle, including swash, surf and shoaling wave processes. Depending on the wave/tide conditions and the position on the beach profile, the ridges will be affected and controlled to varying degrees by each of these hydrodynamic processes.

In the American coastal literature, welded bar systems that develop following storm erosion (see BEACH) are also sometimes referred to as ridges and runnels. This usage of the term 'ridge and runnel topography' is considered inappropriate and should be avoided.

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GERHARD MASSELINK

RIEDEL SHEAR

Refers to a conjugate set of overlapping *en echelon* faults which develops during the early stages of shearing, usually at inclinations of ~15° (R Shears or synthetic fractures) and ~75° (R' Shears or antithetic fractures) to the principal displacement zone (PDZ) boundary. Riedel shear zones form in dip-slip fault regimes and are composed of fault and fracture elements marked by standard physical properties of brittle shear zones (e.g. slickenside surfaces, slickenlines, gouge and/or breccia and abundant fracturing), though others can be observed as deformation bands and zones of deformation (Davis *et al*. 2000). Second *en echelon* synthetic and antithetic shears, termed P-shears, may form through development of the Riedel shear system, although P-shears can sometimes develop before R-shears or at the same time. Lesser shears that can develop in relation to Riedel shear include Y-shears, and T fractures. Riedel shearing was first observed by Cloos (1928) and Riedel (1929) during studies on clay-cake deformation.

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SEE ALSO: shear and shear surface

STEVE WARD

RIFT VALLEY AND RIFTING

Rift valleys are elongate depressions in the Earth's surface that are formed as a result of extension in the crust and upper mantle. Many are large enough to be easily visible from space from where they resemble large cracks that cut across continental landmasses (Plate 95). At ground level they are well defined and easily recognizable geomorphological features that have been the subject for exploration and research for more than a centrury. The term 'rift valley' was first used by Gregory in the nineteenth century to describe the East African Rift (alternatively called the Great Rift or the Gregory Rift) which runs from Afar in the north of Ethiopia to Blantyre to the south of Lake Malawi – a total length of 35,000 km. In Ethiopia this rift reaches its greatest depth of over 3 km. Other well-known rifts include the Baikal Rift of south-central Siberia, the Rhine Graben of Europe and the Rio Grande Rift of western USA.

The morphology of rifts is always similar with a central depression or rift valley flanked on both sides by uplifted areas. The uplifted shoulders of rifts are each associated with a staircase of (mostly) normal faults of varying magnitude which step the underlying rocks down towards the central depression.

The classical view of the structure of a rift was that it is symmetrical, with a flat bottom, flanked by two equally large border faults, a structure for which the German word 'graben' was coined. This interpretation was based almost entirely on an assumption that the surface expression of a rift is the same as the structure, ignoring the importance of erosion and deposition in modifying the landscape. The flat bottom observed e.g. in the East African Rift is due to deposition of lake and river sediments (Frostick and Reid 1989). Seismic evidence from rifts worldwide has

Plate 95 False colour satellite image of part of the Kenyan section of the East African Rift. Rift valley and other sediments show white

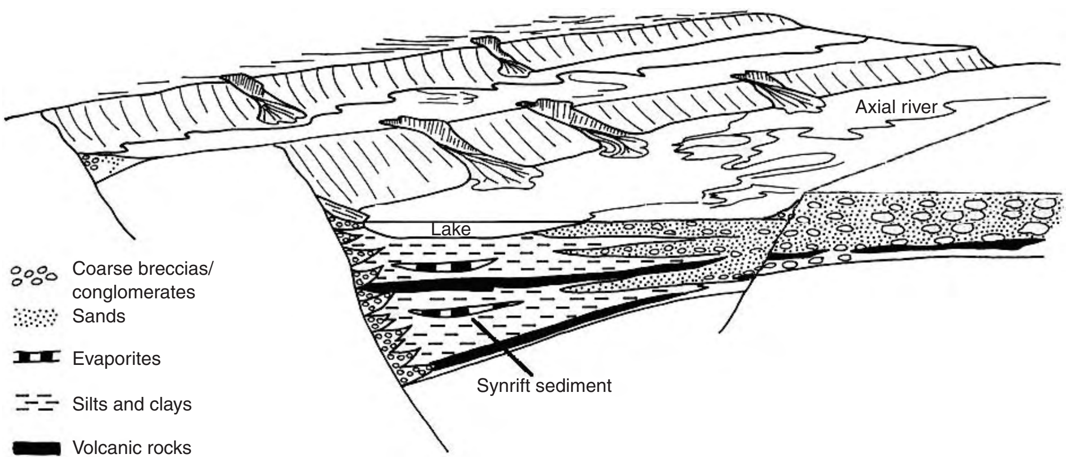

Figure 132 Schematic diagram of a cross section through rift structure. Note the asymmetry with a thicker package of sediments filling the basin close to the border fault

shown that the dominant structure is an asymmetrical half graben with one margin more intensely faulted than the other (Figure 132). The location and character of the main border faults is controlled by the structure and lines of weakness in the pre-rift rocks. In some areas there are a few large faults, with vertical displacement in excess of 2 km, and in others a plethora of smaller ones. The rocks between the faults are tilted away from the rift axis forming parallel valleys between tilted fault blocks. Another characteristic of the faulted margin is uplift, as the valley floor is displaced downwards the shoulder of the rift rises upwards to emphasize the topographic step. Rift valleys vary in width from less than 30 to over 200 km. Along the less faulted margin there are smaller faults inclined both towards and away from the rift axis (antithetic and synthetic faults). Most of these are also normal extensional faults that carve the pre-rift rocks into a series of small HORST blocks with intervening valleys.

Although continental rifts can be viewed overall as continuous elongate depressions, it is interesting to note that the underlying structure, and to some extent its topographic expression, is segmented into a series of smaller basins which vary in length from tens to hundreds of kilometres. At the divides between basins there are transverse or oblique structural elements, the nature of which is the matter of debate, named variously transfer zones, accommodation zones, relay ramps, relay zones and segment boundaries. Across these zones the basin margin occupied by the major fault can alter, giving a sinuous form to the deepest zone of the valley along rift axis.

The origin of rifts has been the subject of great debate over many decades. Rifts develop in a variety of plate tectonic (see PLATE TECTONICS) settings and can be formed anywhere the crust of the Earth is placed under tension. The most obvious circumstance in which this will occur is when a continent is splitting apart to form a new ocean (e.g. in the Red Sea–Gulf of Aden; see Girdler 1991), but it can also occur where plates are moving laterally past or even towards each other in a non-uniform way that places local areas of the crust under tension (e.g. the Dead Sea pull-apart basin). Some researchers favour classifying rifts into those associated with the constructive plate margins that lead to the development of oceans and those that are within a plate that is not splitting apart, so called intraplate settings. However, this division is difficult to justify given that continental rifts may be in an intraplate setting but still associated with oceanic development. This is the case with both the Benue trough in west Africa, which was formed during the opening of the Atlantic Ocean, and the East African Rift which is associated with the opening of the Red Sea–Gulf of Aden (Frostick 1997). Both are failed arms of oceans that had the potential to become mid-ocean ridges. Such failed oceanic rifts are called aulacogens.

The biggest rift systems in the world are on the ocean floor in the centres of mid-ocean ridges. The worldwide network of ocean ridges constitutes the most significant topographic feature on the Earth's surface, surpassing even the Himalayas in scale. A typical ridge is 1,000–2,000 km wide and 2–3 km high. The central rift is the focus of intense earthquake activity and volcanicity.

Although it is well known that rifts form as a consequence of crustal extension, the cause of the instability has been hotly debated. The cause might be convection cells in the mantle that pull apart areas of the crust or the crust might be placed under tension by other plate movements. Whatever the mechanism, the stretching of the Earth's crust to form a rift causes it to thin in a manner similar to the thinning of semi-solid toffee as it is pulled apart. Hot, low-density mantle material wells up close to the surface resulting in high heat flows in and around most rifts. The topographic expressions of hot material from lower in the Earth penetrating closer to the surface can be the development of large domes and extensive volcanic activity. The development of large uplifted domes is a feature of the early stages of ocean opening. For example, examination of the topography and drainage of the West African margin reveals a series of large domes approximately 1,000 km in diameter that predate the opening of the Atlantic Ocean (Summerfield 1991). Similar structures are associated with the East African Rift, centred on Robit in Ethiopia and Nakuru in Kenya.

Rifts are often the focus for volcanic activity which commences at an early stage of rift development and can be extensive, for example in East Africa where an area of over 500,000 km^2 is covered with rift-related volcanic rocks and many of the well-known mountains are volcanoes including Ol Doinyo Lengai, Kilimanjaro and Mount Kenya. The nature of the volcanic rocks in rifts is distinctive and contains high concentrations of so-called volatile elements (particularly carbon dioxide and halogens). Rock types include basalts, trachytes, tuffs and carbonatites. Salts leached from these rocks can contribute to the development of saline lakes, e.g. Lakes Natron and Magadi.

The new topography that results from the development of a rift valley in a continental landmass will impact on hydrology, climate and ecology in a variety of different ways. Uplift along the rift margins reduces the ambient temperature and tends to increase rainfall while the centre of the rift remains warmer and can be more arid, depending on the latitude of the rift. Rift flanks are often the sites of more lush and temperate vegetation which, in the tropics, can form rainforest. Both the topography and the contrast in habitat from flank to valley bottom act as barriers to the migration of animals and, to a lesser extent, plants. The relatively isolated environment of the rift valley bottom is one that has played a unique role in human evolution. It is now widely accepted that the hominids found in the East African Rift, largely by members of the Leakey family, show that critical stages in the evolution of hominids occured in this area prior to migration out of Africa.

The evolution of rift morphology will disrupt the pre-existing continental drainage patterns, reversing, diverting and beheading river systems in a systematic and effective way. Pre-rift, most continental drainage systems comprise a limited number of very large, long-lived rivers fed by a well-integrated network of smaller streams and draining towards the nearest ocean margin. The impact of the incipient rift will depend upon the orientation of the new structure to the existing river system. If the rift is aligned with the main drainage direction it might capture all or some of a local river system. In contrast, a rift which cuts across the pre-existing drainage often diverts and reverses sections of the drainage. Domed sections of a rift are particularly effective at drainage diversion and develop a radial stream pattern that diverts all but very local and small rivers away from the rift basin. As the structure develops further, and faults begin to carve the surface into a series of ridges, there are new adjustments. The uplift and tilting of fault blocks create new river systems which drain along the 'saddle' between adjacent fault blocks, bypassing the basin centre. Most of these bypass rivers finally gain access to the rift axis through transfer zones where the throw on the border fault reduces to zero (see e.g. the Kerio river of northern Kenya described in Frostick and Reid 1987).

In some rifts, topographic barriers pond the drainage, forming lakes. Examples are Lakes Baikal, Tanganyika and Malawi. These lakes vary in salinity from hypersaline to fresh water depending on the surrounding geology and volcanicity. In other rifts there are no lakes and axial rivers drain the length of the valley, for example in the Rhine and Benue rifts. The marginal fault scarps are cut by alluvial fans that feed water and sediment into the basinal rivers and lakes.

As the rift basin floor subsides the uplifted flanks will be progessively eroded and sediments will accumulate in the valley at a rate that largely depends on climate and hydrology. Sediments that accumulate during rifting are normally called 'synrift' sediments. The lowest areas of the valley fill first, generally with lacustrine and river sediment. In the later stages of development into an incipient ocean, sea water may penetrate into the rift valley and the whole area will become a large marine inlet with an uncertain connection to the open ocean. This can lead to the accumulation of thick salt sequences as the sea water evaporates. As the filling progresses wedge-shaped masses of synrift sediments develop and, if subsidence ceased, the valley would eventually lose its topographic identity. In the rifts we see today, subsidence is ongoing and successive wedges of sediment are superimposed on each other. Over geological time many kilometres of sediments can accumulate in rifts.

Continental rifts offer conditions favourable to the development of a number of economic deposits that are rare in other parts of the continents. Some rifts contain sediments that can produce and trap oil and gas in large quantities given the right burial history (e.g. the oil and gas of the North Sea is in a Jurassic rift). Salts that accumulate from both saline lakes and sea water are exploited in some areas for example the Dead Sea Works is situated in the Dead Sea Rift and supplies much of the world's bromine. In addition, the river sands and gravels that accumulate in these basins can be an important source of building materials if the rift is sufficiently close to a developing centre of population.

The spectacular scenery of rifts is, perhaps, their most striking feature and some rifts have therefore become attractive tourist centres. One good example of this is Death Valley in the western USA where the desert conditions reduce vegetation to a minimum and the striking geomorphology is evident even to the untrained eye.

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LYNNE FROSTICK

RILL

At the start of a rainfall event, rainwater which has fallen upon a hillslope begins to 'pond', i.e. OVERLAND FLOW moves rather slowly under the influence of gravity into small closed depressions in the soil's irregular surface (its 'microtopography'). This 'detention storage' gradually fills, although some of the stored water is constantly lost to infiltration into the soil. Meanwhile, if the rain is of moderate or high intensity then each raindrop which impacts upon an unprotected area of soil will possess sufficient kinetic energy to detach soil particles (see RAINDROP IMPACT, SPLASH AND WASH), which are thus redistributed over the soil's surface. Soil in the ponded areas is however largely protected from raindrop impacts. As a result, rainsplash redistribution usually decreases over time within a storm as the area and depth of surface water increases. There is a net downslope movement of splashed soil but this is generally small.

If the rain continues, then provided precipitation rate exceeds infiltration rate the deepening ponds on the soil's surface will eventually overtop their depressions. Overland flow which is released from overtopped ponds is likely to flow downhill more quickly and in greater quantities (i.e. possess greater kinetic energy) than the more diffuse and shallow flow into depressions: it may therefore be sufficiently competent (see SEDIMENT RATING CURVE) to transport soil particles which are splashed into it. Such soil particles can be carried some distance, being deposited only when flow velocity decreases (due to, for example, a reduction in gradient or the presence of vegetation).

Flow with still greater kinetic energy will generate a shear stress which is sufficient to detach soil particles from the body of the soil. These particles will then be transported along with splashed-in sediment. At locations where such detachment occurs, the soil's surface is lowered slightly. Such lowered areas form preferential paths for subsequent flow, and will thus be eroded further. Rather quickly, this positive feedback (see SYSTEMS IN GEOMORPHOLOGY) results in small, well-defined linear concentrations of flow (Favis-Mortlock 1998), known as 'microrills' or 'traces', with a width and depth of a few millimetres.

Many microrills will eventually become ineffective due to deposition within the microrill itself. But a fortunately located subset may grow further to become rills, with a maximum width and depth of a few tens of centimetres. This process of competition between individual channels leads to the self-organized formation (see COMPLEXITY IN GEOMORPHOLOGY) of networks of microrills and rills. Rill networks tend to be dendritic (see DRAINAGE PATTERN) in form on natural soil surfaces, but are constrained by the direction of tillage on agricultural soils. Such networks form hydraulically efficient pathways for the removal of water from hillslopes. However, sediment which is being transported within the rill network may be redeposited after a short distance if the flow loses its competence (such sediment possibly being detached again later in the same rainfall event, if flow conditions change; or during a subsequent rainfall event). Or the sediment may be carried some distance, perhaps even off the field and into a GULLY and/or a permanently flowing channel (see FIRST-ORDER STREAM). Once the rain stops, however, flow in the rill network will gradually cease: all sediment which is being transported at that time will then be redeposited within the network itself.

Rill networks on agricultural land are regularly erased by tillage, and regularly reinitiated (see SHEET EROSION, SHEET FLOW, SHEET WASH). On natural landscapes, however, rill networks persist and may in time cause such serious dissection of the hillslope as to lead to the formation of BADLANDS.

An eroding hillslope, then, normally consists of a flow-dominated channel network in which rill erosion occurs, separated by interrill areas where the dominant processes are rainsplash and diffuse flow. Soil loss from these areas is known as interrill erosion. But it is rill erosion which is the more effective agent for detachment and removal of soil, and so in many parts of the world rill erosion is the dominant subprocess of SOIL EROSION by water on hillslopes (De Ploey 1983). Boundaries between rill and interrill areas of the hillslope are frequently ill-defined and are constantly shifting. Note that subsurface flow may, in some circumstances, rival hillslope topography in importance in determining where channel erosion will begin and develop, e.g. at the base of slopes, and in areas of very deep soils such as tropical saprolites (see GROUND WATER).

Mean flow velocities within individual rills are usually between one and ten centimetres per second. Interestingly, there is evidence that for actively eroding rills, flow velocity is not dependent on rill gradient: this may be due to some compensatory increase in within-rill roughness on steeper slopes (Nearing *et al.* 1997). Velocity-depth profiles and planform patterns of velocity in rills are qualitatively similar to those of larger channels. Velocities may vary noticeably along the rill, however, with increased flow rates and scouring at 'headcuts' (Slattery and Bryan 1992), i.e. breaks of along-channel slope (such as that which is often found at the upstream extremity of each rill). Headcuts tend to move slowly upstream as their headward facets are eroded.

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SEE ALSO: erodibility; erosivity; runoff generation; sheet erosion, sheet flow, sheet wash; soil erosion; universal soil loss equation

DAVID FAVIS-MORTLOCK

RIND, WEATHERING

Weathering rinds are zones of chemical alteration on the outer portions of rocks. In some, but not all cases, a distinct colour difference highlights this zone of intense CHEMICAL WEATHERING. Weathering rinds are important in geomorphology for their role in weathering processes, their role in the development weathering forms such as CASE HARDENING, and in their use in dating landforms. Obsidian-hydration rinds are a related phenomenon.

A weathering rind is not just a zone of chemical alteration at the outer edge of a clast; weathering rinds represent the redistribution of elements. Some rinds are dominated by an enrichment in iron, while others are depleted in such mobile cations as calcium and sodium. A variety of processes develop weathering rinds. Dissolution, for example, leaves void space in the rock and does not necessarily change the colour. Oxidation of iron, in contrast, leaves a band of discolouration. The appearance of the zone of discolouration varies by location and rock type. For instance, rinds can appear white on the upper slopes of Mauna Kea and appear orange on the lower slopes, all in a basalt lithology (Plate 96). Andesite in Japan can appear brown to pale grey (Matsukura *et al.* 1994), and sandstone

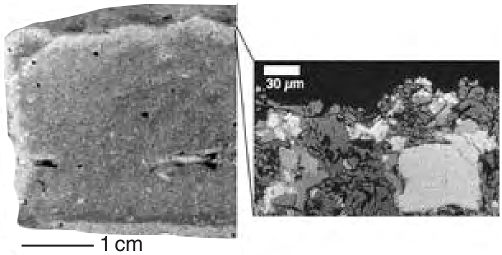

Plate 96 Weathering rind developed on a glacially polished basalt, Mauna Kea, Hawaii. This rind developed over a 16,000-year period. The left photograph shows an optical rind visible in a hand specimen. The right image shows an electron microscope (backscatter) image of a small section of the rind, illustrating three aspects of rind development. First, dissolution of minerals dominates rind formation, as exemplified by the pores (black areas). Second, the bright spots in the image are reprecipitated iron hydroxides, responsible for reddening. Third, rinds may not necessarily thicken over time. Often, they undergo erosion as pieces of weathered minerals progressively detach, that is if rinds are not protected by rock coatings

rinds in New Zealand can appear whitish (Knuepfer 1988).

Weathering rinds form on all three rock types: igneous (e.g. andesites, basalts, granitic), sedimentary (e.g. sandstones) and metamorphic (e.g. schists). Weathering rinds occur in a wide range of locations and in temperate, tropical, arctic and arid environments, for example, Hawaii (Jackson and Keller 1970), the coterminous United States (Colman and Pierce 1986), New Zealand (Chinn 1981), Japan (Matsukura *et al.* 1994) and northern Europe (Dixon *et al.* 2002). Weathering rinds are found in clasts at the surface and within the soil profile (Chinn 1981; Knuepfer 1988).

Weathering rinds are often used in geomorphology to estimate ages of landforms and landscape surfaces (Chinn 1981). This approach assumes that rinds begin to form soon after emplacement of the host rock, and that rinds grow thicker with time (Knuepfer 1988). Weathering rinds thus serve as a relative age indicator where thicker rinds occur on older landforms, and as a calibrated age indicator if accurate forms of age calibration are available in the study area. Prior to the use of cosmogenic nuclides (see COSMOGENIC DATING), use of weathering rinds was prevalent in Quaternary research where moraines, outwash sheets and other landforms correlated climatic changes (Colman and Pierce 1986). The thickness of the discoloured zone of a number of clasts in a deposit is measured normal to the surface, usually with a caliper. Statistical methods differentiate groups of thicknesses among different deposits or surfaces.

Because weathering rinds are so often felt to be synonymous with discolouration, we stress that the study of weathering rinds should not be limited to the measurement of colour changes in hand samples for several reasons. First, a weathering rind can occur without any noticeable colour change. Second, colour change provides only one indication of weathering; microscope studies reveal that the zone of chemical weathering continues into the rock well underneath the zone of colour change. Third, although weathering rinds are not ROCK COATINGS, a single clast may exhibit both a weathering rind and a rock coating (Matsukura *et al.* 1994), a distinction not always recognized in the field. Fourth, where weathering rinds are not protected by rock coatings, weathered mineral fragments readily spall off.

Research into weathering rinds is expanding into exciting new dimensions. Physical and chemical characteristics of weathering rinds are being used to help discern geochemical weathering processes in a given region or area (Dixon *et al.* 2002). The use of cosmogenic nuclides as a dating method has made weathering rind analysis more important than ever. A key uncertainty in cosmogenic dating surrounds the prior exposure history of a possible sample. With each cosmogenic measurement costing about US$2,000 in sample processing and analysis, weathering-rind measurements provide an inexpensive field check on the possibility that a particular sample might have a complex geomorphic history. In addition, *in situ* measurements of weathered minerals in rinds are providing new insight into quantitative rates of weathering; this method is being used, for example, to establish long-term rates of glass dissolution with the goal of understanding GEOMORPHOLOGICAL HAZARDS associated with nuclear waste storage (Gordon and Brady 2002).

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SEE ALSO: case hardening; chemical weathering; rock coating

STEVEN J. GORDON AND RONALD I. DORN

RING COMPLEX OR STRUCTURE

> A petrologically variable but structurally distinctive group of hypabyssal or subvolcanic igneous intrusions that include ring dykes, partial ring dykes and cone sheets. Outcrop patterns are arcuate, annular, polygonal and elliptical with varying diameters ranging from less than 1 to 30 km or greater. The majority of ring complexes represent the eroded roots of volcanoes and their calderas.
>
> (Bowden 1985: 17)

Ring dykes are thick, approximately vertical igneous bodies that form concentric circles around a central intrusion. They are associated with a process called cauldron subsidence. Cone sheets tend to be thinner and have a general form as a set of inverted cones. They result from stresses set up in the Earth's crust as the magma body with which they are associated forced its way upwards. Other circular structures are associated with impact events.

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A.S. GOUDIE

RIP CURRENT

Many of the world's BEACHes are characterized by the presence of strong, concentrated seaward flows called rip currents. The term was introduced by Shepard (1936) to distinguish rips from the misnomers 'rip tide' and 'undertow', which are unfortunately often still used to describe rips today. Rips are an integral component of nearshore cell circulation and ideally consist of two converging longshore feeder currents which meet and turn seawards into a narrow, fast-flowing rip-neck that extends through the surf zone, decelerating and expanding into a rip-head past the line of breaking WAVEs. The circulation cell is completed by net onshore flow due to wave mass transport between adjacent rip systems (Figure 133a). Rip flows are often contained within distinct topographic channels between bars (see BAR, COASTAL) and are a major mechanism for the seaward transport of water, sediments and pollutants (Figure 133b). Rips are also a major hazard to swimmers and it is of concern that many aspects of rip occurrence, generation and behaviour remain poorly understood.

Rip currents are generally absent on pure dissipative and reflective beaches, but are a key component of sandy intermediate beach states in microtidal environments. Short (1985) identified three types: (1) accretion rips occur during decreasing or stable wave energy conditions and are often topographically arrested in position with mean velocities typically on the order of $0.5-1\,\mathrm{m\,s^{-1}}$; (2) erosion rips are hydrodynamically controlled and occur under rising wave energy conditions. They are transient in location, having mean flows in excess of $1\,\mathrm{m\,s^{-1}}$; and (3) mega-rips, which occur in embayments under extremely high waves, and can extend more than 1 km offshore with mean velocities greater than 2 $\mathrm{m\,s^{-1}}$. All are associated with localized erosion of the shoreline and often create rhythmic embayments termed mega-cusps. Relatively permanent rips located adjacent to headlands, reefs and coastal structures, such as GROYNEs, are referred to as topographically controlled rips.

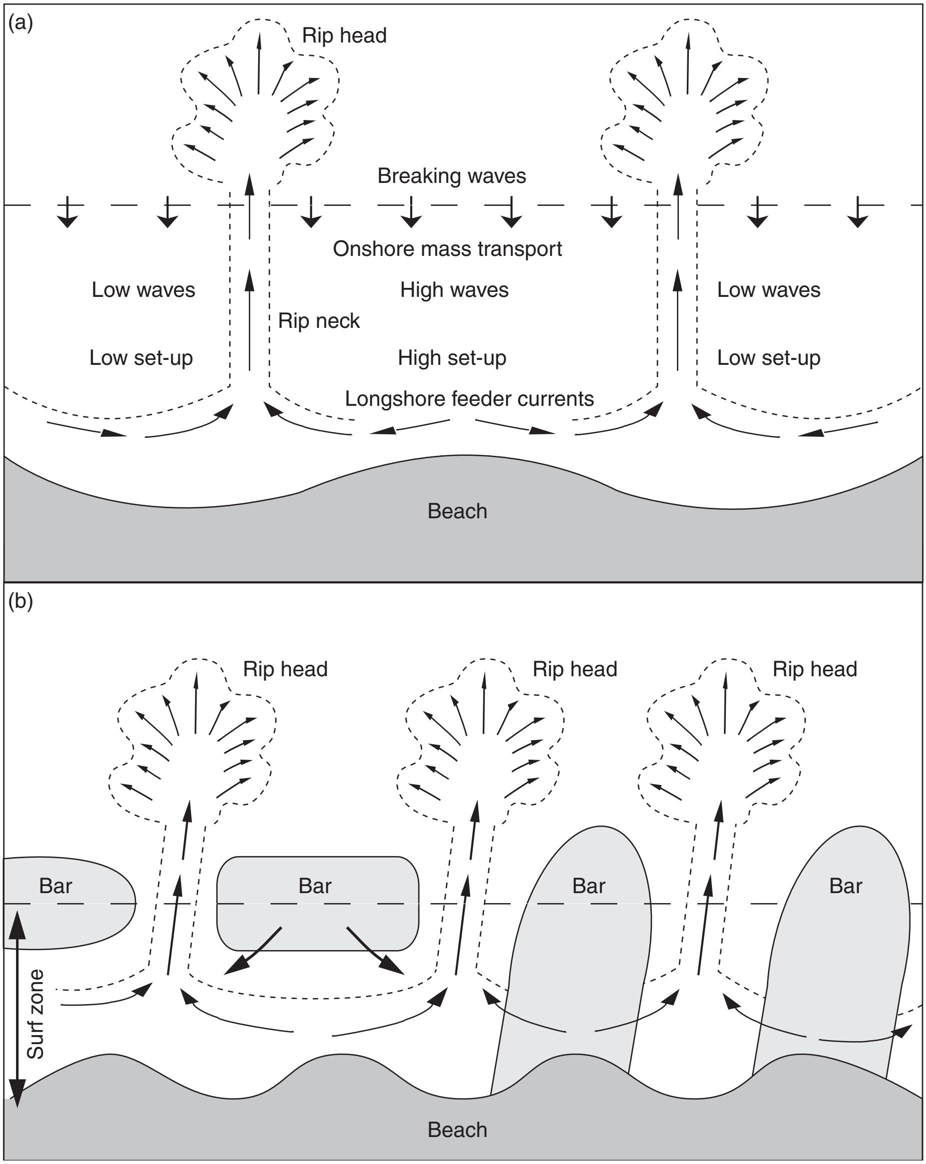

Figure 133 Idealized patterns of rip current flow and components in relation to: (a) nearshore cell circulation and wave set-up gradients; and (b) coastal bar topography

The primary limitation to our understanding of rips has been the difficulty obtaining quantitative field measurements from an energetic environment. Early attempts at describing rips (e.g. McKenzie 1958) were largely qualitative, but correctly identified that rips often display a periodic longshore spacing, increase in intensity and decrease in number as wave height increases, and flow fastest at low tide. Subsequent theoretical, laboratory and field studies have attempted to explain these characteristics with varying degrees of success, although it is generally accepted that rips exist as a response to an excess of water, termed wave set-up, built up on shore by breaking waves. The flow is forced by longshore variations in wave height, which produce gradients in the set-up that drive water alongshore from regions of high to low waves (Bowen 1969; see Figure 133a).

Existing models for the generation of rip cell circulation have thus incorporated various mechanisms to account for the existence of these longshore gradients and can be grouped into three main categories: (1) the wave–boundary interaction model involves wave modification by non-uniform topography and/or coastal structures. For example, wave refraction can produce regions of high and low waves, such that rips can occur in the lee of offshore submarine canyons (Shepard and Inman 1950), but more commonly adjacent to headlands and groynes; (2) wave–wave interaction models have shown theoretically and in laboratory experiments (Bowen and Inman 1969) that incident waves can generate synchronous edge waves that produce alternating patterns of high and low wave heights along the shoreline. Rips occur at every other antinode with a spacing equal to the edge wave length; and (3) instability models suggest that longshore uniformity in set-up is unstable to any small disturbance caused by hydrodynamic or topographic factors and rip spacing is predicted to equal four times the surf zone width. It should be emphasized that validation of these models has primarily been restricted to laboratory experiments and has not been adequately verified in the field. Short and Brander (1999) used a global field dataset to show that rip spacing is related to regional wave energy environments. Patterns of rip spacing (L_r) were consistent within west coast swell ($L_r \cong 500$ m), east coast swell ($L_r \cong 200$ m), and fetch-limited wind wave environments ($L_r \cong 50-100$ m).

The wave–boundary model is best supported by Sonu (1972) who found that on a beach consisting of alternating sandbars and topographic channels with uniform longshore wave height, constant and extensive wave energy dissipation across the bars and local and intense wave breaking over the channels created a set-up gradient towards the channels, which controlled rip flow. Set-up gradients generated in this manner support field data confirmation (e.g. Brander 1999) that rip flows are tidally modulated, since stronger flows at low tide would be expected with increased wave dissipation associated with shallower water depths over the bars. Field studies have also shown that rip velocities increase steadily from the feeders, attaining maximums in the middle of the rip-neck, are greater near the water surface and experience short duration and strong velocity pulses every few minutes, the forcing of which is likely related to infragravity motions such as shear waves or wave groups.

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SEE ALSO: bar, coastal; beach; beach sediment transport; groyne; wave

ROBERT W. BRANDER

RIPARIAN GEOMORPHOLOGY

Riparian geomorphology is concerned with the dynamics, form and sedimentary structure of riparian zones. Riparian zones have been variously described as 'three-dimensional zones of direct interaction between terrestrial and aquatic ecosystems' (Gregory *et al*. 1991: 540); zones that extend 'from recently colonized fluvial landforms exposed at low flow to the limits of the area wherein biota are adapted to, or characteristic community structures are influenced by, flooding' (Dykaar and Wigington 2000: 88); and the 'part of the biosphere supported by, and including, recent fluvial landforms . . . inundated or saturated by the BANKFULL DISCHARGE' (Hupp and Osterkamp 1996: 280). From these and other definitions, it is apparent that the riparian geomorphology of a river reach is dependent upon: present and past flow magnitude and frequency (see MAGNITUDE–FREQUENCY CONCEPT); the amount and calibre of sediment transported by the river; and the slope and degree of confinement of the reach. Whilst the past and present flow and sediment transport regime govern the materials delivered to the reach for landform building, the local slope and confinement of the reach govern the river's energy and its ability to construct and erode landforms.

Nanson and Croke (1992) explored these controls to develop a genetic classification of FLOODPLAIN types that is explicitly linked to the river types that construct the floodplains. They defined three broad groups of floodplain (high-energy non-cohesive, medium-energy non-cohesive, low-energy cohesive) based upon the river's ability to do work and expressed by its specific STREAM POWER at bankfull discharge, and the erosional resistance of the floodplain materials (non-cohesive implies gravel or sand; cohesive implies silt and clay). They subdivided the three groups into thirteen different floodplain classes. These were discriminated by details of the sediment from which they are constructed, the river planform or pattern, its characteristic erosional and depositional processes, and thus the typical landforms present on the floodplain and within the river margins. Importantly, this classification links process and form in a dynamic way, illustrating that riparian zones may possess an enormous variety of landforms and that the nature and dynamism of the landforms varies between floodplain types. Thus, if there is a change in the controlling processes, riparian geomorphology also changes. For example, changes in climate, flow regulation and flood defence engineering affect river flow and sediment transport regimes and the ERODIBILITY of channel margin materials, and so can have far-reaching impacts on riparian zone character (e.g. Steiger and Gurnell 2002).

In most analyses of riparian zone form and process, vegetation has been seen to play a largely passive role, responding to present and past environmental conditions created by fluvial processes (e.g. Hupp 1988). Thus, floodplain vegetation patterns have been interpreted to depend on the type and age of the mosaic of riparian landforms. Migrating, MEANDERING rivers provide a simple illustration. As the river erodes the outer banks of meander bends, POINT BARS develop on the inner banks. Vegetation colonizes point bar surfaces and plant species are gradually replaced as sediment, moisture, light and disturbance on the bars change during their aggradation and incorporation into the floodplain.

Recently, more emphasis has been placed on the active role of vegetation in influencing riparian zone geomorphology. For example, Gurnell and Petts (2002) consider both biotic and abiotic ways in which vegetation can influence the form, sedimentary structure and dynamics of riparian zones. Abiotic influences include the impact of root systems on the erodibility of sediments and the flow resistance of the vegetation canopy. Roots can cause significant reinforcement of riparian sediments, making them more resistant to river erosion. When the riparian zone is flooded, the ROUGHNESS of the vegetation canopy can reduce flow velocities across the vegetated surface, reducing rates of erosion and increasing rates of sedimentation. These abiotic processes can significantly affect patterns of erosion and aggradation, and thus the form and sedimentary structure of riparian zones. The geomorphological significance of these abiotic influences depends on the species, age and density of the vegetation cover, which is related to several biotic processes. The degree to which riparian plants reproduce from seeds or by vegetative reproduction is important because, in general, riparian vegetation growth is more rapid when plants propagate vegetatively. The timing of seed or vegetative propagule release can greatly influence the likelihood of successful vegetation establishment, because many riparian plant propagules are transported and deposited by the river. For example, the timing of propagule release in relation to the climate and river flow regimes can influence whether suitable colonization sites are exposed or inundated by the river, and whether their moisture and temperature characteristics are appropriate to support the successful germination and growth of young plants.

Riparian tree species can be particularly important riparian zone engineers. Poplar and willow species can grow very rapidly, propagating through both seeds and vegetative reproduction. Rivers may erode, transport and deposit whole trees as well as fragments (branches, twigs, root boles) and seeds. Entire trees may survive rafting by floods, deposition and burial within river margins and on bars, and they can sprout to form patches of new sizeable shrubs within a year. The importance of these processes for riparian geomorphology varies with tree species and environmental conditions but also with riparian tree management. The pruning and felling of riparian trees to prevent LARGE WOODY DEBRIS entering rivers is often carried out to maintain the FLOOD conveyance of the river channel. Its impact on riparian geomorphology and ecology is far reaching, leaving little impression of the diverse geomorphological and ecological character and high dynamism of unimpacted riparian zones (Gurnell *et al.* 1995, 2002).

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ANGELA GURNELL

RIPPLE

Ripple is a general term applied to a range of normally unrelated, very small bedforms that occur in trains and record sediment mobilization and transport in various aqueous and aeolian environments (see BEDFORM; BEDLOAD; ROUGHNESS). The main kinds are current and rhomboid ripples, oscillation or 'symmetrical' ripples (see WAVE), ballistic or impact ripples (see AEOLIAN PROCESSES; SALTATION), adhesion ripples and warts (see AEOLIAN PROCESSES; SALTATION), and rain-impact ripples (see RAINDROP IMPACT, SPLASH AND WASH). With the exception of adhesion warts, and some complex oscillation types, ripples are characterized by crests that lie transversely to flow.

Trains of current ripples, restricted to the coarser silts and the finer sands, are typical of rivers, but also appear in tidal environments (estuaries, barred beaches) where flows can be unidirectional for several hours at a time. As equilibrium bedforms, current ripples have linguoid crests in plan, heights of up to about 0.02 m, wavelengths of 0.1–0.2 m, and strongly asymmetrical profiles, the short leeward face lying at the angle of repose (see REPOSE, ANGLE OF). When generated from a smooth bed, however, current ripples evolve toward a linguoid shape through a range of long-crested forms, the crests of which increasingly lose straightness. Internally, current ripples are cross-laminated, commonly in climbing sets, a testimony to high rates of sediment deposition on a scale of minutes or hours. Ripple dimensions are independent of flow depth but increase weakly with grain size. Diamond-shaped rhomboid ripples are developed where ripple-generating flows are sufficiently shallow as to be supercritical.

Other flows being absent, wind waves generate within the affected water-body symmetrical, oscillatory currents which are superimposed on a much weaker drift in the direction of wave-propagation. When sufficiently powerful, their combined effect on sand beds is to create trains of ripples with long, regular crests and steep, almost symmetrical, trochoidal profiles which, as revealed by internal cross-laminae, migrate very slowly in the direction of wave-propagation. Ripple scale depends in a complex manner on the properties of the waves and the sediment, the wavelength and height increasing markedly with grain size. Wavelengths are of the order of 0.01 m in silt, 0.1 m in fine sand and 1 m in coarse sands and fine gravels. Broadly, wavelength is about 500 times the median grain diameter. Wave ripples are most familiar from estuaries and beaches but, after storms, appear on continental shelves to water depths of 100–200 m. Complex forms of ripple occur where barriers reflect waves and where, especially on beaches and in estuaries, unrelated unidirectional and wave currents operate either simultaneously or sequentially. Wave ripples are valuable indicators of shallow water and of shoreline location and orientation.

The saltation of wind-driven grains over a dry bed is generally accompanied by the development of trains of ballistic ripples, resulting from an unstable interaction between the surface and the flow of sediment. These ripples are rather flat, asymmetrical structures which vary in form and scale with increasing grain size and the average length of the jumps made by the particles. Typically, ripples in the finer sands have crests that are long and regular in plan and wavelengths of about 0.05 m. Those in sediments of very coarse sand or granule grade take wavelengths of the order of 1 m and generally have short, irregular crests, along which the coarser particles conspicuously lie. Ballistic ripples are cross-laminated internally, but the structure is difficult to see in the well-sorted sands of which the smaller examples are formed. The ripples have long been reported from deserts and sandy coasts, wherever the wind is free to mobilize sufficiently coarse grains.

The capture of saltating particles by a damp or wet surface, such as a coastal sand beach, river bar or sabkha, gives rise to upwind-facing, centimetre-scale adhesion ripples (uniform wind-direction) or adhesion warts (wind-direction variable). These common and widespread structures have no particular climatic significance but are valuable proofs of surface exposure and aeolian activity. Advancing in the opposite direction to the wind, adhesion ripples create a steep internal bedding that dips downwind.

Rain-impact ripples are centimetre-scale, upwind-facing, transverse ridges shaped when heavy rain driven by a strong wind descends at a fine angle onto an exposed, water-saturated sand bed, such as a beach, tidal sand shoal or river bar. The ridges advance very slowly in the direction of the wind under the repeated impact of the drops. If rain-impact ripples have a fossil record, which is uncertain, they would afford a further proof of atmospheric exposure.

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J.R.L. ALLEN

RIVER CAPTURE

River capture, sometimes called stream capture or stream piracy, refers to the occurrence of the seizure of the waters of a stream or drainage system by a neighbouring one. It is based on the difference in local BASE LEVEL heights, with the captured stream having a higher base level and for that reason with a low erosion potential. The predatory stream, with a lower base level, is capable of diverting in its favour the waters of the less active stream, and in this way enlarging its drainage net and catchment area. Integration of both drainage systems leads to a higher order network. It does not only occur because of a steeper gradient but also because the pirate stream is cutting its valley in softer rock.

Capture constitutes a common event in the erosional evolution of a drainage net of a region and is a traditional concept in geomorphology that can be found in classical authors. Gilbert (1877) described the process in relation to the role of unconsolidated materials in mine dumps, calling it abstraction, a term often applied to the simplest type of capture, which results from competition between adjacent consequent gullies and ravines. He was also aware that a stream flowing down the steeper slope of an asymmetrical ridge erodes its valley more rapidly than the one flowing down a more gentle slope, and as a result the divide migrates away from the more actively eroding stream. This principle has been refered to as the *law of unequal slopes* (Thornbury 1969). The same concept was integrated by Davis (1899) in his model of relief evolution by the geographical cycle, capture taking place in the young or early mature stages of development. Another classical author, Horton (1945), in his slope runoff model also takes into consideration the capture process and uses it in order to explain the development of a hierarchical drainage net, that is, the process by which the drainage lines become integrated into a few dominant stream courses. Unequal rainfall on two sides of a divide may contribute to divide migration, especially where winds are prevailing from one direction, as in the trade wind belts (Thornbury 1969).

At the point at which the capture takes place, the captured stream bends sharply, forming a right angle turn into the pirate stream, which is called the elbow of capture. The valley stretch in which the captured stream continues to flow after losing the upper part of its catchment becomes a beheaded valley. This valley is then too large for the stream that continues to flow in it and thus becomes an UNDERFIT STREAM, that is a stream too small to be hydrologically related to the valley in which it now flows. On the other hand, the captured part of the stream now has a lower local base level, which increases its erosion potential and makes it able to incise into its former alluvial valley floor, producing a terrace in its former floodplain.

The capture process mainly occurs in two different ways: by headward erosion and by lateral erosion. *Headward erosion* is the probable cause of most easily recognizable stream captures. It takes place when the tributaries of the high energy stream are working back towards its head, and eventually reach the neighbouring valley head and cut through the divide. Capture by *lateral erosion* occurs when two streams flow parallel at no great distance from each other. Progressive erosion produces a lateral shifting of the stream which can finally produce a planation of the water divide. If this continues at the cost of one of the neighbouring streams, it ends in the lateral capture of its waters. Capture by subterranean waters can also occur in soluble rocks when water from a stream at a higher level percolates and meets an underground stream flowing at a lower level.

Examples of river captures have been described in many regions of the world, both at large and at small scales. In the large scale, one of the classical examples is that of a tributary of the Indus captured by the Ganges which implied the transfer of the drainage of a large area of the Himalayas from Pakistan to India. In Yunnan Province, China, rivers flowing towards the Red river were captured by the middle Yangtze tributaries. In Queensland, eastern Australia, the Fitzroy River has reached an old divide at the Connors Range and captured several of the rivers flowing in this area. In New Zealand the capture of the Silver Stream by the Karori near Wellington is well known. In Europe waters were diverted from the Danube towards the Rhine by a small head-ward tributary. In North America, in the Appalachian region of the eastern United States, there are many captures which are controlled by differences in rock hardness.

Amongst the implications of river captures is their role and significance for the evolution of relief and the history of drainage patterns, providing an interesting geomorphic challenge. In that sense, a highly integrated stream system with a large main stream is usually an indication of a long period of development (Ahnert 1998). Another important issue mentioned by Schumm (1977) is the role of captures in the discovery of new placer deposits, which are alluvial deposits containing valuable minerals, because the regional distribution of placers can be strongly influenced by stream capture. The result of this process from an economic point of view is that the source of the valuable minerals may be abruptly isolated from the downstream depositional area.

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SEE ALSO: base level; gully; underfit stream

MARIA SALA

RIVER CONTINUUM

The biological concept of a river continuum describes a regular downstream progression of such physical variables as channel width, diel temperature pulse and stream order, in relation to biotic adjustments (Vannote *et al*. 1980). The concept was originally developed for rivers in regions with deciduous forests. In these regions, headwater streams (orders 1–3) are narrow and shaded by riparian vegetation. The vegetation reduces instream or autotrophic production from algae by shading, and contributes large amounts of coarse organic detritus (>1 mm diameter), such as leaf litter. The ratio of photosynthesis/respiration (P/R) is less than 1. The diversity of soluble organic compounds is high, and the diel temperature pulse is low. Communities of aquatic insects in headwater streams are dominated by insects that shred coarse organic matter (shredders), and insects that filter finer organic matter from transport, or gather such material from sediments (collectors). Fish populations have cool-water species that feed mainly on invertebrates. Biotic diversity is low.

Medium-sized streams (orders 4–6) are sufficiently wide that sunlight reaches a greater portion of the stream channel. Algae and rooted plants in the stream are more plentiful, and the ratio of P/R exceeds 1. The diversity of soluble organic compounds drops sharply relative to headwater streams, and the diel temperature pulse reaches a maximum. Fine particulate organic matter (50 μm–1 mm diameter) becomes more important. Collectors remain important, shredders form a smaller percentage of insect communities, and grazers that shear attached algae from surfaces in the stream increase in abundance. Fish populations now have more warm-water species that feed on invertebrates and other fish. Biotic diversity reaches a maximum.

Large rivers (>6 order) are very broad and open to sunlight, but photosynthesis may be limited by depth and turbidity. Large quantities of fine particulate organic matter from processing of dead leaves and woody debris come from upstream, and the ratio of P/R again drops below 1. The diel temperature pulse is low. Aquatic insects are primarily collectors. Fish are warm-water species that feed on plankton, invertebrates and other fish. Biotic diversity drops off again.

The general pattern described above may differ in mountainous areas where headwater streams flow through alpine meadows, in dry regions where riparian vegetation is restricted, or along deeply incised channels where shading from valley walls limits photosynthesis. However, the river continuum does provide a conceptual model of spatial gradients in physical and biological variables. This conceptual model is one of the first holistic theories of a river as an ecosystem, rather than individual segments. The river continuum emphasizes the connections between the river and its terrestrial setting. The continuum also suggests that aquatic communities can be explained by the mean state of environmental variables and their degree of temporal variability and spatial heterogeneity (Minshall *et al*. 1985). Together with hypotheses of stream succession that predict changes in habitat and species following a disturbance such as a FLOOD, the river continuum concept facilitates predictions of reach-specific patterns in habitat, communities or life-history strategies.

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SEE ALSO: fluvial geomorphology; large woody debris; stream ordering

ELLEN E. WOHL

RIVER DELTA

River deltas are coastal accumulations of terrestrial sediments that rivers have brought to the sea. The Greek historian Herodotus (*c*.450 BC) originally applied the term 'delta' to the triangular subaerial deposit surrounding the mouth of the Nile River. In modern usage, however, deltas can be either subaerial or subaqueous accumulations and may have a variety of geometries. Although deep sea fans may also be considered deltas, they are not discussed here. Here, the 'subaqueous delta' is assumed to be confined to deposits on the continental shelf. The prevailing shape of any given delta depends on the rates of sediment supply by the rivers and the patterns and rates of sediment dispersal by coastal ocean processes and by gravity. In many cases, the subaqueous deltaic deposits are much more extensive than the subaerial deposits and, in some cases such as that of Papua New Guinea's Sepik River which discharges directly into deep water, the subaerial delta may be missing altogether. Historically, deltas have played important socio-economic roles. Subaerial deltas were the sites of early agriculture and formative civilizations and presently support some of the world's largest urban centres (e.g. Shanghai, Bangkok, Cairo). Subaqueous deltas are sinks for terrestrial carbon and are sources of fossil fuel.

Deltas vary immensely in both area and volume. The size of a delta depends at the lowest order on the annual sediment discharge of the river but the most extensive deltas also tend to be developed where wide, low gradient continental shelves provide a platform for prolonged sediment accumulation and morphological progradation. Hence, the largest deltas are found on passive (as opposed to active) continental margins (Wright 1985). Despite this fact, active margins are probably equally or more important than passive margins in supplying river sediment to the sea; Milliman and Syvitski (1992) showed that the numerous small mountainous streams, particularly those of the humid tropics, are collectively the most important source of terrestrial sediment to the sea. However, since these rivers are spatially distributed and since much of this sediment is bypassed to deep water, large deltas typically do not result. Other factors that influence delta area and the relative sizes of subaerial vs subaqueous components include tectonic subsidence and the energy of waves and currents that resuspend or prevent shallow water deposition of sediments. Table 38 lists characteristics of five deltas.

Satellite images of the Changjiang (Yangtze) and Mississippi Deltas (Plates 97 and 98) illustrate the diversity of major deltas. In the case of the Changjiang (Plate 97), the river-supplied sediments have built a large lobate protrusion into the East China Sea that supports and surrounds Shanghai. More exaggerated seaward protrusion of this delta is constrained by strong currents and waves, which disperse newly discharged sediments over the shelf and into Hangzhou Bay. These sediments are accumulating on the shelf at a rate of $5\,\mathrm{cm\,yr^{-1}}$ (DeMaster *et al.* 1985) to form the subaqueous component of the delta. The Mississippi Delta, one of the world's most extensively studied deltas, is composed of sediments from a catchment that covers 60 per cent of the continental United States. Its elongated and narrow digitate shape is rare and is attributable to a combination of fine cohesive sediment, low wave

Table 38 Characteristics of five major river deltas

Property	River delta				
	Amazon	Ganges–Brahmaputra	Changjiang (Yangtze)	Huanghe (Yellow R.)	Mississippi
Drainage Basin Area ($\mathrm{km^2} \times 10^6$)	6.15	1.48	1.94	0.77	3.27
Water discharge ($\mathrm{km^3\,yr^{-1}}$)	6,300	971	900	42	580
Sediment discharge (10^6 tonnes $\mathrm{yr^{-1}}$)	900	1,620	480	1,060	210
Sediment/water ratio ($\mathrm{kg\,m^{-3}}$)	0.14	1.67	0.53	25.25	0.36
RMS wave height (m)	1.6	2.5	1.5	2.0	1.1
Spring tide range (m)	5.8	3.6	3.0	1.4	0.4
Total delta area ($\mathrm{km^2} \times 10^3$)	467	106	67	36	29
Ratio subaerial/subaqueous area	6.4	2.4	1.7	3.3	5.3

Sources: Coleman and Wright (1975); Milliman and Meade (1983); Wright and Nittrouer (1995)

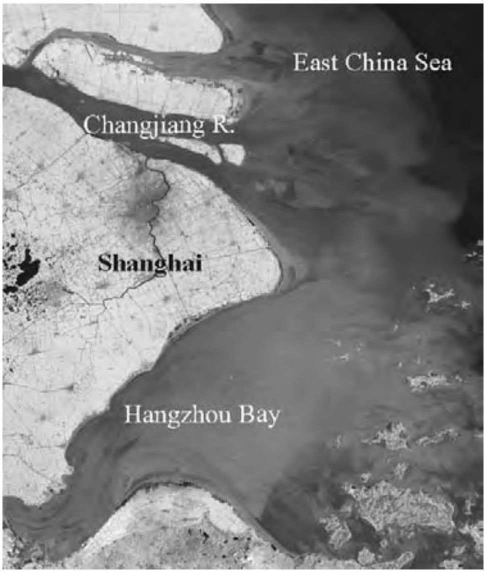

Plate 97 Satellite image of the Changjiang (Yangtze) delta and estuary showing the city of Shanghai, the turbid river effluent and Hangzhou Bay which serves as a sink for much of the sediment

Plate 98 Satellite image of the Mississippi Delta showing the active 'bird's foot' in the right portion of the image and the abandoned La Fourche delta lobe on the left

height and negligible tidal currents, a regime that allows sediments to accumulate near the river mouths without being widely dispersed by oceanographic forces (Wright and Nittrouer 1995). The modern Mississippi 'bird's foot' (Plate 98) now extends across the continental shelf creating a barrier to east–west currents. In recent geologic history, a series of such lobate overextensions have been followed by avulsions: at least sixteen such lobes have been created and abandoned in Holocene time (Kolb and Van Lopik 1966). Abandoned deltaic lobes make up most of Louisiana's coastal plain, which is experiencing a high rate of coastal land loss because of regional subsidence and erosion.

The processes that disperse, transport and deposit the sediment discharged by a river determine the configuration of the resulting delta. This is true not only for the subaqueous component but also for the subaerial delta, which must surmount the subaqueous deposits in order to prograde. Wright and Nittrouer (1995) argued that the fate of sediment seaward of river mouths involves at least four stages: (1) supply via river plumes; (2) initial deposition; (3) resuspension and onward transport by marine forces (e.g. waves and currents); and (4) long-term net accumulation. Different suites of processes dominate each stage. Immediately upon leaving the confines of a river mouth, a river effluent spreads as either a positively buoyant (lighter than seawater because of the salinity difference) or negatively buoyant (because of very high suspended sediment concentrations in the river water) plume while mixing and exchanging momentum with the ambient seawater. This is the first stage of dispersal. Most of the larger rivers that drain to passive continental margins have positively buoyant effluents because they carry low concentrations of suspended sediment and large volumes of low-density fresh water. Examples include the Mississippi, Amazon, Ganges–Brahmaputra and Changjiang among numerous others. The most prominent exception among the large rivers is the Huanghe (Table 38), which often transports suspended sediment in concentrations greater than $25\,\mathrm{kg\,m^{-3}}$ creating an effluent bulk density greater than that of seawater (Wright *et al.* 1990). Such negatively buoyant effluents are referred to as hyperpycnal (excessively dense) and tend to move downslope within the near-bed layer under the influence of gravity. Hyperpycnal conditions are somewhat more common at the mouths of smaller rivers that drain mountainous catchments near the coasts of active margins. The Eel River of northern California is a prominent example (Geyer *et al.* 2000).

The second stage of sediment dispersal is represented by initial, but usually temporary, deposition from the expanding and decelerating effluents of stage 1. River-mouth bars of varying geometries are among the morphologic products of this deposition. The deposition is caused by sediment flux convergence produced by effluent deceleration and sediment particle settling. The more rapidly the effluent gives up its momentum through mixing and bed friction and the greater the particle setting velocity, the closer the one would expect initial deposition to be to the river mouth. On the other hand, high waves and strong coastal currents enhance mixing and effluent momentum exchange with the sea but may also act to resuspend sediment or to inhibit initial deposition. Along high-energy coasts, the initial deposition may be delayed until the sediment reaches a deeper, more quiescent environment such as the mid-shelf region (Ogston *et al.* 2000). Energetic oceanographic processes also disperse sediment parallel to the coast or isobaths, generally preventing the formation of digitate or protruding deposits.

The third, resuspension and transport, stage of dispersal may act concurrent with or subsequent to the initial deposition stage (stage 2). When the coastal regime is highly energetic throughout the year or when high energy coincides with high river discharge, deposition is delayed as explained above. However, in many cases (e.g. Huanghe; Wright *et al.* 1990), the maximum input of river sediments to the sea and the maximum agitation of the bed by waves occur in different seasons. In such cases, the initial deposition may take place near the river but be removed, partially or wholly, by wave-induced resuspension a few months later. Depending on the amount of time that elapses between initial deposition and eventual resuspension, sediments may undergo varying degrees of consolidation making them more resistant to erosion and more likely to remain at the initial deposition site.

In the fourth dispersal stage, the river sediments reach their 'final' resting place and the rate of net accumulation exceeds the rate of erosion and resuspension. It is the accumulated products of this final stage that leave the most lasting geologic record. In the case of the Amazon delta, Pb-210 analyses of cores (Kuehl *et al.* 1986) indicate that on century timescales, roughly half the river's sediment load is accumulating on the mid-shelf (depth 30–50 m) at an average rate of $10\,\mathrm{cm\,yr^{-1}}$. The remainder of the sediment is sequestered within the subaerial delta. In contrast, the mild energy regime of the Mississippi Delta, together with a rather rapid rate of tectonic subsidence, has permitted the formation of thick accumulations at the mouths of prograding distributaries on the centuries timescale. On a longer timescale, episodic delta lobe switching has yielded multiple thick and elongate accumulations distributed along the Louisiana coast.

As deltas prograde seaward over a continental shelf and spread laterally along a coast, a subaerial delta plain usually surmounts the underlying subaqueous platform. Although this subaerial surface, which includes the intertidal environments, is typically quite thin vertically, at least in comparison to subaqueous deposits, it is this component that supports most human activities and with which people are most familiar. The suites of geomorphologic features that distinguish any particular deltaic surface are, as for the subaqueous delta, products of the coastal process regime that moulded the delta as well as of the regional climate and human land use practices. Other factors include the degree to which a delta is undergoing submergence because of rising sea level, tectonic subsidence or both. Wright (1985) describes some of the most common subaerial features.

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L.D. WRIGHT

RIVER PLUME

A plume is a vertically or horizontally moving, rising or expanding fluid body, such as the contrails of a fighter jet, emissions from a stack, or river discharge into a lake. Under the strong influence of gravity, rivers can enter a lake or ocean as a fully turbulent jet (Baines and Chu 1996), such as from discharge across steep riverbeds (> 0.5°), or under the influence of a flood wave. A river will reduce its velocity at its mouth, if it is moving too fast, by undergoing a hydraulic jump and thickening its flow (Bursik 1995). Many rivers discharge more slowly. The river's momentum and the hydraulic head at the river mouth carry the plume up to hundreds of kilometres into the lake or ocean, depending on the size and power of the river (Syvitski *et al*. 1998).

The plume's behaviour is dependent on the density contrast between the river water and the standing water. Compared to most lake water, the contrast in effluent density is small and controlled by the river's suspended load. Ocean water has a higher density, and the plumes often flow buoyantly on the surface (hypopycnal: Plate 99). The pathway that a hypopycnal plume will take, depends on a variety of factors:

1. Angle between the river course at the entry point and the coastline;
2. Strength and direction of the coastal current;
3. Wind direction and its influence on local upwelling or downwelling conditions;
4. Mixing (tidal) energy near the river mouth; and
5. latitude of the river mouth and thus the strength of the Coriolis effect.

Often there are strong interactions between these factors. For example if the angle of entry is in the direction of the Coriolis effect (i.e. move to the right in the northern hemisphere), then the plume will likely form a coast-hugging plume. Otherwise the plume will detach from the coast.

While hypopycnal plumes may form when river water enters a freshwater lake, they are just as likely to form hyperpycnal plumes (Plate 99). Sometimes referred to as underflows or turbidity currents, these dense flows penetrate the lake

(a)


(b)


Plate 99 (a) A hyperpycnal plume forming seaward of Skeiđaràrsandur (Iceland) (1996 photograph by M.T. Gumundsson and F. Pálsson). The surface plume disappears at the plunging point and subsequently flows seaward along the seafloor. (b) SeaWiFs image showing hypopycnal plumes emanating from the Mississippi River

under the influence of gravity, remaining in contact with the lake floor (Kassem and Imran 2001). If a hyperpycnal plume accelerates, additional sediment can be resuspended into the flow from the lake floor. As the hyperpycnal plume spreads and thickens due to the entrainment of ambient fluid, velocity is reduced and sediment is deposited. Hyperpycnal plumes rarely occur in rivers that discharge to the ocean (Wright *et al.* 1986). Globally, only a few dozen rivers generate hyperpycnal plumes on an annual basis, and most rivers would see such events happen once every hundred or so years, if at all (Mulder and Syvitski 1995).

Another major difference in comparing sediment plumes flowing into oceans and lakes is in the dynamics of particle settling. In freshwater environments, finer particles settle out of the plume slowly and individually. In ocean environments, river particles quickly flocculate and settle out rather quickly (Syvitski *et al.* 1995). Flocculation is the process that sees particles come into contact with one another and stick, wherein the new larger particles (flocs) have greatly enhanced settling velocities.

River plumes may also enter a lake or ocean at depth: tidewater glaciers directly discharge their stream water at or near the base of the ice front.

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JAMES SYVITSKI

RIVER RESTORATION

River restoration is a term used to describe a wide range of approaches aimed at improving the environmental quality of engineered river systems (see CHANNELIZATION; DAM; STREAM RESTORATION). The objective may be to recreate the river's natural forms and processes (sometimes referred to as 'naturalization'), although the assumption that nature can be created has been criticized. Restoration has also been described as 'nudging nature'. This reflects the fact that, to date, much restoration work has been focused on lower energy streams which have less ability to recover naturally following river channelization and therefore require active intervention (see Brookes 1987, for a discussion of the recovery of channel sinuosity on straightened rivers in Denmark). Furthermore, improvements in urban rivers, sometimes undertaken for aesthetic reasons, have also been referred to as 'restoration'. And some restoration schemes may also involve the development of a resource, such as a riverside wetland, that did not previously exist at the site. 'Creation' may therefore be a more appropriate term in these situations. Consequently, there is no simple definition of river restoration. However, Brookes and Shields (1996: 4) make the following useful distinction between enhancement, rehabilitation and restoration.

Enhancement they define to be 'any improvement in environmental quality'. This would include, for example, the increased diversity of marginal river vegetation achieved following works to raise flood banks on the River Torne (UK). The enhancement works comprised bank re-profiling to create narrow wetland shelves (berms), shallow bays, channel margins of varying shape and depth and linear still ponds from borrow pits on the floodplain (Clarke and Wharton 2000). The opportunity for river enhancement, which was achieved at negligible cost, arose because contractors changed the usual practice of importing materials and obtained the spoil for the flood banks from the channel margins and the floodplain. A large number of case studies describing river enhancement techniques are also illustrated in *The New Rivers and Wildlife Handbook* (RSPB *et al.* 1994) and there is much scope for these small-scale river improvements.

Rehabilitation, as defined by Brookes and Shields (1996: 4), is 'the partial return to a pre-disturbance structure or function' (see Brookes and Shields 1996 for examples from northern Europe and the USA). An example from the UK of a small-scale rehabilitation project is the Redhill Brook. This is typical of many lowland streams in England which have undergone rehabilitation since the mid-1980s. A 100 m reach had been artificially straightened and a further realignment was planned in 1991 as part of a floodplain development project. However, in issuing a land drainage consent for this work the National Rivers Authority required the realigned section to be designed to reflect the characteristics of a natural lowland stream. This included creating a channel with varying channel cross sections incorporating pools, riffles and point bars. Sediment was also reinstated which would not erode in a bankfull flood event (Brookes and Shields 1996: 246–247). Much more extensive rehabilitation has been undertaken on the rivers Brede, Cole and Skerne comprising a joint Danish and British EU–LIFE demonstration project (see Holmes and Nielsen 1998 for information on the background to the project; and Kronvang *et al.* 1998 for details on the restoration of the channel morphology and hydrology).

Finally, 'restoration' in its strictest sense is the term employed by Brookes and Shields (1996: 4) to describe 'the complete return to a pre-disturbance structure or function'. There are, however, several constraints to full river restoration. First, there is likely to be disagreement about the most appropriate pre-disturbance state. Practically, there is a need to establish whether the baseline for restoration should be set immediately before the most recent channelization works, before the first evidence for channel modification or at some point in between. Second, there is the related problem of establishing pre-disturbance data. Rarely are data comprehensive and accurate enough for reconstruction to be fully informed. In this context, Tapsell (1995) has asked 'what are we restoring to?' and Graf (1996, 2001) has discussed the issue of what is natural and how closely restored systems can approximate natural conditions. And third, the desirability of river restoration must be questioned. If sustainable river management is the aim, then the river must be considered within its catchment context, with river forms and processes able to respond to controlling factors such as flow regimes and sediment transport rates, that in turn respond to changing drainage basin conditions. A river restored to some pre-disturbance state is unlikely to be in balance with present conditions. Erskine *et al.* (1999) also document how restoration of the pre-dam situation on the Snowy River (Australia) was neither possible nor desirable because the conditions below the dam had stabilized themselves to a new regime.

Thus, the term rehabilitation is more appropriate in reflecting the reality of river restoration. In the UK, the River Restoration Centre, a non-profit making organization working to restore and enhance rivers, views restoration as a visionary target 'of pristine rivers that are wholly returned to an undisturbed state requiring no management' (Holmes 1998: 139) while accepting rehabilitation as a more practical alternative.

The involvement of all stakeholders, including the participation of the public, is a key element in the success of river restoration schemes by helping to engender a sense of ownership by the local community. In the EU–LIFE demonstration project on the River Skerne (UK) a community liaison officer was a vital member of the team working with the experts and local residents to ensure effective public consultation and dialogue (Holmes and Nielsen 1998). And Waley (2000) describes the socio-cultural value of river restoration in Japan and how science and ethics combine in restoration programmes.

The restoration of rivers may be driven by many factors, including environmental, economic and political. Attempts to restore the geomorphology, hydrology, water quality and ecology of rivers may arise from a desire to redress the environmental impacts of past engineering schemes (see CHANNELIZATION; DAM). Thus most river restoration activity is concentrated in developed countries which have a long history of river engineering. And in the UK, improvements to the physical habitat and ecology have been shown to be the main drivers behind restoration initiatives. The removal of dams which no longer generate hydro-electric power at a competitive rate and the restoration of a more natural flow regime and river environment has been reported in the USA (Graf 1996). This can have benefits for wildlife and generate income from the recreational use of the river (e.g. fishing and canoeing). Changes in environmental legislation also have a significant influence on river restoration. For example, in Denmark the 1982 Watercourse Act provides powers for safeguarding the physical environment of streams by focusing on ecologically acceptable maintenance practices, and incorporates special provisions for stream restoration and the potential for financial support of such activities. Across Europe, the European Union Water Framework Directive (2000/60/EC) is providing further impetus to river restoration by requiring member states to protect, enhance and restore all bodies of surface water not designated as artificial or heavily modified.

Brookes (1988: 217–237) and Wharton (2000) describe procedures for restoring channel capacity, river-bed sediments, cross-sectional form and pattern. Clearly, however, these components should not be viewed separately in the process of restoring a river's geomorphology. Based on research in Sweden, Petersen *et al*. (1992) advocate a 'building block approach' for restoring river environments in a number of stages. By combining different elements such as the construction of pools and riffles, the re-meandering of reaches and the creation of buffer strips and wetlands, the design and implementation of the restoration scheme can be tailored to specific sites. Brookes and Shields (1996) have also published guiding principles on river restoration and the UK River Restoration Centre has produced a second edition of its *Manual of River Restoration Techniques* (RRC 2002). By maintaining a database on completed projects and river restoration practitioners and researchers, the RRC also plays a pivotal role in disseminating information on river restoration and forms part of the European Centre for River Restoration (ECRR).

As more river restoration projects are undertaken it becomes increasingly important to appraise these schemes so that failures as well as successes can be documented and evaluated (Kondolf 1998). Importantly, it should be recognized that river restoration can have impacts on the fluvial system similar to those reported for channelization. Information from immediate post-project appraisal and longer term monitoring will help to develop the science of restoration. Specifically, there is a need to improve predictions of river channel sensitivity to change and to incorporate this understanding in integrated catchment management planning. There is also a need for further evidence on the link between the restoration of geomorphology and physical habitat and subsequent improvements to river ecology. And the appraisal of river restoration schemes in terms of their management and implementation will help inform the development of future policy and practice.

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SEE ALSO: anthropogeomorphology

GERALDENE WHARTON

ROCHE MOUTONNÉE

Roches moutonnées are asymmetric bedrock bumps or hills with one side ice-moulded and the other side steepened and often cliffed. They are widespread features in formerly glaciated hard-rock terrain and often to be found in clusters or fields. The name was first introduced by de Saussure (1786), based on a fancied resemblance to the wavy wigs of that period, which were called moutonnées after the mutton fat used to hold them in place. The term encompasses a wide range of feature sizes. Typically, roches moutonnées vary from 1 to 50 m in height and a few metres to hundreds of metres in length, but Sugden *et al*. (1992), for instance, describe large roches moutonnées hills in eastern Scotland with lee side cliffs up to 160 m high.

The morphology of roches moutonnées seems to reflect the contrast between ABRASION on the smoothed up-ice side and plucking on the lee side (see GLACIAL EROSION).

Abrasion acting on the stoss side is marked by STRIATIONs at a variety of scales together with polished facets and crescentic fractures on more gently sloping surfaces. As the glacier moves against the upstream side of an obstruction the ice overburden pressure increases. The basal ice may reach the PRESSURE MELTING POINT and partially melt, causing the glacier to slide. The direction of basal ice flow in this position is pointed towards the bed and the embedded clasts are dragged over the bedrock with some force, effectively abrading it.

On the lee side of the obstruction the ice overburden pressure is lower than average, encouraging the formation of a subglacial water-filled cavity. The presence of a cavity together with fluctuations of the water pressure within it strongly promotes the process of glacial plucking. Sugden *et al*. (1992) showed that block removal starts at the furthest point down ice in the cavity and from there extends successively up ice, thereby transforming the lee side of the bedrock bump into a staircase cliff. The detailed morphology of the plucked surface is also influenced by the properties of the parent bedrock, since plucking is encouraged by a favourable oriented system of joints.

Some roches moutonnées do appear to be only slightly modified preglacial hills, but in many areas initial bedrock eminences were clearly sculptured and reshaped by differential glacial erosion. Between these end-members there is likely to be a continuum of forms with varying degree of inherited topography. If the quarried and rough lee side as a distinctive feature of a roche moutonnée is little developed, it may also be difficult to distinguish roches moutonnées from whalebacks or rock DRUMLINs. Whalebacks are elongated and approximately symmetrical bedrock bumps whereas rock drumlins are asymmetrical with a steep upstream side and a gently inclined downstream side. Both are smoothed and rounded on all sides.

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CHRISTINE EMBLETON-HAMANN

ROCK COATING

About 15 per cent of the Earth's landscape consists of bare rock surfaces. Yet the common phrase 'bare rock' is truly a misnomer, because paper-thin accretions coat almost all of these rock surfaces in all terrestrial environments. Studies on the physical and chemical characteristics, origin, geography and utility of these deposits has spawned over 3,000 scientific papers. Plate 100 illustrates a few examples.

Alexander von Humboldt (1812) initiated the scholarly study of rock coatings by studying the composition, origin, spatial distribution and environmental relations of coatings such as those found along tropical rivers. In the past two centuries, researchers have documented hundreds of different types of rock coatings found within the fourteen major categories listed in Table 39.

The three most common rock coatings are rock varnish (see DESERT VARNISH), silica glaze and iron films. Silica glazes occur in warm deserts, cold

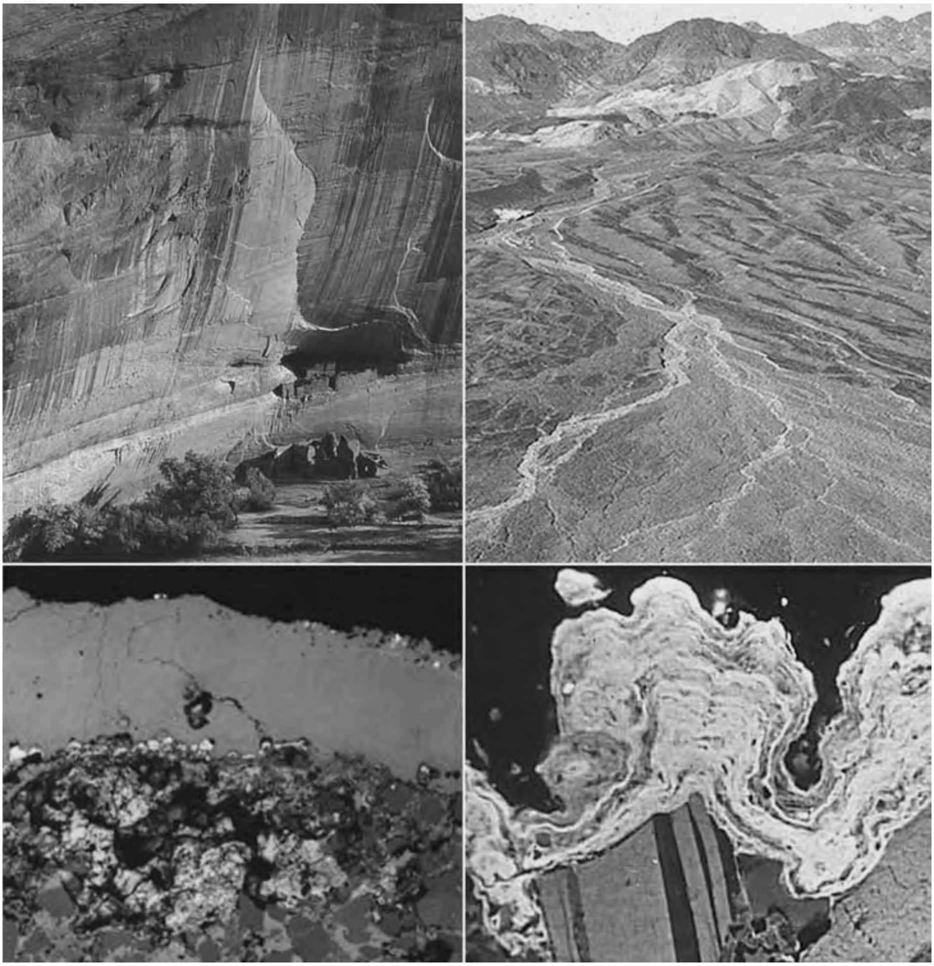

Plate 100 Upper left: a vertical face at Canyon De Chelly, USA, is streaked with heavy metal skins, iron films, lithobiotic coatings, oxalate crusts, rock varnish and silica glaze. Upper right: alluvial fan in Death Valley deposits the same light-coloured rock types in active streams. But over time, rocks in abandoned stream courses are darkened by desert varnish. Lower row: the electron microscope images (backscatter detector) illustrate that rock coatings are external accretions, exemplified by an oxalate crust in the lower left image that is about 0.5 mm thick and desert varnish on the lower right that is about 0.1 mm thick

deserts like Antarctica, on dry tropical islands, along tropical rivers, mid-latitude humid temperate settings, and various archaeological settings. Silica glazes probably precipitate from soluble Al-Si complexes $[Al(OSi(OH)_3)^{2+}]$ that are released from the weathering of clay minerals. Rust-coloured iron films display a wide variety of characteristics in very different climates and microenvironments. For example, rocks in the Dry Valleys of Antarctica host iron hydroxides that both form a micron-scale accretion and a weathering rind (see RIND, WEATHERING) over a millimetre thick. In a very different setting, iron oxyhydroxides impregnate rocks in arctic streams (Dixon *et al.* 2002).

Geomorphologists have long used intuition to interpret rock coatings and their relationship to the geomorphic setting. For example, some have believed that PEDIMENTs are fossil landforms, in part because the presence of rock coatings must

Table 39 Major categories of rock coatings

General type	Description	Related terms
Carbonate skin	Coating composed primarily of carbonate, usually calcium carbonate, but could be combined with magnesium or other cations	Caliche, calcrete, patina, travertine, carbonate skin, dolocrete, dolomite
Case hardening agents	Addition of cementing agent to rock matrix material; the agent may be manganese, sulphate, carbonate, silica, iron, oxalate, organisms or anthropogenic	Sometimes called a particular type of rock coating
Dust film	Light powder of clay- and silt-sized particles attached to rough surfaces and in rock fractures	Gesetz der Wüstenbildung; clay skins; clay films; soiling
Heavy metal skins	Coatings of iron, manganese, copper, zinc, nickel, mercury, lead and other heavy metals on rocks in natural and human-altered settings	Described by chemical composition of the film
Iron film	Composed primarily of iron oxides or oxyhydroxides; unlike orange rock varnish because it does not have clay as a major constituent	Ground patina, ferric oxide coating, red staining, ferric hydroxides, iron staining, iron-rich rock varnish, red-brown coating
Lithobiontic coatings	Organic remains form the rock coating, e.g. lichens, moss, fungi, cyanobacteria, algae	Organic mat, biofilms,
Nitrate crust	Potassium and calcium nitrate coatings on rocks, often in caves and rock shelters in limestone areas	Saltpetre; nitre; icing
Oxalate crust	Mostly calcium oxalate and silica with variable concentrations of magnesium, aluminium, potassium, phosphorus, sulphur, barium and manganese. Often found forming near or with lichens. Usually dark in colour, but can be as light as ivory	Oxalate patina, lichen-produced crusts, patina, scialbatura
Phosphate skin	Various phosphate minerals (e.g. iron phosphates or apatite) that are mixed with clays and sometimes manganese	Organophosphate film; epilithic biofilm
Pigment	Human-manufactured material placed on rock surfaces by people	Pictograph, paint, sometimes described by the nature of the material
Rock varnish	Clay minerals, Mn and Fe oxides, and minor and trace elements; colour ranges from orange to black produced by variable concentrations of different manganese and iron oxides	Desert varnish, desert lacquer, patina, manteau protecteaur, Wüstenlack, Schutzrinden, cataract films
Salt crust	The precipitation of sodium chloride on rock surfaces	Halite crust, efflorescence, salcrete

Table 39 Continued

General type	Description	Related terms
Silica glaze	Usually clear white to orange shiny lustre, but can be darker in appearance, composed primarily of amorphous silica and aluminium, but often with iron	Desert glaze, turtle-skin patina, siliceous crusts, silica-alumina coating, silica skins
Sulphate crust	Composed of the superposition of sulphates (e.g. barite, gypsum) on rocks; not gypsum crusts that are sedimentary deposits	Gypsum crusts; sulphate skin

infer long-term stability. Others have guessed at the ages of such features as flooding events on ALLUVIAL FANs, based on an intuitive feeling about the appearance of rock coatings (see gradual darkening of alluvial fan surfaces in the Plate 100). The complexities associated with formative processes have made rock coatings extraordinarily difficult to use as geomorphological tools to indicate either age or infer palaeoclimate. Rock coatings will be getting increased attention in future years as they are identified on Mars and as planetary scientists attempt to use rock coatings to infer Martian geomorphic processes (Kraft and Greeley 2000).

Rock coatings have applied significance in a variety of contexts. Heavy metal skins assist in identifying metal pollution (Dong *et al*. 2002). Some believe that artificial rock coatings have potential to aid in the conservation of priceless stone monuments (Borgia *et al*. 2001). Construction and development in desert regions contrasts bright uncoated rocks and darker natural rock coatings; the desire to live in natural-appearing settings leads to the application of artificial rock coatings to mimic natural colouration (Henniger 1995). Rock coatings, called patina in archaeology, are also used in the study of surface artefacts, rock paintings and rock engravings.

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SEE ALSO: alluvial fan; desert pavement; desert varnish; pediment

RONALD I. DORN

ROCK CONTROL

Rock control in geomorphology is defined as the influences of differences in earth materials on the development of landforms. Earth materials that form the Earth's surface or landforms are simply called landform materials, and include rocks, weathered materials and soil. The concept of *rock control* was first proposed explicitly and argued passionately by Yatsu (1966) and then expanded by him to a concept of *landform material science* in 1971. Yatsu stressed that to understand the formation of landforms, it is necessary to quantitatively evaluate the behaviours of landform materials in terms of their physical, mechanical, chemical and mineralogical properties in relation to the geomorphological processes concerned. His severe criticism of geomorphology is based on the fact that geologic structure and lithology have only been used qualitatively to explain the development of erosional landforms since the birth of modern geomorphology.

Typical examples often described in textbooks on geomorphology (e.g. Thornbury 1954; Sparks 1971) as structural landforms or landforms resulting from rock control include cuestas, hogbacks, mesas, structural benches, dyke ridges, knickpoints, karst and inversion topography. These landforms are relatively higher or steeper than their surroundings, and are generally composed of a relatively *resistant* or *hard* rock (e.g. sandstone, limestone, lava) that adjoins the relatively *less resistant* or *weak* rocks (e.g. mudstone, shale, tuff). However, rock control is not as simple as the vague terms *resistant* and *less resistant imply*. This is because the resistance and behaviour of landform materials vary markedly with geomorphological process and geomorphic setting.

For instance, the rocky coast of Arasaki, southwest of Tokyo, Japan, is underlain by steeply dipping, alternating beds of mudstone and scoria tuff. Differential erosion between the two rocks varies with altitude (Figure 134). On the vegetation-free sea cliffs behind the uplifted wave-cut benches, mudstone forms ridges and tuff forms shallow furrows. On the benches, mudstone forms the furrows and tuff forms the ridges, producing a washboard-like relief. On the shallow offshore sea bottom, there is no differential erosion. The mudstone is mechanically about two times as strong as the tuff. However, it is well jointed and forms fragments about 1 cm in size due to wet–dry slaking, whereas the tuff does not. The explanation for this differential erosion is that (1) both rocks are eroded at rates proportional to their mechanical strengths on the sea cliff above tidal zone, (2) the fragments of mudstone produced by wet–dry slaking are rapidly washed away by waves in the tidal zone, and (3) wave abrasion offshore erodes both rocks at the same rate (see Suzuki 2002).

Thus, the behaviour of landform materials generally does not merely depend on their geological structure and lithology, but also strongly depends on their physical and mechanical properties. This is because even lithologically similar rocks have wide ranges of physical and mechanical properties,

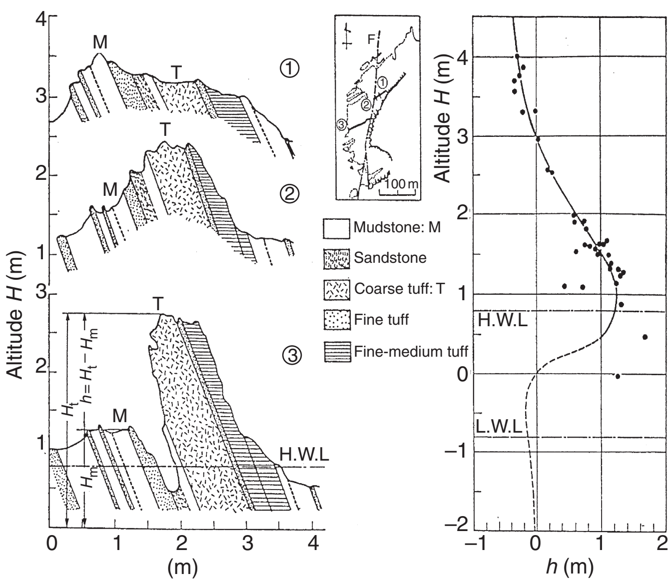

Figure 134 Change in relative relief between a tuff bed (T) and a mudstone bed (M) with altitude (H) on the Arasaki coast, Japan. *Left*: three geologic sections that are different in the altitude (H_m) of the mudstone surface (M). *Right*: relationship between relative relief (h) and H_m

reflecting their origin and history, such as diagenesis, tectonic deformation, unloading and weathering. Further, weathering results in changes in rock properties and hence is of importance as a preparatory process for erosion and mass movements.

Properties of landform materials are grouped into two major categories: geological properties and rock (material) properties. Geological properties are those described in terms of geology, and include lithology (such as grain size, mineral composition and texture), chemical composition, geological structure (such as stratification, joints, faults, unconformities, and their dip and strike), occurrence of rock masses (such as lava flow, dyke, batholith, etc.), weathering grades, and so on. Rock (material) properties, on the other hand, are those described in terms of physics and engineering (particularly rock and soil mechanics), and which can be further subdivided into physical properties and mechanical properties. Physical properties are the intrinsic characteristics (such as density, porosity and pore-size distribution) that do not depend on applied forces. Mechanical properties are the behaviours and responses of rocks against forces acting upon them, and hence vary with the kind of forces, the conditions of the rocks, such as water content, and the test methods used. Mechanical properties include strength (compressive, tensile, shear and bending strengths), hardness (e.g. abrasion, impact and scratch), deformation properties (e.g. deformation modulus, adhesive forces and internal friction), dynamic properties (elastic wave velocities), thermal properties (e.g. thermal conductivity and thermal expansion coefficient), permeability (e.g. permeability coefficient and infiltration capacity), behaviour in relation to water (such as swelling, slaking and solution) and so on.

These physical and mechanical properties are determined by the measurements of the rock mass in the field and for test pieces in the laboratory using precise instruments and equipment. Some practical test methods have also been applied to evaluate the mechanical properties of rocks, including standard penetration tests (N-value), rebound hardness with a Schmidt rock hammer, cone penetration hardness, needle penetration hardness for the rock mass, and rock quality designation for drilling cores.

Rock control problems are addressed by looking for the important rock properties in each process and quantitatively evaluating the roles of the properties with respect to the process. In the context of Yatsu's argument, therefore, physical, mechanical and chemical properties of landform materials have been intensively measured in both field and laboratory and in field and laboratory experiments since the 1970s, particularly by Japanese geomorphologists. Based on the measurements and experiments, much persuasive substantiation has been found for the modes and rates of various erosional processes and landforms (Suzuki 2002). Notable examples include formative processes along rocky coasts (Sunamura 1992), wind abrasion, lateral planation, slope evolution, hillslope morphology, valley development and some minor landforms such as tafoni. Processes and rates of bedrock weathering have also been studied actively in both field and laboratory.

Landform evolution is controlled not only by rock properties and geological properties but also by many other variables, such as geomorphological setting (initial landform), climate, geomorphic agents from which the various forces derive, and elapsed time. The research on rock control problems mentioned above, therefore, has been directed toward developing quantitative models of geomorphological processes that are capable of predicting types and rates of landform development. The models have been expressed as geomorphological equations including the geomorphic quantities concerned and the main controlling variables, i.e. site- and time-independent process equations with dimensionless constants. To develop the models, it is indispensable to study the rock control problems all over the world, because landforms are never changed unless the landform materials are moved. Thus, research on the rock control problems will be one of the core fields in process geomorphology in the twenty-first century.

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SEE ALSO: rock mass strength; weathering

TAKASUKE SUZUKI

ROCK AND EARTH PINNACLE AND PILLAR

Within areas built of poorly consolidated sediments, subject to intensive linear erosion, sheet wash and susceptible to weathering, bedrock may be sculpted into groups of weirdly shaped erosional residuals in the form of pinnacles, pillars and cones. They are relatively common in semi-arid areas, where scarce vegetation provides little protection against surface erosion, hence pinnacles are typical components of BADLANDS. Steep slopes underlain by erodible sediments, for example of newly deposited MORAINES, may also support pinnacle assemblages.

Two types of sediments yield to this type of erosional relief in particular, tills and pyroclastic deposits. Some tills and other glacial deposits contain boulders 'floating' in an otherwise fine-grained material. After exposure, boulders will protect an underlying softer mass against erosion, whereas the surrounding unprotected material will be eroded away, leaving the boulder-capped part standing as a residual pillar. Later, the boulder cap will provide a shield against the direct destructive impact of rain and the pillar may increase in height as long as the cap remains in place. Once the boulder falls from the top of the pinnacle, the column built of soft rock will rapidly be destroyed. Classic localities of this type of earth pinnacles have been described from the Tyrol in the Alps.

Plate 101 A group of rock pinnacles in Cappadocia, central Turkey. Remnants of resistant welded tuff act as a cap to the underlying softer sediment

In pyroclastic deposits, volcanic bombs within softer tuff play the similar protective role as boulders in tills do. In Cappadocia, central Turkey, bomb-capped pyramids reach up to 20 m high. In other cases, caps are provided by remnants of a welded tuff horizon overlying thicker and softer strata beneath.

Not all rock and earth pinnacles have a protective cap, and there are other reasons why they remain as isolated residuals. In the semi-arid badlands of Cappadocia, surfaces of tuff cones isolated by fluvial and sheet wash erosion are subject to case hardening, and it is the crust which protects the cones from further destruction. Owing to the presence of the crust, the earth pinnacles of Cappadocia could have grown up to 15–20 m high and were found stable enough for churches, hermitages and cave dwellings to be dug into them in early Christian times.

Tufa (see TUFA AND TRAVERTINE) deposits may form curiously shaped pinnacles too, but in these cases pinnacles are constructional, and not erosional features. At Mono Lake, California, and Searles Lake, California, tufa pyramids and pillars up to 15 m high formed around underwater springs and were exposed at the surface, when lake levels were lowered.

SEE ALSO: hoodoo

PIOTR MIGOŃ

ROCK GLACIER

Rock glaciers (German: *Blockgletscher, Blockstrom*; French: *Glacier rocheux*; Russian: *Kammeni gletscher*; Spanish: *glaciar rocoso*) occur in most alpine-mountainous regions and are distinct tongues of rock rubble which flow slowly downhill. Most features are elongate and are generally distinct from blockstreams (see BLOCKFIELD AND BLOCKSTREAM) which occur on very low angle slopes.

The substantial literature on these features is complex and often confusing, being hindered by difficulties of *in situ* investigation. Markedly different viewpoints have been taken to explain their origin, dynamic behaviour and environmental significance (Potter *et al.* 1998). It is important to avoid explicit designation of an origin so they are best defined by their morphology and appearance; a simple morphological definition, after Washburn (1979), is: ‘a tongue-like body of angular boulders that resembles a small glacier, which generally occurs in high mountains and usually has ridges, furrows and sometimes lobes on its surface with a steep front or snout at the angle of repose’.

Plate 102 An active rock glacier (maximum velocity about $0.25\,\mathrm{m\,a^{-1}}$) in northern Iceland with a small corrie glacier at its head. There is a gradation from a thin cover of debris to much thicker debris (*c.*1 m) near the snout. The lateral margins are distinct from the sides of the valley and the longitudinal furrow is conspicuous. The whole rock glacier is about 800 m long and estimated to have formed in the past 200–300 years

Plate 103 Merging rock glaciers, Wrangell Mountains, Alaska. These are typical rock glacier forms, emerging from corries now containing little or no ice. That there is probably very little forward movement of the wrinkled and furrowed surfaces is indicated by the vegetated surface

Distribution maps and reviews can be found in Whalley and Martin (1992) and Barsch (1996). They may even occur on Mars (Whalley and Azizi 2003). Although originally thought to be confined to continental areas and to give way to glacier bodies in more maritime regions examples in the latter have been found. They were first recognized in North America and Greenland (Martin and Whalley 1987; Barsch 1996).

The surface velocity is generally $<1\,\mathrm{m\,a^{-1}}$, although some with velocities $>5\,\mathrm{m^{-1}}$ have been described (Gorbunov *et al.* 1992). If no movement can be detected they are generally referred to as ‘relict’ and, if highly vegetated with subdued features, as ‘fossil’ and are recognized by morphology alone. However, ‘inactive’ rock glaciers are sometimes recognized where creep rates are very low and may even have trees growing on them. The steep fronts (snouts) may advance over other features; valley floors, moraines and lakes. These characteristics, variable over time, make it difficult to show that they result from one origin or relate to a single set of environmental conditions.

Rock glaciers are generally about 1 km long but many shorter examples exist and some may be up to 3 km long; width is generally a few hundred metres. Typically, they have their heads in corries (see CIRQUE, GLACIAL) in which there may be a small glacier, although this may not always be visible. The elongate forms are regarded as rock glaciers proper, but have also been called ‘tongue-shaped’, ‘valley floor’ or ‘debris rock glaciers’. Some forms may be ‘spatulate’ where they spread over a main valley floor. Rock glaciers are usually separated from valley sides by ‘lateral furrows’. The term ‘rock glacier’ has also been used for features which are broader than long and which typically have their upper sections against cliffs or scree rather than emanating from corries. This form has been called a ‘valley side rock glacier’, ‘lobate rock glacier’ or ‘talus rock glacier’. It has been suggested that the latter features are best termed ‘protalus lobes’ rather than rock glaciers because of the differences in form and location (Hamilton and Whalley 1995).

Flow-related features are commonly seen as ridges and furrows on the surface although it is not known if these extend to any depth. Some rock glaciers have mainly transverse ridges, especially near the snout, others have a predominantly longitudinal ridge pattern. Such flow features have been related to flow regime; compressing or extending (Whalley and Azizi 2003). Many rock glaciers have distinct longitudinal furrows and some show pools or small lakes developed in flatter areas ('thermokarst ponds').

There are four main theories of rock glacier origin. One suggests that they are glacially derived with a veneer of weathered rock debris (a few cm to >1 m thick) over a thin (<50 m) glacier ice core. The debris protects a thin body of ice which flows only slowly. This may be termed the 'glacial model'. It has been suggested that rock glaciers are nothing more than debris-covered glaciers. However, the subdued dynamic behaviour of rock glaciers indicates that ice volumes are small and that SUPRAGLACIAL debris has been supplied via the surface of the small glacier. Debris-covered glaciers gain debris in their lower reaches by ablation of ice releasing englacially transported debris but there is probably a transition between the two. What gives rise to rock glaciers is the relative abundance of debris to active glacier ice.

The 'permafrost model' explains the slow movement as creep of ice dispersed within weathered rock debris (derived from SCREE) and that a glacial derivation is not necessary to explain flow. It does require PERMAFROST conditions (mean annual air temperature $<-1.5\,^{\circ}C$) for the formation and continued creep of the ice. The ice needs to be above 'saturation', i.e. more than fills void spaces, or as ice lenses, for creep to be efficient. Ice-cemented debris (at or below saturation) colder than the PRESSURE MELTING POINT of ice is mechanically stronger than ice and will not flow unless at high shear stresses (high surface slope and/or thickness).

The third model suggests that some rock glaciers (or protalus lobes) are formed by sudden, perhaps catastrophic, failure of scree slopes (Johnson 1974) or as a single catastrophic rock avalanche (Bergsturz or STURZSTROM). This view is not widely held although there is evidence that *some* rock glacier forms might have been constructed in this way to produce topography similar to a rock glacier. Where the features are old there might be confusion with a fossil rock glacier.

A fourth model, a variant of the first (glacier-derived) and third (catastrophic), is that a rockfall covered a small or decaying glacier. The thin debris cover would thus insulate the thin glacier core but suddenly rather than progressively.

Plate 104 Complex protalus lobes, Alpes Maritimes, southern France. The inner ridges look a little like protalus ramparts and the feature lies below a cliffed area which probably had a thin glacier or large snowpatch at its foot. These features differ in form to rock glaciers per se

Rock glaciers have been used as indicators of permafrost (past permafrost for relict features) but only if the permafrost model is valid. This may be considered as being a 'zonal' model. The glacial model is thus 'azonal' as the contributing glacier may occur whether or not permafrost is present (Washburn 1979).

The origin and flow mechanism of rock glaciers is frequently attributed exclusively to creeping permafrost (Barsch 1996). Although traces of glacier ice have been seen in some rock glaciers, permafrost conditions were considered to be the only way in which the features could exist. Observations of glacier ice down the length of some rock glaciers have now been established and thus show that at least some rock glaciers have glacier ice cores. It is possible that modern dating and isotopic techniques will allow ice from such rock glaciers to provide a climatic record. The full implications for recognizing climate change through rock glaciers still needs investigation.

Geophysical measurements (seismic, gravity, resistivity and ground penetrating radar) have been used to investigate the structure of rock glaciers. Resistivity has been used to differentiate between the mode of ice formation. It is claimed that high resistivity ($>10\,M\Omega m$) is indicative of glacier ice but that rather lower values are typical of ice of permafrost origin. This has been disputed by some authors who claim that the high resistivity is not typical of the small glaciers which provide glacier cores because such ice is contaminated by dust which lowers the value. The difficulty is of linking geophysical signals with an appropriate mixture (ice and debris) model (Whalley and Azizi 1994). The complexity is enhanced because there may be grades of mixture, from permafrost to glacier ice core, in one feature and is particularly significant near rock glacier snouts. This ambiguity of origin also suggests that using rock glaciers to identify past conditions may be difficult.

'Protalus lobes' are related to rock glaciers where permafrost conditions may be required to preserve ice but where a glacier is unlikely to have formed. These are distinctive enough to be given a separate name. PROTALUS RAMPARTs are long, rather narrow, ridges below cliffs and are thought to have a snow-bank (nival) origin. Suggestions have been made that they might be precursors to rock glaciers of permafrost origin (Barsch 1996).

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BRIAN WHALLEY

ROCK MASS STRENGTH

Rock mass strength (RMS) refers to the specific properties of the rock mass that control its strength and subsequent rock slope stability. Importantly, it allows the prediction of the stable inclination of natural rock slopes, as well as the recognition of strength EQUILIBRIUM SLOPEs in the landscape adjusted to the prevailing SUBAERIAL processes. The standard method of RMS determination applied in geomorphology was developed by Selby (1980, 1982) and has been extensively tested over the past twenty years as a reliable method for the assessment of rock slope stability. This classification scheme is a modification of RMS classifications developed for engineering purposes (e.g. Bieniawski 1979) which have been extensively used to aid excavation design for tunnels, slopes and foundations. However, these classifications often do not incorporate a quantitative assessment of the influence of a reduction in rock mass strength due to WEATHERING. Furthermore, the engineering classifications contain different definitions of the rating classes so that the results derived from the various methods are not directly comparable. The method of Selby (1980), and modified by Moon (1984), explicitly and quantitatively incorporates and weights the influence of weathering on the strength of the rock mass in the field through evaluation of intact rock strength, estimation of state of rock weathering, joint spacing, continuity and infilling. Since weathered rock is the norm, the scheme developed by Selby (1980) is a more appropriate measure of the RMS of a natural rock slope than those developed for engineering purposes.

Although geomorphologists have long recognized that rock slope failure often occurs along discontinuities such as joints, bedding planes and faults (see FAULT AND FAULT SCARP) rather than through intact rock, logistical difficulties and frequent inability to access the appropriate equipment for the laboratory and field assessment of rock strength has often meant that studies of the strength of the rock mass tended to be qualitative in nature. It is in this context that the development of the rock mass strength classification has had important consequences for our understanding of the morphology and evolution of rock slopes as it provides a basis for understanding the features of the rock mass that provide resistance to weathering and EROSION, as well as the maintenance of slope stability. The only equipment required is a SCHMIDT HAMMER, tape measure and inclinometer, and it can be applied to any rock mass where there is enough exposure to allow measurement of the rock JOINTING (usually at least $10\,m^2$). If the slope contains more than one morphological element, it must be subdivided into zones with similar RMS properties, with the RMS assessment undertaken within each slope element.

The rock mass strength classification system developed by Selby (1980, 1982) was based on an examination of rock slopes in Antarctica and New Zealand, and has subsequently been applied in a range of settings (e.g. Augustinus 1992; Moon and Selby 1983; Allison and Goudie 1990). Slopes adjusted to their RMS (strength equilibrium slopes) are common in nature, and the recognition of over steepened slopes that have been undercut by erosion, as well as structurally controlled rock benches of lower slope angle and RICHTER DENUDATION SLOPES, indicates its utility in resistance-form studies. The RMS classification involves measurement of a range of properties of the rock mass: (1) Schmidt hammer impact as a measure of the strength of the intact rock; (2) state of rock weathering; (3) jointing characteristics of the rock mass: spacing of rock joints, joint width, joint continuity, joint infilling and orientation of the dominant joint set; and (4) water seepage from the rock face (Table 40). Since not all these parameters are of equal importance in controlling rock mass strength, each of these factors is weighted and given a rating value according to their perceived influence on stability of the rock slope using the scheme given in Selby (1980) or as modified by Moon (1984). However, the usefulness of the further subdivision of the classification proposed by Moon has been questioned, so that the simpler scheme of Selby (1980) is preferred. The sums of the individual weightings for the rock mass being evaluated is its RMS rating. A maximum value of 100 applies and the range is divided into five classes (Table 40). The higher the

Table 40 Geomorphic rock mass strength classification and ratings

Criteria	*(1)* Very strong	*(2)* Strong	*(3)* Moderate	*(4)* Weak	*(5)* Very weak
Intact rock# strength	100–60	60–50	50–40	40–35	35–10
Rating	*20*	*18*	*14*	*10*	*5*
Weathering	Unweathered	Slightly weathered	Moderately weathered	Highly weathered	Completely weathered
Rating	*10*	*9*	*7*	*5*	*3*
Joint spacing	>3 m	3–1 m	1–0.3 m	0.3–0.05 m	<0.05 m
Rating	*30*	*28*	*21*	*15*	*8*
Joint orientation	>30° into slope	<30° into slope	Horizontal and vertical	<30° out of slope	<30° out of slope
Rating	*20*	*18*	*14*	*9*	*5*
Joint width	<0.1 mm	0.1–1 mm	1–5 mm	5–20 mm	>20 mm
Rating	*7*	*6*	*5*	*4*	*2*
Joint continuity	None continuous	Few continuous	Continuous, no infill	Continuous, thin infill	Continuous, thick infill
Rating	*7*	*6*	*5*	*4*	*1*
Groundwater outflow	None	Trace	Slight	Moderate	Great
Rating	*6*	*5*	*4*	*3*	*1*
Total rating	100–91	90–71	70–51	50–26	<26

Source: Modified from Selby (1980), and Moon (1984)
Note: # N-type Schmidt hammer rebound values

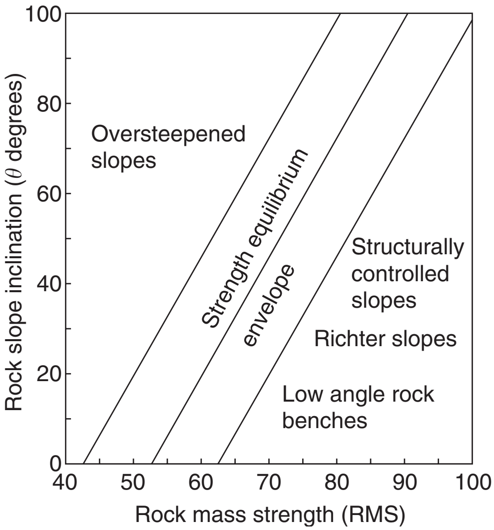

Figure 135 Plot of slope gradient and rock mass strength, with strength equilibrium envelope (after Moon 1984; Abrahams and Parsons 1987)

RMS rating, the higher mass strength and the steeper the slope inclination that can be sustained.

The graphical representation of the RMS classification involves plotting the total RMS rating against the slope inclination at each site (Figure 135). Note that the slope (θ) vs RMS graph is accompanied by an equilibrium line which relates the RMS rating to the stable slope angle, as defined by numerous measurements of slopes assessed to be in a stable equilibrium condition (Selby 1980, 1982). Superimposed on this plot is the RMS envelope as modified from that of Selby (1980) by Moon (1984), and further refined by Abrahams and Parsons (1987). Within this envelope there is a 95 per cent probability that the slopes are in strength equilibrium (Figure 135). Abrahams and Parsons (1987) re-evaluated the published RMS data for strength equilibrium slopes and produced a more statistically rigorous relationship between slope inclination and RMS. Using this plot and envelope, predictions of stable slope angles can be achieved, and it is possible to identify equilibrium or non-equilibrium slopes, with the latter either oversteepened or low angle, structurally controlled or Richter denudation slopes (Figure 135).

Strength equilibrium slopes have inclinations in balance with their RMS and are not controlled by other exo- or endogenic processes such as structural or tectonic factors. These slopes also require considerable time for this balance to develop (>10,000 years) so that young slopes are often not in strength equilibrium (Selby 1987). Nevertheless, many slopes have an inclination adjusted to their RMS, and oversteepened slopes undercut by processes such as GLACIAL EROSION can be easily recognized, although Augustinus (1995) demonstrated that equilibrium and structurally controlled slopes are more common in youthful, tectonically active mountains with deeply incised glacial valleys. Furthermore, the widespread recognition of strength equilibrium rock slopes suggests that many of them probably retreat whilst preserving strength equilibrium (Moon and Selby 1983; Selby 1987). Consequently, a change in the slope angle during retreat can occur where RMS changes as a consequence of progressive WEATHERING or rapid rupture of the rock mass as a consequence of external factors such as earthquake shaking. The tendency for slopes to equilibrate rapidly as a consequence of a change in RMS means that oversteepened slopes will have a short life span (on a geological timescale) before they evolve towards strength equilibrium forms as soon as fractures open, increase in continuity, widen or rotate.

The importance of rock mass control in geomorphology is exemplified by the application of the rock mass strength classification to the development of an understanding of rock slope form evolution and stability. However, rock mass strength and its resistance to EROSION processes may also be crucial to understanding the evolution of erosional landforms. For example, the development of features such as glacial valley longitudinal profiles (as well as the glacial valley cross-profile forms) will be dependent on the RMS of the rock being eroded as well as the EROSIVITY of the processes, since the intact rock strength, orientation of the rock joints and their spacing will influence the rock mass resistance to glacial erosion processes such as plucking. Clearly, in this situation the rock mass properties that control the stability and morphology of slopes will not be applicable to quantifying rock resistance to erosion and would require redefinition for this purpose.

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PAUL AUGUSTINUS

ROCKFALL

Rockfall is the free or bounding fall of rock debris down steep slopes. Rockfalls vary from individual pebbles to catastrophic failures of several million cubic metres (STURZSTROM, rock avalanches). Smaller rockfalls ($<10^1$–$10^2\,m^{-3}$) are the primary process associated with the formation of SCREE (talus) slopes and may be classified into two types (Rapp 1960). Massive vertical cliffs are dominated by primary rockfalls where detachment is followed by direct transfer to the scree below. These are triggered mainly by pressure release or FREEZE–THAW CYCLE activity. However, debris may accumulate on irregularities in the cliff (benches, gullies, etc.) and subsequently be dislodged by other rockfalls, snow avalanches (see AVALANCHE, SNOW), surface flows, animals, etc. These secondary rockfalls have different magnitude–frequency characteristics than primary rockfalls. Triggering mechanisms for rockfalls are inferred from inventory studies that compare the pattern of rockfalls with simultaneous observations of temperature and precipitation. Most investigators identify diurnal maxima during times of solar illumination and seasonal maxima in spring and fall (Luckman 1976; Gardner 1980).

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SEE ALSO: geomorphological hazard; hillslope, process; mass movement; pressure release; unloading

BRIAN LUCKMAN

ROCKPOOL

Rockpools (synonymous with tidepool, pool) can broadly be defined as depressions in eulittoral and supralittoral rocky SHORE PLATFORMS which store surface water and form as a result of dissection of rock material by a combination of chemical, physical and/or biological means. It is generally accepted that the presence or creation of an initial depression enables the commencement of a positive feedback loop – where surface water storage provides a suitable environment where weathering and erosion processes widen and deepen pits in rock surfaces to develop rockpools (Elston 1917).

Rockpool development can be divided into three phases: (1) pool initiation, (2) pool widening and deepening and (3) coalescing of smaller pools. Pool initiation is thought to be largely controlled by geological conditions such as rock hardness (with softer rocks such as sandstone and limestone being more prone to erosion), joint planes, irregular bedding and concretions which provide an initial depression from which rockpools gradually develop. Pool deepening and widening are often caused by a suite of biological, chemical and physical processes. Biological processes include bioerosion by boring and/or grazing species such as polychaete worms, sea urchins and limpets.

Dissolution is often caused by biochemical activity when respiration increases the CO_2 concentration in the pool. Chemical weathering also occurs due to evaporation and drying of seawater. The dominant physical erosion process is scouring and abrasion of pools by harder rock debris being moved or rotated by the action of waves. The third phase of pool formation is caused by the continual erosion of narrow walls separating adjacent pools. This leads to the coalescing of smaller pools into large irregular or elliptical pool forms and, in some instances, secondary, inset pools develop as another line of stratification is eroded in the base of pools.

Rockpools vary in size from small features a few centimetres in diameter to large, irregular forms which are up to 6 m in diameter and range from 0.1 to 2 m in depth. Two main types of rock pools have been defined in geomorphological literature: solution pools and pot-holes. Solution pools (synonymous with shallow pools) are typically defined as shallow, flat-bottomed depressions found on gently sloping shore platforms (Sunamura 1992). While solution pools have greater width than depth and vary in shape, pot-holes are typically cylindrical or bowl-shaped depressions that have more similar depth to width ratios. Pot-holes are thought to form primarily by abrasion. This classification is quite narrow in scope as rockpools typically form by a combination of biological, chemical and physical means rather than being dominated by one process over another.

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LARISSA NAYLOR

ROCKY DESERTIFICATION

Rocky DESERTIFICATION is the process that leads to KARST lands being turned into stony ecological deserts. It is a consequence of devegetation followed by intensive agriculture and extreme soil erosion. Soils on karsts are usually thin, because limestones are often composed of at least 90 per cent calcium carbonate and so there is little insoluble residue from their solution that can form the mineral basis of a soil. In the humid subtropical karst of China, it has been calculated that 0.25–0.85 million years are required to form 1 m of soil. Where thick soils are found on karst, it is usually because they are formed of foreign materials transported from beyond the boundary of the karst, such as loess, alluvium, volcanic ash or glacial drift. But even thick soils can be stripped from karst, although it takes longer.

It is well known in any environment that deforestation leads to accelerated SOIL EROSION. In karst this is exacerbated because of soil loss down countless voids opened by corrosion into caves, where it has a major deleterious impact on subterranean biota. It is eventually evacuated from cave systems via underground streams that discharge at springs, but lowered water quality results. The free draining nature of karst therefore contributes to the loss of its soil if the fragile hold accomplished by plant cover is disturbed.

In parts of the Mediterranean basin, the rocky nature of karst is so characteristic that it has come to be considered natural, rather than a consequence of millennia of human impact (Gams *et al*. 1993). The word *karst* itself can be traced back to pre-Indo-European origins, where it stems from *karra* meaning stone. But the landscapes involved were originally forested and have been cleared, tilled and overgrazed. The first evidence of forest clearance around the northern Mediterranean was in the Neolithic about 6,000 years ago. This continued through Greek, Roman and more recent times, and as population increased lands were subdivided and grazing became more intensive. This relentless impact contributed to the stripping of the hills; so that naked karst now seems the norm. But recent political and land use changes in Slovenia have seen rural migration and abandonment of farms, followed by natural regrowth and spread of forests, indicating that recovery is possible if human pressure is reduced.

A similar sequence of events has occurred in China (Yuan 1995), especially in Guizhou Province, where removal of subtropical monsoon forest and intense population pressure in recent centuries has seen the denudation of karstic hillsides. The expansion of rocky desertification in the area was at a rate of 933 square kilometres per year during the 1980s. Similar problems though of smaller scale are encountered in deforested and intensively farmed karstlands of the Gunung Sewu of Java and in parts of Central America and the Caribbean.

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PAUL W. WILLIAMS

ROUGHNESS

The term 'roughness' refers to the roughness of a channel bed, which is an important component of the overall resistance to water flow along the channel. Water flows along a channel under the influence largely of two forces: the downslope component of its own weight (which acts to propel it along the channel) and the resistance of the channel (which acts to hold it back). If the resistance is low, then a given flow has a high velocity and a low depth. If the resistance is high, the same flow has a low velocity and a high depth. Quantification of flow resistance is thus fundamental to the calculation of flow conditions in a channel.

Several relationships linking flow resistance, velocity and depth have been in use for a century or more, each quantifying the resistance with a single coefficient. They are the Darcy–Weisbach equation:

$$U = (8\,g\,R\,S_f / f)^{1/2}$$

the Manning equation (in SI units):

$$U = R^{2/3}\,S_f^{1/2} / n$$

and the Chézy equation:

$$U = c\,(R\,S_f)^{1/2}$$

where U is mean flow velocity, R is hydraulic radius (flow cross-sectional area/channel wetted perimeter), S_f is friction slope (a measure of energy loss often approximated by water surface slope), g is acceleration due to gravity and f, n and c are respectively, the Darcy–Weisbach, Manning and Chézy coefficients. The central problem in flow resistance is therefore evaluation of the coefficient.

In a straight channel of uniform slope, uniform cross section and large width/depth ratio with no sediment transport or BEDFORMS, the resistance to flow is determined primarily by the frictional resistance of the bed. This varies with the roughness of the bed, itself dependent on the material of which the bed is composed, e.g. sand or gravel. In a popular approach, fluid mechanics and boundary layer theory are invoked to calculate resistance as a function of the logarithm of relative roughness, defined as the ratio of bed roughness height to flow depth. Roughness height is often evaluated as an equivalent sand grain size or as a selected percentile from the measured size distribution of the bed material. For example

$$(8/f)^{1/2} = 5.62 \log (d/D_{84}) + C$$

where D_{84} is the particle size for which 84 per cent of the particles are finer and C is a coefficient. Because theory cannot yet provide a full quantification of the resistance, the coefficient is determined empirically from experimental data.

Natural channels rarely conform to the ideal conditions described above and additional terms may be required in the resistance equation to account for deviations from these conditions. For example, sand beds develop bedforms such as ripples and dunes which increase the resistance above that of the roughness of the grains alone. Consequently there are a variety of formulae and methods for determining the resistance coefficient. Users should be careful to select a method which is appropriate for the conditions and data availability with which they are concerned.

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JAMES C. BATHURST

RUGGEDNESS

A property of the landscape which describes the complexity of the topography and the roughness of the terrain. More rugged landscapes tend to exhibit a greater amount of complexity, having rough and uneven surfaces. Ruggedness is a naturally qualitative term, though several ruggedness indexes have been proposed (e.g. Riley *et al*. 1999) that provide a quantitative frame. Melton (1958) developed the ruggedness number to describe the ruggedness of land on a drainage-basin scale. This is a dimensionless number calculated from the formula $H/\sqrt{A}$ where H is the vertical relief above fan apex (miles^2), and *A* is basin area (miles^2). In general, the ruggedness number can be as high as 2.0 or 3.0 for a first or second-order basin, and rarely above 1.0 for a third or fourth-order basin.

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STEVE WARD

RUNOFF GENERATION

Runoff generation refers to a suite of processes that produce and route flow from landscape segments to stream channels in response to precipitation (i.e. rainfall and/or snowmelt). Runoff is generated by three different mechanisms: infiltration-excess overland flow (Horton-type), saturation-excess overland flow (Dunne-type) and subsurface stormflow. Infiltration-excess overland flow is overland flow that results from saturation from above (Horton 1933, 1945). This occurs where the water-input rate exceeds the infiltration capacity of the soil long enough for ponding to occur; the excess water then flows quickly over the surface to stream channels (see QUICK FLOW). Once the precipitation volume exceeds the moisture storage capacity of the soil, however, saturation-excess overland flow occurs. First described in detail by Dunne and Black (1970), this mechanism is controlled by the saturated hydraulic conductivity of the soil. The soil becomes saturated from below, either by (1) the presence of an impeding layer which causes a perched water table to develop that may gradually rise to the surface; (2) extension of the capillary fringe to the ground surface; or (3) the presence of a permanent water table at or near the ground surface. Saturation-excess overland flow consists of direct water input to the saturated area plus the return flow contributed by the exfiltration of ground water from upslope.

The third runoff mechanism, subsurface stormflow, consists mainly of the displacement of old pre-event water by new rainwater. This is due to near-stream ground water mounding (the same process that ultimately produces saturation overland flow) and/or via flow from perched saturated zones (saturated wedges). The process also includes throughflow or interflow, which is water that infiltrates into the soil and moves laterally within the soil matrix either in the unsaturated zone between the ground surface and a perched or regional water table or through macropores such as cracks, root and animal holes and pipes (Wilson *et al*. 1990; Jones 1997). The latter reaches the stream channel quickly and differs from other subsurface flow by the rapidity of its response and its relatively large magnitude.

It is widely accepted that Hortonian overland flow is the dominant response mechanism in semi-arid to arid regions, on impermeable zones and in human-disturbed areas. On the other hand, the response mechanism in humid regions is generally saturation-excess overland flow and/or some form of subsurface flow. Saturation-excess overland flow most frequently occurs near stream channels but it also can be generated in hillslope hollows, where subsurface flowlines converge in slope concavities, at concave slope breaks, or where soil layers conducting subsurface flow are locally thin. Subsurface storm flow dominates where soils are deep and permeable, especially under forest cover, and where slopes are steep. All three runoff mechanisms can occur simultaneously in a basin, even during a single water-input event.

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SEE ALSO: quick flow

MICHAEL SLATTERY

S

SABKHA

A sabkha is the English form of the Arabic word *sebkha* which means 'salt flat'. Kinsman (1969) defines a sabkha as a surface of deflation down to the level of ground water or the zone of capillary evaporation. Neal (1975) describes a sabkha as a geomorphic surface the level of which is dictated by the water table. Warren (1989) and Briere (2000) describe a sabkha as a marginal marine and continental mudflat where evaporite minerals are forming in the capillary zone above the saline water table. Sabkhas were described subsequently in many other areas of the world, such as the coast of Baja California, Mexico and the coast of Sinai. Equivalent features are the Solonchak salt flats of the Caspian Sea and some kavir depressions of Iran. Certain salt pans in South Africa and playas in the southwestern United States may be similar but true sabkhas are not fed by streams or runoff.

In North America, the words 'playa' and 'salina' have both been applied to sabkha-like areas in the desert. Holm (1960) states that 'playa' is synonymous with 'mamlahah' (inland sabkha). Von Engeln (1942), on the other hand, says that if the percentage of salts in a playa is high enough for a salt crust to form when the flat is dry, it is then called a salina.

If the word sabkha means 'salt flat' then there are both coastal and continental sabkhas. Some coastal sabkhas, such as those along the Trucial Coast, pass laterally into continental sabkhas without any noticeable change in surface morphology on the sabkha (Kinsman 1969). The marine portion of the Abu Dhabi sabkha is characterized by a matrix of marine sediments soaking in largely marine-derived ground water and the continental portion by non-marine sediments and ground water.

Modern marine sabkhas are forming along the coasts of many stable land areas such as the western and southern coasts of the Arabian Gulf, the coasts of Australia and northern Africa. Some sabkhas are slightly above present sea level (0.5–3.0 m) and may have been inherited from short Mid-Holocene eustatic oscillations.

A review of the global distribution of sabkha indicates its extensive presence in the Middle East, including Egypt, Sudan, Libya, Tunisia, Algeria and Ethiopia. Sabkha also exists in India, Australia and southern Africa. Contrary to expectations, sabkha and sabkha-like sediments can occur also in relatively cold climates. Aridity, therefore, seems to play a more important role than hot weather in the formation of sabkha. Figure 136 shows the distribution of sabkha around the world.

Sabkhas are part of a landform sequence that extends from the shoreline, with barrier islands or dunes, through a lagoon then to the sabkha and perhaps into a dune system before truly terrestrial systems are reached. Sabkha surfaces are extremely flat and often extend for long distances. The sabkha depositional sequence can be divided into three units: the subtidal, the intertidal and the supratidal. When the sequence is prograding the three units are superimposed one on top of the other to form a shallowing upward sabkha cycle.

The subtidal unit is usually divided into open marine and restricted marine sedimentation. Restricted marine sedimentation occurs on the landward side of barrier islands. The lagoon sediments contain a diverse biota of many benthic species, molluscan sands occur in the more restricted areas and pelleted carbonate mudstones occur in the more open areas of the

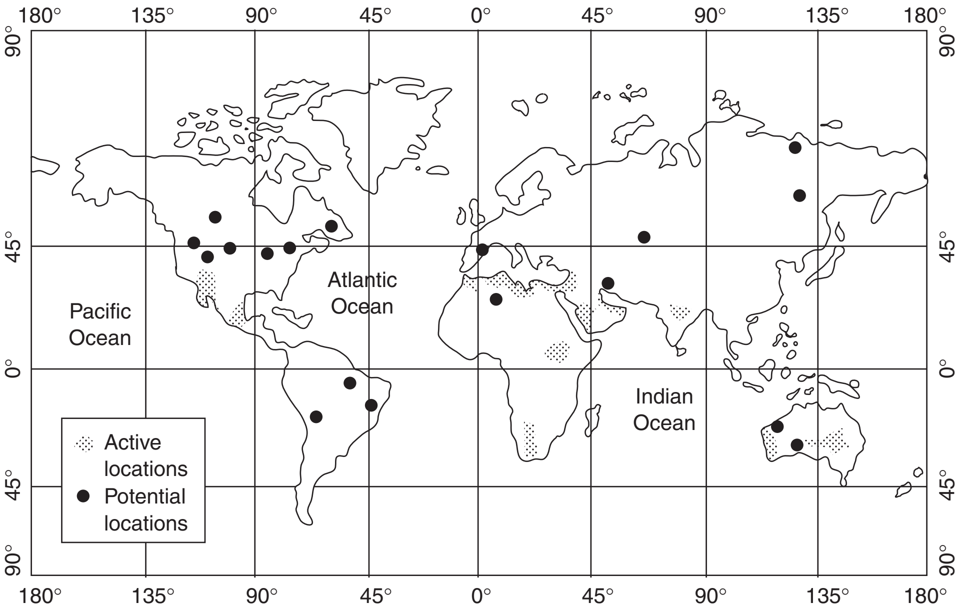

Figure 136 Map of the world showing active and potential sabkha locations (Al-Amoudi 1994)

lagoon. The intertidal zone can be divided into an upper and lower intertidal facies. However, the lower intertidal facies may be the same as the lagoonal facies as described by a number of workers. This facies is dominated by an algal peat composed of the bioturbated remnants of an algal mat. The upper intertidal facies is usually a laminated algal mat often cross-cut by desiccation cracks containing lenticular gypsum crystals. Aragonite, magnesite and dolomite may locally cement the surface sediments (Butler *et al*. 1982). The lower supratidal belt, which is flooded once or twice a month, is characterized by gypsum up to 30 cm thick. Diagenetic nodular anhydrite occurs in the deposits of the middle supratidal zone and there is often a surface crust of ephemeral halite. Such deposits are flooded on less than monthly intervals. In the upper supratidal zone, flooded once every four to five years, the gypsum has been replaced by coalesced nodules of anhydrite.

Sabkhas are broad coastal supratidal and intertidal flats developed along the margins of arid landmasses. Sediments that accumulate on sabkhas include: (1) siliciclastic detritus sediments that are eroded from adjacent land and washed onto the sabkha; (2) offshore deposits of sand and mud that periodic storms wash up and onto the sabkha; and (3) the indigenous sediments of the sabkha itself.

Much of the evaporite sediment produced in sabkhas precipitates as saline ground water seeps into and out of the sabkha (Figure 137). Much of this ground water is seawater that is continually recharged beneath the sabkha, but ground water from the adjacent landmass can also feed the system. Groundwater circulation is driven by capillary action and evaporative pumping. Intermittent flooding by the sea also occurs, and beach ridges can trap a reservoir of additional seawater.

Typical sabkha evaporite minerals are anhydrite, gypsum and dolomite. Much of the anhydrite occurs as irregularly shaped lumps or nodules. These nodules replace altered gypsum crystals originally formed within layers of interbedded carbonate mud or shale. The term chickenwire structure is used to refer to this mixture of elongate, irregular clumps of anhydrite separating thin stringers of carbonate and/or siliciclastic mud (Plate 105a,b). This structure is particularly common in sabkha evaporites.

Cyclicity is common in sabkha evaporite sequences. As deposition proceeds, sabkha deposits

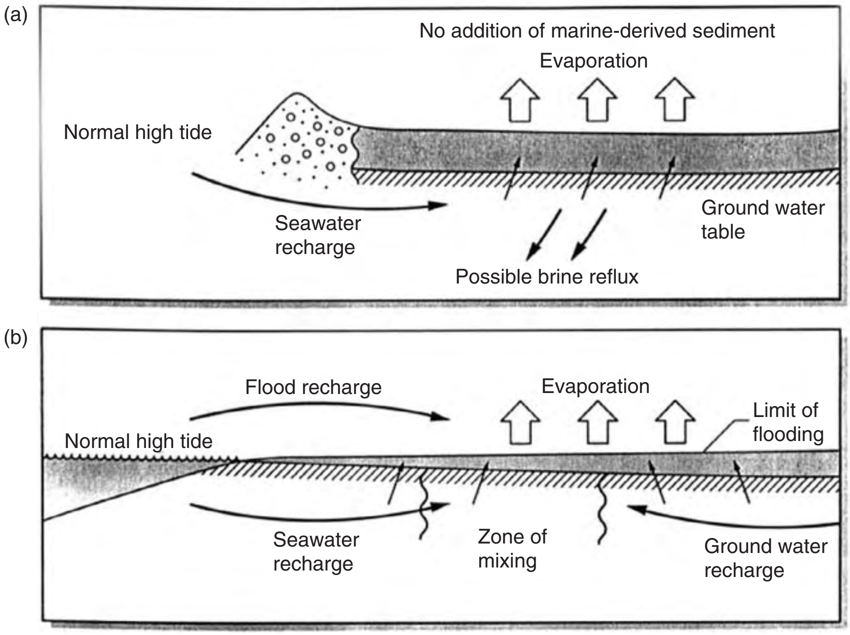

Figure 137 Sabkhas receive water from a variety of sources: (a) sabkha with seawater recharged through the subsurface and with relatively little groundwater influx; (b) sabkha groundwaters are recharged by a mixture of seawater and groundwater, plus seawater flooding from major storms (from Walker 1984)

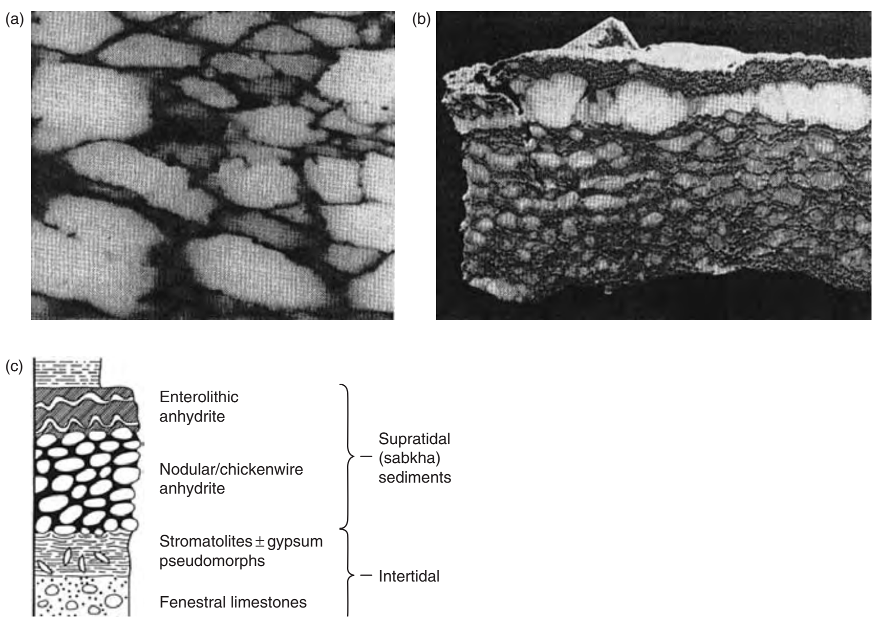

Plate 105 Sabkhas produce a number of distinctive structures: (a) mosaic anhydrite (chickenwire structure) commonly formed when anhydrite nodules coalesce, shown at actual size; (b) nodular gypsum is common just below the surface of the sabkha; (c) typical vertical cycle of sabkha sediments. Such cycles range from several metres to several tens of metres in thickness (after Tucker 1981)

naturally prograde oceanward and eventually lie upon intertidal sediments (STROMATOLITES, gypsum, fenestral birdseye pelleted carbonate mud). These in turn rest on oolitic and bioclastic subtidal carbonate rocks of the subtidal zone (Plate 105c).

The Arabian Gulf coastal sabkhas occur on the southern margin of the Gulf with the best-studied sequences being situated south-east of Abu Dhabi City, where they are now partially covered by urban and industrial development. If an idealized landward transect is traced, it passes in order through offshore open-marine skeletal sediments, belts of oolitic grainstones, belts of lagoonal and/or barrier sequences, and then crosses intertidal algal mats to terminate in a supratidal sabkha sequence (Butler *et al*. 1982). If the intertidal algal mats are included, the sabkhas constitute a zone 10–15 km wide, with an along-strike continuity of more than 150 km. The surface transect from the lagoonal sands and muds up onto the flat plain of sabkha first crosses the dark, black to grey, flat-laminated, algal mats of the intertidal zone (Figure 138).

In some arid areas the surface of the sabkha is encrusted by a veneer of salt and scattered discoidal crystals of gypsum. Dust and sand storms occur through most of the year, which results in the sabkha plain being covered by a layer of drifting quartz sand.

The landward boundary of the supratidal area is characterized by a zone of vegetation and is dominated by the halophyte *Halocnemon Strobilaceum*. The vegetation on the sabkha surface acts as a trap for the moving sand. Most of this drifting sand is usually washed off during storm tides and redistributed on the sabkha surface.

Inland sabkhas develop where water flowing in wadis intermittently floods low-lying depressions (see PAN) to leave behind damp, salt-encrusted sediments. They are also found in depressions where, for one reason or another, the water table reaches the surface.

In inland sabkhas, the salt crust forms as the result of the concentration of salts caused by evaporation of the water. Gypsum crystals are common in the sediments of inland sabkhas. Algae are known, but the algal mats so commonly associated with coastal sabkhas have not been recognized. A coastal sabkha, on the other hand, is characterized by marine flooding and evaporitic conditions. It is a diagenetic environment whose sediments are of continental and adjacent marine origin.

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SEE ALSO: deflation

ADEEBA E. AL-HURBAN

SACKUNG

A German term describing a type of slope-sagging, gravitational lateral spreading or deep-seated gravitational deformation in mountainous alpine landscapes. Sackungen (plural) display rounded morphology, commonly trend parallel to the contours of the slope, and form a characteristic ridge-top trench in the adjacent valley. They typically display a bulge at the toe

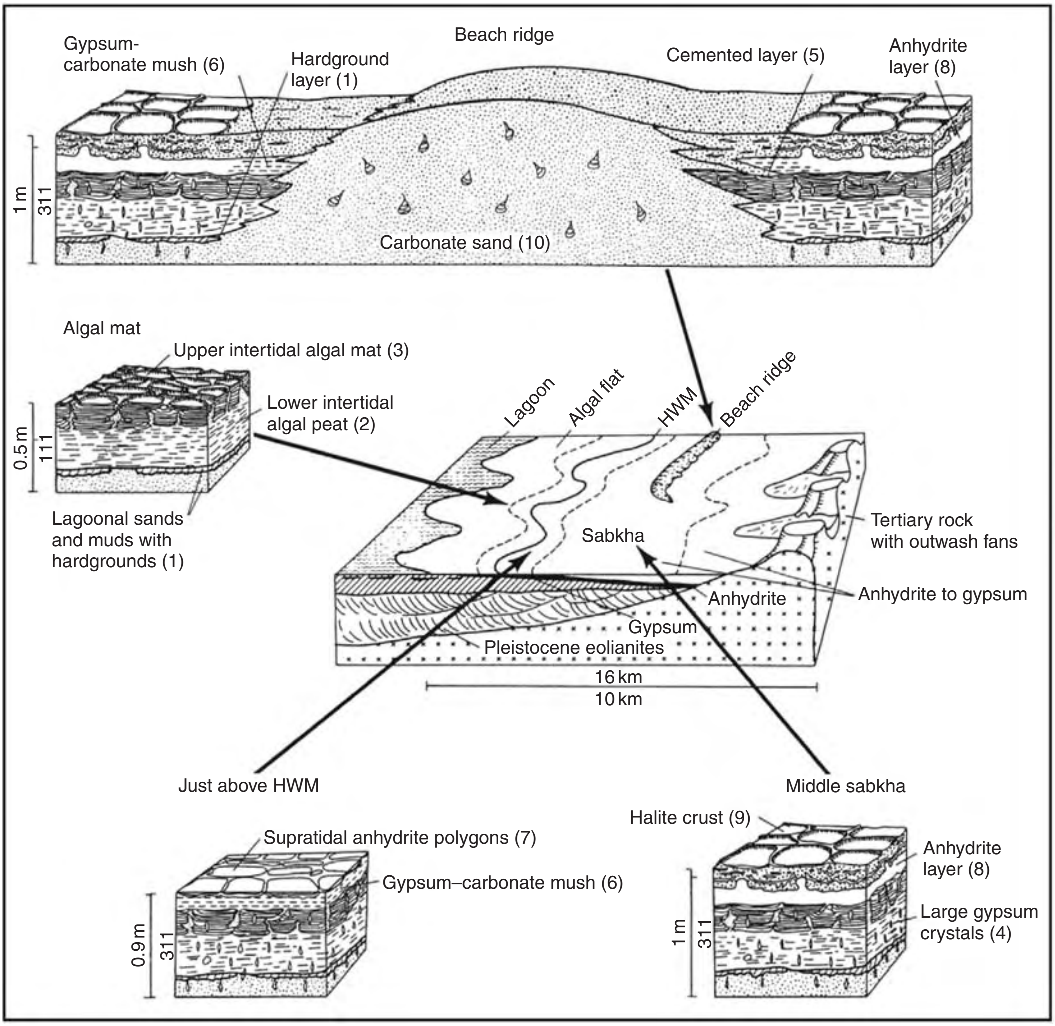

Figure 138 Schematic block diagrams showing sediment and evaporite distribution in Abu Dhabi sabkha, Arabian Gulf. All peripheral diagrams are keyed to central block of sabkha. HWM = high-water mark. (1) Lagoonal carbonate sands and/or muds with carbonate hardgrounds; (2) vaguely laminated lower tidal-flat carbonate-rich algal peat; (3) upper tidal-flat algal mat formed into polygons; (4) large gypsum crystals (many lenticular); (5) cemented carbonate layer; (6) high tidal-flat to supratidal mush of gypsum and carbonate; (7) supratidal anhydrite polygons with wind-blown carbonate and quartz; (8) anhydrite layer replacing gypsum mush and forming diapiric structures; (9) halite crust, formed into compressional polygons; (10) deflated beach ridge of cerithid coquina and carbonate sand. (After Butler *et al.* 1982)

of the slope, known as the ‘Talzuschub’, as well as tensile or normal faults near their crest, termed ‘Bergzerreißung’ (Zischinski 1969). Rates of material creep vary significantly (1 mm to several metres), with variations in activity corresponding to changes in precipitation and water table (increased creep activity with higher precipitation). Sackungen are indicative of gravitational spreading (Varnes *et al.* 1989), resulting from low mass strength in the underlying material (a product of high density jointing and faulting) to a substantial depth.

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SEE ALSO: mass movement

STEVE WARD

SALCRETE

A salty surface crust, primarily composed of sodium chloride, that cements a sand surface as a result of evaporation of moisture and the consequent chemical concentration of dissolved material. The term, which was coined by Yasso (1966), has been mainly used to describe crusts developed through evaporation of sea spray on beaches (e.g. Pye 1980).

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A.S. GOUDIE

SALT (EVAPORITE) KARST

Evaporites, including common salt (halite) and calcium sulphate (gypsum and anhydrite) are the most soluble of common rocks (see GYPSUM KARST). They are also widespread, being found, for example, in 32 of the 48 contiguous states of the USA (Johnson 1997) in rocks of every geologic system from the Precambrian through to the Quaternary. Karst-creating evaporites are also extensively developed in Canada (Ford 1997), and have been described from many parts of Europe, including the Ripon area of England and the Betic Cordillera and Ebro basins of Spain. Salt dome structures are often affected by solutional processes (see SALT RELATED LANDFORMS).

Evaporite outcrops display a full array of solutional features, including subsidence depressions, collapse breccias, sinkholes, vertical shafts and water-filled chimneys (Last 1993; Calaforra and Pulido-Bosch 1999; Gutierrez-Elorza and Gutierrez-Santollala, 1998). Because of the great solubility of evaporites, rates of karst denudation can be high, and even under the dry conditions of the arid parts of Israel, can approach 500–750 mm 1,000 yr^{-1} (Frumkin 1994). Salt karst processes present a range of engineering problems and hazards (Paukstys *et al*. 1999) and these can be exacerbated as a result of human activities, including mining, ground water abstraction and other hydrological modifications.

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A.S. GOUDIE

SALT HEAVE OR HALOTURBATION

A cause of damage to engineering structures in desert areas as a result of the presence of soluble salts (including sodium chloride, magnesium sulphate and sodium sulphate). The process is akin to frost heave, in that the hydration and crystallization of salts plays a major role, but it is also akin to needle-ice (pipkrake) formation in that salt whiskers may grow vertically. The problem is especially severe when saline ground water approaches the ground surface. Possible techniques to deal with it have been developed (Horta 1985), including brooming, embankments, barriers and the use of thick, impervious surfacings. In the case of some gypsum areas, volume changes associated with solution and recrystallization can produce what are termed 'mega-tumuli' and dome-shaped hills (Ferrarese *et al*. 2002).

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A.S. GOUDIE

SALT RELATED LANDFORMS

Evaporites, including halite (sodium chloride) are widespread both geographically and in the geological record. As much as one-quarter of the world's continental areas may be underlain by evaporites of one age or another. When evaporite beds are thick they can have many important geomorphological consequences: the development of diapiric structures (including salt domes), the production of folds and faults (see FAULT AND FAULT SCARP), tectonic uplift and associated drainage modification, the creep of salt as 'salt glaciers', and widespread solution, subsidence and karst formation. Salt is important also as a cause of weathering (see SALT WEATHERING).

Because it has a low density and low rheidity (the ease at which it flows as a viscous solid), salt flows readily under burial conditions when the surrounding sediment is still undeformed. Flow rates can increase in the presence of water (as brine) and at temperatures in excess of around 245 °C. The flowage of salt can transform relatively tabular evaporite bodies into a wide variety of structures that tend to evolve from concordant, low-amplitude features through to discordant high-amplitude intrusions (DIAPIRS), and thence to extrusions. The immature concordant structures include salt anticlines (which have approximately symmetrical cross sections, planar bases and arched roofs), salt rollers (which are also ridge-like, but are asymmetric with a faulted scarp) and salt pillows (periclinal, subsurface domes). Sediment covers tend to be thin over the crests of such structures, but the area above the pillow tends to be a topographic high surrounded by a topographic low or primary rim syncline. These form simultaneously with the accumulation of salt in the area of ongoing uplift, and result from the downwarping of the overlying strata into the space vacated by the salt flowing into the growing structure.

High amplitude diapiric intrusions, involving piercement, with uplift characteristically at 2–4 mm per year, develop in the next stage of structure growth. Among the forms described are salt walls. These are elongated like salt anticlines but are intrusive and of much greater amplitude. They tend to be 4 to 5 km in breadth, have a length of over 120 km and are generally 8 to 10 km apart. Another intrusive form is the salt stock. These vary in shape from squat to columnar and can be conical or barrel-shaped. The round varieties are 2 to 8 km in diametre in their upper parts, and in many places they are linked together in parallel-striking straight or winding lines that have been likened to strings of pearls. When the whole or part of the shallower portion of a diapir extends laterally beyond the cross-sectional area of the diapir roots, an overhang develops, producing balloon or mushroom shapes (see Jackson *et al*. 1990).

The final stage of salt structure evolution is the postdiapir stage (Warren 1989), during which the salt supply is dwindling as the volume of the salt mass decreases. Meteoric processes become important and solutional loss occurs. Large collapse depressions can form.

If the rate of diapiric growth is greater than the rate of salt solution, salt will be extruded at the ground surface. This may tend to be favoured in arid areas where low precipitation values cause low rates of dissolution. Because of its mechanical properties such extruded salt may begin to flow, and such flows are called 'salt glaciers'. These are termed 'namakiers' (from *namak* – the Farsi word for salt – and glacier) (Talbot 1979).

Some of the Zagros Mountains' salt glaciers in Iran are large. The example at Kuh-e-Namak is 2,000 m long, 3,500 m wide and up to 50 m thick, but perhaps the biggest example is Kuh-e-Gach, which is also 50 m thick, but attains a width of around 4,700 m and has an area of around 23.5 km^2. Their speed of movement is less than that of true glaciers, and the average speed is only a few metres per annum with movement tending to occur after rainfall events. They display complex folds where they flow over bedrock irregularities, and at their distal ends they may feather out in a mass of unbedded detritus analogous to a terminal moraine. Salt glaciers are also subject to retreat if the balance of wastage by solution should exceed that of outward flow of salt from the diapir, and this explains the presence of isolated exotic blocks, analogous to erratics, some kilometres beyond the plugs themselves.

The growth of salt structures can create characteristic drainage patterns and slope forms. A model of this has been proposed by Berger and Aghassy (1982) who envisage three development phases. In the 'positive relief stage' there is a central dome on which radial drainage by outbound consequent streams is dominant, and these dissect long isoclinal slopes. In the 'breached stage' the initial topographic high in the centre of the dome becomes lowered, an inversion of topography begins to occur and a major depression develops at the dome's centre. Inward-facing scarp slopes and inbound obsequents develop. In the 'obliterative stage' the inbound obsequents expand headwards and capture much of the consequent outward-bound drainage. Sedimentation, floodplain development and marsh formation occur in the core.

Finally, a large number of the world's 'fold and thrust' belts have developed over evaporites. Examples include the Jura, the Pyrenees, the Franklin Mountains of north-west Canada, the Canadian arctic fold mountains, the Salt Range of Pakistan, the Zagros of Iran, the Sierra Madre Oriental of Mexico, the Cordillera Oriental of Colombia, the Atlas of Tunisia and Algeria, the South Urals, the mountains of the Tadjik Republic, and the anticlinal province of the Amadeus basin in Australia. The presence of salt encourages what is termed 'thin-skinned deformation'.

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A.S. GOUDIE

SALT WEATHERING

The weathering of rocks and building materials by salt. This is an important group of processes particularly in deserts, on coasts and in cities. Salt weathering probably plays an important role in the formation of PANs, TAFONI, STONE PAVEMENTs, SHORE PLATFORMs, rock flour and split cobbles. The build up of salt in rocks can also cause SALT HEAVE OR HALOTURBATION and SLAKING. Conventionally, salt weathering can be divided into mechanical and chemical mechanisms (Goudie and Viles 1997).

The most cited cause of salt weathering is generally the process of salt crystal growth from solutions in rock pores and cracks (Evans 1970). Various mechanisms can cause crystal growth to occur. For example, some salts rapidly decrease in solubility as temperatures fall. This is particularly true of sodium sulphate, sodium carbonate, magnesium sulphate and sodium nitrate. Thus nocturnal cooling could cause salt crystallization to occur. Such a crystallization of a salt solution on a temperature fall affects a much larger volume of salt per unit time than crystallization induced by evaporation, which is a more gradual process.

Nevertheless, evaporation helps to create saturated solutions from which crystallization can occur, and when this happens highly soluble salts will produce large volumes of crystals. In this context it is important to note that of the common salts, gypsum is very much less soluble than many of the others, and that less crystalline material will be available in a given volume of solution to cause rock disruption.

A salt's crystal habit may also affect its power to cause rock breakdown. For instance, the needle-shaped habit of sodium sulphate crystals (mirabilite) might tend to increase their disruptive capability.

Air humidity is an important control of the effectiveness of salt crystallization, for a salt can crystallize only when the ambient relative humidity is lower than the equilibrium relative humidity of the saturated salt solution. If that is the case on a rock surface then the salt will crystallize and cause decay. The equilibrium relative humidities of different salts vary considerably, and those with low values will be prone to dissolution in humid air (Plate 106). The equilibrium relative humidities of hydrated sodium carbonate and sodium sulphate are high, whereas those of sodium chloride, sodium nitrate and calcium chloride are relatively lower.

Another type of salt weathering process is salt hydration. Certain common salts hydrate and dehydrate relatively easily in response to changes in temperature and humidity. As a change of phase takes place to the hydrated form, water is absorbed. This increases the volume of the salt and thus develops pressure against pore walls. Sodium carbonate and sodium sulphate both undergo a volume change in excess of 300 per cent as they hydrate.

For some salts a change of phase may occur at the sorts of temperatures encountered widely in nature; sodium sulphate's transition temperature is 32.4 °C for a pure solution, and falls to 17.9 °C in a NaC1 saturated environment. Moreover for some salts the transition may be rapid. At 39 °C the transition from thenardite (Na_2SO_4) to mirabilite ($Na_2SO_4 \cdot 10H_2O$) may take no longer than twenty minutes (Mortensen 1933).

Winkler and Wilhelm (1970) have calculated the hydration pressures of some important common salts at different temperatures and relative humidities, and find that the greatest hydration pressures (maximum value 2,190 atm at 0 °C and 100 per cent relative humidity) occur when anhydrite changes into gypsum. This exceeds the crystallization pressure of ice at −22 °C, and is in excess of the pressure required to exceed the tensile strength of rocks.

The number of occasions upon which rock surface temperatures cycle across the temperature thresholds associated with the change of phase is probably substantial in many desert areas. If one assumes that an air temperature of *c*.17 °C translates into a rock surface temperature of *c*.32 °C (the transition temperature for sodium carbonate and sodium sulphate) then that value is crossed daily between 5 and 9 months of the year depending on the desert station selected. In other words, there may typically be around 150 to 270 days in the year in which rock temperature conditions are favourable to the salt hydration mechanism of rock decay.

A third possible mechanism of rock disruption through salt action has been proposed by Cooke and Smalley (1968), who argue that disruption of rock may occur because certain salts have higher coefficients of expansion than do the minerals of the rocks in whose pores they occur. Halite expands by around 0.9 per cent between 0 and 100 °C, whereas the volume expansion of quartz and granites is generally about one-third of that value. Gypsum and sodium nitrate are other common salts that have a relatively greater expansion potential compared to rock minerals.

It is difficult to assess the actual importance of this process, and while some early experimental simulations (e.g. Goudie 1974) suggested that it was not very effective, much more work is required on this mechanism before its potential can be dismissed.

In addition to these three main categories of mechanical effects, salt can cause chemical

Plate 106 Damaged water pipelines in central Namibia. They proved to be unable to withstand the corrosive conditions associated with the foggy, salty environment of the Namib, and had to be replaced after only a few years in service

Plate 107 Raised beach cobbles being split by halite and nitrate in the Atacama Desert near Iquique in Chile. (Beer can for scale)

weathering. Some saline solutions can have elevated pH levels. Why this is significant is that silica mobility tends to be greatly increased at pH values greater than 9. Indeed, according to various studies, silica solubility increases exponentially above pH9. The presence of sodium chloride may also effect the degree and velocity of quartz solution. At higher NaCl concentrations quartz solubility and the reaction velocity both increase. The growth of salt crystals may be able to cause pressure solution of silicate grains in rocks, for silica solubility increases as pressure is applied to silicate grains. This is a mechanism that has been identified as important in areas where calcite crystals grow, as for example in areas of calcrete formation.

Schiavon *et al.* (1995) have found petrographic evidence from granites in urban atmospheres that suggests chemical reactions occur between granite minerals and weathering solutions responsible for the precipitation of gypsum. They found feldspar minerals that were partially or totally replaced by sulphate crystalline salts while still retaining their primary mineral outline and texture.

The attraction of moisture into the pores of rocks or concrete by hygroscopic salts (e.g. sodium chloride) can accelerate the operation of chemical weathering processes and of frost action (MacInnis and Whiting 1979; Plate 108) and the disruptive action of moisture trapped in rock capillaries is well known.

Plate 108 Concrete buildings in Ras Al Khaimah, United Arab Emirates, illustrating salt weathering caused by migration of salt solutions up the pillars by the wick effect. The process is exacerbated by high ground water levels associated with the spread of irrigation

Salt can have a deleterious impact on iron and concrete. Many engineering structures are made of concrete containing iron reinforcements. The formation of the corrosion products of iron (i.e. rust) causes a volume expansion to occur. If one assumes that the prime composition of such corrosion products is $Fe(OH)_3$, then the volume increase over the uncorroded iron can be fourfold. Thus when rust is formed on the iron reinforcements, pressure is exerted on the surrounding concrete. This may cause the concrete cover over the reinforcements to crack, which in turn permits the ingress of oxygen and moisture which then aggravates the corrosion process. In due course, spalling of concrete takes place, the reinforcements become progressively weaker, and the whole structure may suffer deterioration.

Rates of corrosion are accelerated by chloride ions which may occur in a concrete because of the use of contaminated aggregates or because of penetration from a saline environment. However, the electro-chemical corrosion of metals can also be produced by sulphates for there are often sulphate-reducing bacteria in a saline soil containing sulphates, which can cause strong corrosion to metals.

Sulphates can cause severe damage to, and even complete deterioration of, Portland cement concrete. Although there is controversy as to the exact mechanism of sulphate attack (Cabrera and Plowman 1988), it is generally accepted that the sulphates react with the alumina-bearing phases of the hydrated cement to give a high sulphate form of calcium aluminate known as ettringite. Magnesium sulphate is particularly aggressive because in addition to reacting with the aluminate and calcium hydroxide as do the other sulphates, it decomposes the hydrated calcium silicates and, by continued action, also decomposes calcium sulphoaluminate. The formation of ettringite involves an increase in the volume of the reacting solids, a pressure build up, expansion and, in the most severe cases, cracking and deterioration. The volume change on formation of ettingite is very large, and is even greater than that produced by the hydration of sodium sulphate.

Another mineral formed by sulphates coming into contact with cement is Thaumasite. This causes both expansion and softening of cement and has been seen as a cause of disintegration of rendered brick work and of concrete lining in tunnels. Building materials may also be damaged by SULPHATION, which creates disruptive gypsum on rock surfaces.

Finally, one controversial issue in weathering studies is whether the presence of salts accelerates the rate at which frost action operates, and, if it does, why this should be. Some laboratory studies (see Williams and Robinson 1991) have indicated that some rocks disintegrate more rapidly when they are frozen after soaking in salt solution rather than in pure water, and various studies have shown that de-icing salts can promote the breakdown of concrete by freeze–thaw. However, in some laboratory simulations salts may reduce or even prevent frost weathering.

This whole issue is important in terms of understanding weathering in high latitude coastal situations (e.g. on shore platforms), to understand the effects of de-icing salts on road surfaces and engineering structures like bridges, and also because salts generated by acid rain could conceivably enhance frost action. Williams and Robinson (1991), in a useful review, have looked at some of the mechanisms that might explain why under some conditions salts could accelerate frost weathering.

Identification of areas where salt weathering is a hazard to engineering structures is an important task for applied geomorphologists and a detailed discussion of this in the context of ground water conditions in arid regions is provided by Cooke *et al*. (1982). As irrigation spreads, leading to ground water rise, areas subject to salt attack may increase. There are already many examples of the acceleration of salt weathering by human activities causing the decay of cultural treasures, such as the Sphinx in Cairo, Petra in Jordan and Monhenjo-Daro in Pakistan.

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A.S. GOUDIE

SALTATION

Derived from the Latin *saltare*, saltation, coined by Gilbert (1914), refers to the hopping, jumping or leaping of sediment grains transported by a fluid, whether wind or water (Bagnold 1956). The grains are launched from their bed in a high angle trajectory by lift forces. In air this trajectory is between 20° and 40° downwind. Grains then accelerate in the flow direction as a result of fluid drag. Then, as a result of gravitational and drag forces they fall back to the bed on a more gentle trajectory (10°–15° in air). The entrainment of a grain in water is due to direct fluid lift and drag forces. In air, entrainment can also result from grain collision (Plate 109). The force at which a

Plate 109 Saltating sand grains leaping to a metre or so above the surface during a sand storm at Budha Pushkar in the Rajasthan Desert, India

grain is set in motion (the entrainment threshold), as well as the height and length that it attains in its subsequent jump, is proportional to the shear velocity of the fluid and to grain size. In air, saltation hop-lengths are about 12 to 15 times the height of bounce. Because a sand grain is only two to three times denser than water, the inertia of the rising grain only carries it to a height of about three grain diameters.

In air, on the other hand, saltating grains rise higher, sometimes to several metres, especially after bouncing impacts on rock or pebble surfaces. It is the mode of travel of most wind-blown sand. In addition, the impact of saltating grains can cause a slow downwind movement of grains by surface creep. There is a transition between surface creep, where grains do not lose contact with the bed, and saltation. Some grains may make very short trajectory paths where they barely leave the bed. This process is termed reptation (Rice *et al*. 1995).

Saltation also affects snow and contributes to the development of avalanches (Sato *et al*. 2001).

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A.S. GOUDIE

SALTMARSH

Coastal and estuarine saltmarshes are depositional landforms situated within the upper part of the intertidal zone. Implicit in their definition is the presence of halophytic (salt-tolerant) vegetation. This distinguishes saltmarshes from tidal flats, from which they commonly develop. Inland areas characterized by alkaline soils may also develop a similar vegetation cover. Associated with aridity, these areas are more commonly referred to as salt flat, salt steppe or salt desert and are not considered further in this entry.

Saltmarshes have a wide geographic distribution along temperate and high latitude coasts, but are replaced by MANGROVES SWAMPS in the tropics. Locally, their occurrence is restricted to low wave energy environments which favour the accumulation of fine, generally muddy, sediments. The morphology of most saltmarshes is characterized by a seaward sloping vegetated platform, dissected by networks of tidal channels (also termed ‘creeks’ or ‘sloughs’). The subtle topography of the marsh surface is often associated with a zonation in plant productivity and/or species composition, which results from complex linkages between factors such as salinity stress, nutrient availability, frequency of flooding (a function of elevation) and plant competition.

At a global scale, major differences in marsh character result from the interaction between ecological, climatic, edaphic and hydrographic influences. These are mediated at a regional scale by the nature and abundance of fine sediments and by the range of depositional settings afforded by particular coastal configurations. Saltmarshes are characterized by a particularly strong interplay of physical, biological and geochemical processes and, accordingly, have long been the subject of considerable scientific interest.

Early scientific studies were concerned mainly with the processes of halophyte colonization under the influence of various environmental factors (notably elevation, as the crucial factor determining the frequency of tidal inundation, salinity and soil aeration) and the importance of vegetation (especially dense swards of tall marsh grasses) in the trapping and binding of fine sediment. Research in Europe and North America emphasized the role of coastal halophytes as ‘land-building agents’, leading to a model of saltmarsh morphological development under the influence of an autogenic plant succession (Chapman 1974). Geographical variations in the plant succession provide one basis for the classification of marsh types (e.g. Adam 1990). Subsequent ecosystem studies have shown that saltmarshes are sites of high biological productivity, the cycling of which is governed by complex vegetation–substrate–fauna interactions and by tidal exchanges of water, sediments and nutrients with the marine environment. The so-called ‘outwelling hypothesis’ (Odum 1971) attributed much of the enhanced biological productivity of coastal waters to exports of organic material and nutrients from intertidal marshes and subtidal seagrass beds. Empirical studies provide only partial support for such nutrient exports, and highlight the importance of geomorphological controls on marsh configuration and processes, operating over a range of scales (Nixon 1980).

Geomorphologists have tended to assign a secondary, more opportunistic, role to the colonization of intertidal surfaces by halophytic vegetation. From this perspective, marsh ecological characteristics are viewed as contingent upon the provision of viable substrates for plant colonization. Four main sets of physical factors are implicated: sediment supply; tidal regime; wave climate; and relative sea-level movement (Allen and Pye 1992).

The configuration and extent of coastal margins exert a first-order control on both sediment supply and the 'accommodation space' for saltmarsh development. Frey and Basan (1985), for example, draw attention to physiographic (see PHYSIOGRAPHY) contrasts between the Pacific and Atlantic coasts of North America. On the tectonically active and sediment-deficient Pacific coast, saltmarshes are fragmented and are restricted to narrow fringes around protected embayments, estuaries (see ESTUARY) and (in the north) FJORDs. The more extensive Atlantic coastal plain marshes are continuous over larger areas. Regional variation in the width of the CONTINENTAL SHELF also exerts a control on tidal range which, in turn, defines a zone within which saltmarsh sedimentation can occur.

Saltmarshes are important sinks for fine sediment and play an important role in sediment exchange between estuarine and coastal waters. The nature of saltmarsh sediments varies markedly between *allochthonous* systems, characterized by the deposition of externally derived inorganic sediments, and *autochthonous* systems, dominated by the accumulation of internally produced organic material (Dijkema 1987). The relative importance of inorganic and organic matter accumulation determines the nature of saltmarsh morphodynamic development as well as the ability of both physical and ecological components of the system to adjust to changes in environmental boundary conditions (notably tidal range and mean sea level; French 1994).

Tidal hydrodynamic processes are especially important in controlling the rate and pattern of sedimentation within allochthonous marshes, some of the best-developed examples of which occur under large tidal ranges (e.g. Davidson-Arnott *et al*. 2002). In these systems, the introduction of muddy sediment is controlled by both elevation (which determines the frequency and duration of flooding, or 'hydroperiod') and by proximity to the tidal channels through which most of the tidal water movement occurs (French and Spencer 1993). The feedback between elevation, tidal flooding and sedimentation is a key determinant of long-term marsh morphodynamics (Allen 2000; Friedrichs and Perry 2001). Thus, newly formed marshes typically exhibit rapid rates of vertical sedimentation, whilst the sedimentation is much slower in older marshes, which are higher in elevation and therefore less frequently inundated. Non-tidal inundation, such as that resulting from occasional storm surge events, is of greater importance in introducing sediment to marshes with a very small tidal range.

Wave climate exerts an important local control on horizontal marsh extent, even in otherwise sheltered embayments and estuaries, where small geographical variations in fetch may give rise to significant differences in the character and energetics of the intertidal zone. Wave-induced stresses determine the viability of vegetation establishment, although the influence of waves on the stability of the underlying substrate seems to be of more importance than the mechanical strength of the plants. Wave climate also determines the morphology of the saltmarsh to tidal flat transition. Under moderate wave energies this may take the form of an erosional 'micro-cliff'.

The formation and development of saltmarshes is also related to sea-level change. SEA LEVEL provides a moving boundary condition which, along with tidal range, determines the vertical extent of saltmarsh growth. Modern saltmarshes formed in response to Holocene sea-level rise, and minor oscillations in sea level appear to have been associated with distinct episodes of saltmarsh expansion in areas conducive to fine sediment accumulation. This 'depositional paradigm' (Stevenson *et al*. 1986) has been challenged by the discovery of sedimentary deficits within subsiding deltaic marshes. This has focused attention on the extent to which saltmarshes are able to accumulate sufficient material to keep pace with the forecast rates of sea-level rise under global warming scenarios. In general, contemporary rates of sedimentation within non-deltaic marshes in both North America and Europe exceed present rates of sea-level rise. Furthermore, marsh elevations can adjust to higher rates of sea-level rise through the increased sedimentation which accompanies more frequent inundation (French 1994). This adjustment does, of course, depend on sediment supply. The effect of increased water levels on vegetation and soils is also important, especially in autochthonous marshes with limited inputs of inorganic sediment.

Saltmarshes in many parts of the world have experienced high rates of historical loss through reclamation and destructive industrial uses (such as salt production using evaporation ponds). In addition, large areas of estuarine and open coastal marsh have been lost through recent erosion. This erosion is widely attributed to a combination of contemporary sea-level rise and the presence of seawalls and other structures which restrict any natural landward migration of the intertidal zone. From an ecological perspective, remaining saltmarshes are not only important for the maintenance of estuarine and coastal food chains, but also provide valuable wetland habitats. They also act as a naturally dissipative landform that forms an important element of sustainable coastal defence and flood protection strategies. In particular, a number of studies have shown saltmarsh to be much more effective than unvegetated tidal flats in dissipating wave energy. This function translates into significant engineering cost savings when sea defences are constructed behind a strip of saltmarsh.

These ecological and engineering functions have stimulated efforts to restore saltmarsh, for example through the re-establishment of tidal conditions in reclaimed areas no longer required for agriculture. The success of such schemes has been mixed, and it is now clear that successful engineering of ecological and flood defence functions is crucially dependent upon a sound understanding of the geomorphological processes which act to shape the corresponding natural systems (French and Reed 2001).

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SEE ALSO: mangrove swamp; mud flat and muddy coast; tidal creek; tidal delta

J.R. FRENCH

SAND-BED RIVER

An alluvial river in which the bed material is predominantly sand-sized (0.0625–2 mm). Sediment size is a primary control on river form and process and sand-bed rivers therefore exhibit a number of characteristic process and morphological attributes that distinguish them from channels dominated by other sediment sizes. They are recognized by geomorphologists as a fundamental river type, distinct from GRAVEL-BED RIVERS, in which the bed material is coarser (>2 mm) and less well sorted.

Within the drainage network DOWNSTREAM FINING ensures that sand-sized sediments dominate the sediment load delivered to distal reaches. Gravel-bed rivers in upland and piedmont settings therefore give way to sand-bed channels in the lowlands, though sandy reaches will develop wherever sand is supplied in abundance, for example as a function of local lithology or land use.

The grain characteristics and relative importance of SUSPENDED LOAD and BEDLOAD transport vary in sand-bed rivers with sediment supply and flow characteristics, but suspended grains are typically finer than 0.2 mm and account for a greater proportion of total sediment yield. In the limit case, sand-bed rivers can be thought of as suspension-dominated (e.g. Parker in press). Examination of such channels reveals that the boundary shear stresses generated by modest flows (for example bankfull) are at least one order of magnitude larger than the stresses needed to entrain median sizes. This indicates that in sand-bed rivers mass sediment entrainment and transport occur at discharges well below discharges associated with rare floods, and that modest flows are responsible for maximum cumulative sediment yield. Transport conditions are fundamentally different in bedload-dominated, gravel-bed rivers where flows reach entrainment thresholds relatively infrequently. Sand-bed channels are therefore characterized by excess rather than threshold hydraulic stresses, have more mobile and responsive 'live' beds, and carry sediment loads that are limited by supply issues rather than the flow's competence to move available particle sizes.

When stresses are sufficient to entrain grains but turbulent eddying is insufficient to suspend them, grains move as bedload, primarily by SALTATION. Near their entrainment thresholds grains move randomly over a plane bed, but as flow intensity increases patterns of erosion and deposition become ordered in space and groups of sand grains move together as migrating BEDFORMS. Ripples and then dunes form, but a point is reached where structured transport collapses and dunes are replaced by an upper-regime plane bed. With further increases in flow the water surface may develop waves, beneath which antidunes grow. This bedform sequence correlates with both bedload and suspended sediment transport rates and increases of an order of magnitude have been observed between successive stages in flume experiments. The complex mutual adjustments between bedform generation, macroturbulence (burst cycles), flow resistance, and suspended sediment concentrations that ultimately determine hydraulic characteristics and sediment transport rates are incompletely understood, but sufficient has been learned to allow palaeohydraulic interpretation of bedform traces preserved in the modern and lithified deposits of sand-bed rivers.

Because bedform dimensions far exceed grain dimensions in sand-bed channels, a large proportion of total boundary flow resistance is due to bedforms and grain resistance is relatively unimportant. Drag increases as flat beds (grain roughness only) develop ripples then dunes. However, the transition from dunes back to a plane bed is accompanied by a significant drop in resistance that marks a shift from the so-called lower to upper flow regime and ensures that stage-velocity relations are non-linear. Energy losses rise again if the flow becomes supercritical and the water surface develops breaking waves. An additional control on flow velocity in sand-bed channels is suspended sediment concentration, which reduces resistance by dampening turbulence intensity.

The common transition from gravel to sand bed is often abrupt and accompanied by a distinct break of slope. This has been explained in terms of the gradient required to transport sand and gravel loads and the respective importance of channel capacity and competence in suspension and bedload-dominated systems. Slopes in sand-bed reaches are relatively small, typically between 0.002 and 0.0001 (2–0.1 m per kilometre) and exhibit less downstream adjustment than in gravel-bed reaches where large changes in grain size require consequent changes in flow competence.

Cross-section morphology may be regarded as the more adjustable morphological dimension in sand-bed rivers. An element of distinctiveness is apparent in HYDRAULIC GEOMETRY relations for sand-bed channels, but the multivariate nature of the problem makes widespread generalizations difficult. For example, many sand-bed channels transport significant quantities of fine-grained, cohesive sediment (silts and clays) that are deposited in over-bank positions during floods. This facilitates floodplain accretion and increases bank strength, producing channels that have lower width–depth ratios than gravel-bed rivers carrying similar discharges. However, sand-bed channels that do not carry significant quantities of fines have particularly weak banks and tend to be comparatively wide.

Straight, meandering, braided and anabranching rivers may all have sand or gravel beds. Although the patterns are common to both, Lewin and Brewer (2001) suggest that braid and meander formation are fundamentally different in gravel- and sand-bed channels as a function of differences in excess shear stress and thence sediment mobility and bedform persistence. Certainly a degree of distinctiveness is apparent in the controls on planform type. For example, numerous studies have identified threshold values of discharge and channel slope that discriminate meandering and braided forms, but detailed analysis indicates that the threshold slope for braiding also depends on bed material size, with sand-bed rivers tending to occur on more gentle gradients for a given discharge (Knighton 1998: 211). Similarly, examination of the controls on meander geometry suggest that channels carrying greater suspended loads and more cohesive materials (by extension, suspension-dominated sand-bed channels) tend to be more sinuous and have smaller wavelengths than meandering channels in bedload-dominated systems.

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SEE ALSO: bedform; downstream fining; gravel-bed river; hydraulic geometry; suspended load

STEPHEN RICE

SAND RAMP

Topographically controlled aeolian deposits amalgamated with layers of fluvial, colluvial and talus sediments derived from local mountain sources, and palaeosols representing relatively stable geomorphological periods. Mountains act as barriers to sand transport and sand accumulates on piedmont slopes in their lee and to their windward sides. Where sections are present, multiple periods of sand accumulation and palaeosol development can be identified and such phases can be dated by techniques such as luminescence dating. They can provide a record of past episodes of aeolian activity and stabilization (see, for example, Allchin *et al*. 1978 on the Thar; and Lancaster and Tchakerian 1996 on the Mojave).

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A.S. GOUDIE

SAND SEA AND DUNEFIELD

Collections of dunes are best called by easily understood terms: ‘dunefields’, and, when large, ‘sand seas’. Many geomorphologists use the term ‘erg’ for any large collection of dunes (see DUNE, AEOLIAN), but, apart from the indiscriminate use, there are problems with this term. First, ‘erg’ is an indigenous term from a small part of the northwestern Sahara, where it was used on the topographic maps of over a century ago for anything from a large dune to a very large sand sea. Second, there are many other indigenous terms for bodies of dunes in other parts of the world (even in the northwestern Sahara). And third, better mapping shows there to be two distinct groups of dunefield.

There is a sharp break in the size distribution of collections of dunes at about 32,000 km^2 and the larger group has a sharp peak in size at about 200,000 km^2 (Wilson 1973) (Figure 139). In other words, large dunefields, termed here ‘sand seas’, are a distinct population. A very similar size break also distinguishes ‘seas’ in common English usage from smaller bodies of water. The peaked distribution is a function of tectonics. The gentle folds in the basement rock of the African, Asian and Australian deserts are of this size order (Figure 140). Most dunefields and sand

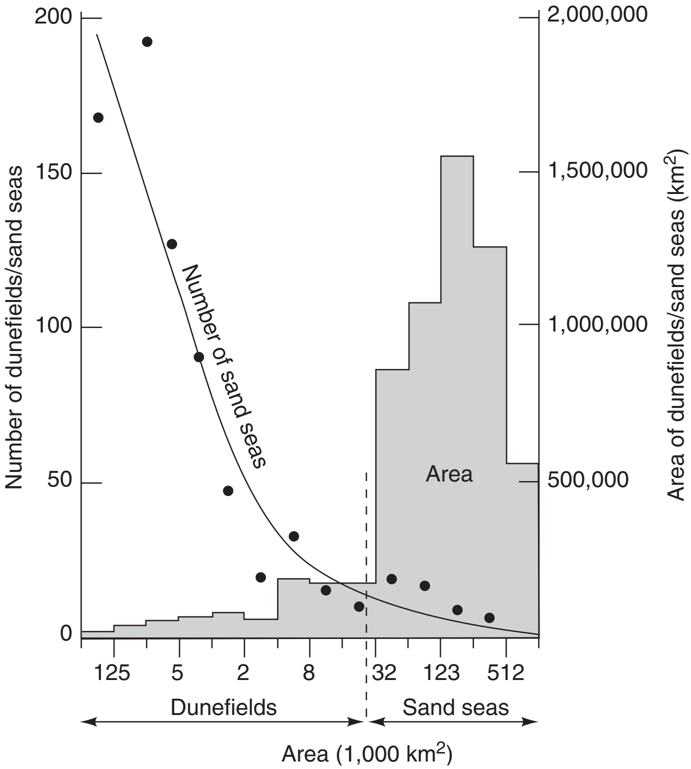

Figure 139 Distribution of the areal extent of aeolian sand bodies, showing the distinction between sand seas and dunefields (after Wilson 1973)

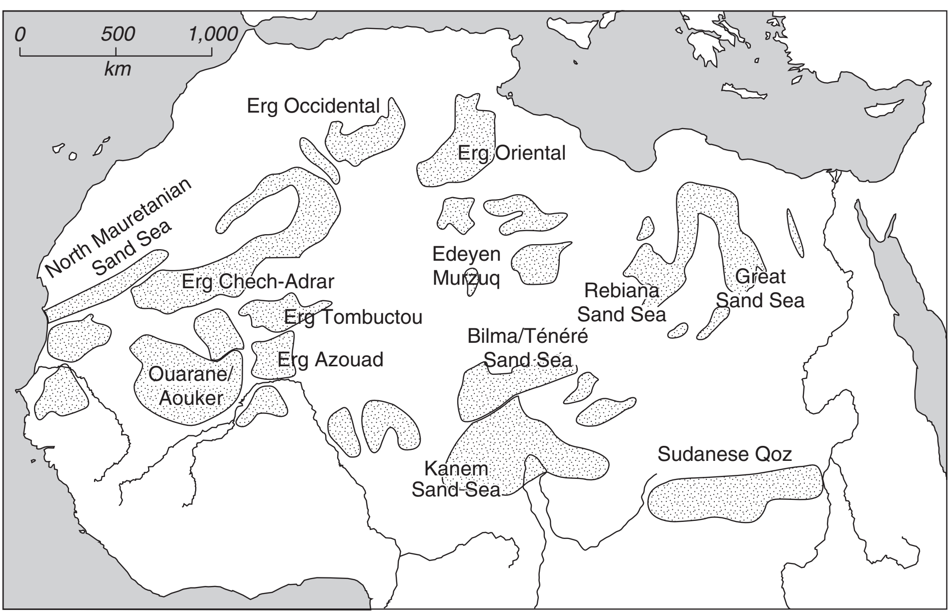

Figure 140 The principal sand seas of the Sahara and Sahel

seas occur in basins or lowlands, and many sand seas nearly fill their basins. Dunefields do so less commonly, although some, as in the Kelso dunefield in California and the Great Sand Dunes in Colorado, fill a large proportion of their small basins.

If the aeolian sand in a dunefield is broken by a patch of desert floor (pavement or bare rock) that is bigger than an interdune, then strictly speaking, this is the edge of the field, but this is a somewhat pedantic restriction. Judgements like this are just one example of the ambiguities in defining the extent of a sand sea (or dunefield), but even with this proviso, the largest active/semi-active sand seas include: the Rub' al Khali in Arabia (about 560,000 km^2) and the Great Eastern Erg in Algeria (about 192,000 km^2). Another somewhat vain definition is of the lower size limit of a dunefield: it could be two dunes.

Sources of sand

Apart from topographic control, the most important control on the distribution of bodies of dunes is proximate supplies of sand. These are of various kinds: weathered bedrock, fluvial deposits, coastal sedimentary cells, and earlier or other dunefields are the most common. Many large dunefields and sand seas derive their sand from a mixture of these sources. In many of the older sand seas, sand has accumulated and been reworked over many cycles. Many dunefields are parts of regional systems of sand movement, one field feeding sand to the next. In the Sahara the movement is generally from north-east to south-west. Regional movements also occur in the Mojave, in the Namib, and probably in Australia. The issue of source, which is best illuminated by studies of mineralogy, has been extensively debated in relation to individual sand seas (Muhs 2003).

Dune assemblages

Many sand seas, unlike most dunefields, contain a variety of dune types. There are several reasons. First, variety is a function of size, for different wind regimes can occur over such large areas, and wind regime is a major determinant of dune type. The northern parts of the Saharan and Arabian sand seas experience frontal winds in winter, and trade winds in summer. Further south in the same sand sea, the influence of the frontal winds fades. In the lower latitude sand seas of southern Africa and Australia, the trade wind systems dominate and dune form is less varied.

Second, variety of dune form is a function of age. Large bodies of sand can only accumulate over many thousands of years, lengths of time that have seen changes in climate. Older dunes may differ in type and orientation from younger ones. Moreover, many older dunes have been subdued by erosion, and have developed soils at times when the climate was wetter. Their gentle topography may be scarred by haphazard reactivation or it may be buried by younger sands to varying degree. This can be seen in the Great Western Erg in Algeria. The northern portion is dominated by ancient, mostly linear, and now subdued dunes, which have developed soils. Further south there are more active and much higher dunes in various formations (transverse, linear and star) (Callot 1988). Another example of how age creates variety is the Wahiba sand sea in Oman. The oldest dunes here have been lithified (see AEOLIANITE), and eroded to an almost level plain. A later generation of large linear dunes partially covers these lithified sands and has moved over them to the north. In the south, these linear dunes have in turn been covered by transverse and network dunes built with new Late Pleistocene sand. These sands have invaded the sand sea from the coast, and are themselves derived from the marine erosion of the ancient lithified aeolianites (Warren 1988). Age, in a sense, allows greater entropy and disorder in dunefields and sand seas.

Third, variety of dune type in large sand seas is a function of the movement of the dune body as a whole. Porter (1986) developed a model for ancient sand seas, now forming aeolian sandstones. It is also to be applicable to some Saharan sand seas. In the model, the finer sand and associated dunes moves forward more quickly than the coarse sand, which remains as a trailing edge of low zibars.

Non-aeolian features

Sand seas and dunefields have some distinctive non-aeolian features. The most striking of these is associated with the blockage of the pre-existing and marginal drainage by the dunes. The drainage and dune water tables feed or have fed many thousands of lakes, some large. These fill up even after the rare storms of today, but were more often full in the wetter periods of the past. These ancient lakes have left deposits ranging from chara limestones to diatomites, sometimes in very thin (5–10 cm thick) drapes on the sand. Round these lakes, in many Saharan and Arabian sand seas, there are signs of human (Palaeolithic and Neolithic) settlement, in the form of tools and middens. Lakes like these are common in the Taklamakan in China and in the Nebraska Sand Sea.

Activity/planetary sand seas

Dunefields and sand seas are in a spectrum from fully active to lithified and largely buried. Some small dunefields are almost wholly active, in the sense that all the dunes are in movement, and most of the surface bare and blowable. But even in the hyper-arid central Sahara and Arabia, many sand seas have portions that are partly stabilized, being remnants of earlier phases bearing remnants of soils. As one moves to the margins of these deserts, as in northern Arabia, or either edge of the Saharan or Australian sand seas, dunes become progressively more stabilized by a cover of vegetation and a developed soil. The climatic limits of aeolian activity are debatable, if only for the reason that a migrating climatic boundary has left a complex legacy of stabilized, reactivated and new dunes. There is no dispute, however, that there are dunefields and sand seas that are now almost wholly inactive, some indeed covered by deep forest. Some of the largest surface sand seas, for example of the proto-Kalahari (2,500,000 km^2), or the Nebraska Sand Hills (57,000 km^2), and the Sudanese *qoz* (about 240,000 km^2) are now almost wholly stabilized. Stabilized dunefields and sand seas also occur in areas in which there are now no deserts, as in northern Canada, the North European Plain, Hungary and northern Tasmania.

Sand seas and dunefields slowly lithify under various influences, and become aeolian sandstones. The sediments of ancient 'ergs' occur from as far back as the Precambrian. The deposits of these sand seas generally show great complexity, as a result of changes in climate and tectonics (Blakey 1988). There are also sand seas and dunefields on Mars, some very similar to terrestrial features (Lancaster and Greeley 1990).

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SEE ALSO: dune, aeolian

ANDREW WARREN

SANDSHEET

An area of predominantly aeolian sand where dunes with slipfaces are generally absent. Sandsheet surfaces can be rippled or unrippled, and range from flat to regularly undulatory to irregular (Kocurek and Nielson 1986). They form in ergs (sand seas) where conditions are not suitable for dunes, or particular factors act to interfere with dune formation. These factors include a high water table, periodic flooding, surface binding or cementation, the presence of vegetation, and a significant coarse grain-size component. These same factors are also effective in promoting sandsheet accumulation where otherwise only sand transport without deposition would occur.

The classic sandsheet is the Selima Sand Sheet of the Libyan Desert. It covers around 120,000 km^2 and is a largely featureless surface of lag gravels and fine sand broken only by widely separated dunefields and giant ripples. Maxwell and Haynes

(2001) have stressed the role of both fluvial deposition and aeolian modification in its development.

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A.S. GOUDIE

SANDSTONE GEOMORPHOLOGY

Landforms developed on sandstone can be arranged in a hierarchical series increasing from the microscopic to the regional scale. Examples of features at these various levels are: (1) etch pits on quartz grains, silica skins; (2) TAFONI, tesselated surfaces, solution runnels; (3) cliffs, domes, towers, arches (see ARCH, NATURAL); (4) CUESTA and scarp assemblages; plateau and canyon assemblages; PSEUDOKARST assemblages, ruiniform assemblages. Explaining features at any of these scales requires an understanding of the variable properties of sandstones.

Sandstones can be most simply defined as clastic rocks in which sand-sized fragments are dominant, but there is considerable variation amongst them, and this variation is of geomorphological significance. The size of the dominant clasts is important, for a fine, silty sandstone responds to erosion differently to a coarse, pebbly one. The proportion of intergranular matrix also is significant, and 'clean' sands or arenites (<15 per cent matrix) need to be distinguished from 'dirty' sands or wackes (>15 per cent matrix). The composition of the grains can vary greatly, for arenites can be divided into quartz (<5 per cent feldspar or rock fragments), lithic (>25 per cent rock fragments), arkose (>25 per cent feldspar) and volcanic (>50 per cent volcanic fragments) subtypes. The amount and composition of intergranular cement, porosity and permeability, and the pattern of bedding and jointing exert important influences on the mechanical and chemical properties of sandstones.

The strength of sandstone depends greatly on its composition. For example, the uniaxial (or unconfined) compressive strength varies from 200 MPa (1 Megapascal = 145 lb/in^2) in a strongly cemented quartz arenite, 20 MPa in a weak sandstone, to 2 MPa in very weakly consolidated sands. Moreover, the strength of sandstones varies greatly with the type of stress applied. Their shearing strength is generally less than half, and their tensile strength only 5 to 15 per cent, of their compressive strength. Furthermore, the strength of saturated sandstone is from 90 to only 10 per cent of the strength of the same rock when dry; the difference between wet and dry strengths increases mainly with the clay content of sandstone. The clay content, together with the degree of cementation, also largely determines the deformability of sandstones. Those which are highly cemented and have little clay tend to fail by brittle fracture, whereas those which are poorly cemented and have substantial clay contents tend to deform elastically before rupturing.

The strength of sandstone mass depends not only on the strength of the intact pieces, but also on their freedom of movement, which, in turn, depends on the spacing, orientation and shearing strength of the discontinuities. Jointing and bedding provide major pathways along which water penetrates sandstones, and thereby influence both weathering of the rock, and the build up of pore-water pressure that may promote failure in the rock.

Strong sandstones generally form major cliffs, but the modes of failure on cliffs, and therefore the morphology of the cliffs themselves, varies with the characteristics of particular rock masses. Because sandstones are weakest when in tension, horizontal projections of rock from cliffs, unless in the form of supported arches, generally extend no more than 10 to 20 m. The patterns of scars caused by tensional rupturing indicate that about 50 per cent of the failures through intact sandstone are generated upwards, about 40 per cent laterally, and, contrary to the predictions of commonly used models, only about 10 per cent downwards. In well-jointed sandstones, the dominant mechanism of failure is the collapse of individual blocks that have been undercut. Once undercutting has penetrated beneath the central third of a block, the block will generally topple outwards, unless held in place by pressure exerted by adjoining blocks.

Undercutting of sandstone is mainly the result of the breakdown of less resistant, underlying rocks. While seepage promotes the plastic failure of clay-rich rocks such as shales, even in these rocks, failure frequently takes the form of brittle fracture along closely spaced beds and joints. Undercutting can also occur along prominent bedding planes in the sandstone itself, and in beds of conglomerate that commonly occur at the base of upward fining cycles of sandstone deposition. Seepage promotes undercutting, especially in very permeable sandstone, but undercutting is essentially the result of the very high concentration of stress generated at the base of a cliff by the weight of the rock above. As the rock at the base is compressed, it deforms laterally, and when the lateral stress pushing outwards into the undercut exceeds the tensional strength, the rock fractures. Tensional stresses generated in this fashion at the base of sandstone cliffs can be greatly augmented by active, or residual, tectonic stresses. For example, the measured horizontal stresses south of Sydney, Australia, are three times greater than the vertical stress. In these conditions, zones of tension, in which joints separating the blocks are opened, extend up the entire cliff face and onto the edge of the adjacent plateau surface. Where sandstones are tilted, joint-bounded blocks may topple or slide down the slope. High and narrow blocks tend to topple; wide and flat blocks tend to slide. Movement of blocks depends also on the steepness of the slope and the frictional resistance at the base of a block.

The form of cliffs thus depends not only on the ROCK MASS STRENGTH of the sandstone itself, but also on the properties of the rocks exposed beneath it. Along the Arnhemland Escarpment, in northern Australia, the Kombolgie Sandstone stands in vertical cliffs where less resistant schists and granites are exposed beneath it, but forms

Plate 110 Cliffs with cavernous weathering in the Nowra Sandstone, south-east Australia

irregular slopes of lower declivity where it occupies the full height of the escarpment. Armouring of weaker rocks by debris from sandstone outcrops also retards undercutting. In the south-west of the USA the presence or absence of thick mantles of talus seems to determine whether slopes are in equilibrium with the properties of the sandstone, or whether they are controlled by foot-slope processes and undercutting.

Extensive masses of sandstone may be incorporated in major failures in weaker, underlying rocks (see TOREVA BLOCK). Major rotational failures in clays on the southern part of the Msak Mallat Escarpment, in central Libya, have incorporated slabs of the Nubian Sandstone in a 3-km wide belt of mounds over 100 m high. Removal of confining stresses resulting from the incision of the Cataract Canyon of the Colorado River has allowed evaporites to slowly deform down the dip, causing fracturing and collapse of the overlying sandstones. The result is a series of spectacular graben-like depressions, which are 150 to 200 m wide and are 25 to 75 m deep. Gliding of sandstone blocks away from cliff lines is not limited to areas of highly deformable substrate. South of Sydney, Australia, the movement of large blocks of Nowra Sandstone away from cliff lines is the result of the very slow creep of underlying sandy siltstone (Plate 110).

Many outcrops of sandstone have been shaped into domes and rounded slopes, known as slickrock. In some cases curved sheeting in the sandstone roughly parallels the rounded topography. Most rounded slopes, however, appear to be the result of the granular breakdown of the sandstone and of the peeling of thin, weathered layers from the surface. This is so both in arkosic rocks, such as the Navajo Sandstone of the Colorado Plateau, and in quartz arenites, such as those which are cut into complex arrays of rounded towers in the Bungle Bungle Range of north-west Australia. Granular breakdown and slab failure also are the primary processes in the development of arches, which are quite common in some sandstones. The domes and slickrock slopes that are characteristic of many coarse conglomerates can likewise be attributed to granular breakdown and slab failure.

Weathering of sandstones varies with the mineralogy of its constituents. Where there is considerable matrix, hydration of clay is important. In arkosic sandstone, the primary process is the weathering of feldspars. Even highly quartzose sandstones are subject to extensive, though slow, CHEMICAL WEATHERING, with the order of decay generally being first the siliceous, intergranular cement, then the quartz grains, and finally the quartz overgrowths. The presence of sodium chloride, either as sea spray or as aerosols in rain, has a strong accelerating effect on the dissolution rates of silica. These various processes are dependent on the ability of water to penetrate the sandstone. The opening of joints and generation of microfractures by initial release of confining pressures create numerous pathways for seepage. Even the alteration from fresh to slightly weathered sandstone can result in a reduction by a factor of two or three in mean strength, and of about six in deformability, as porosity and moisture content increase. The advance of weathering depends very much on the permeability of the sandstone, for, where the original pore spaces are filled by overgrowths or siliceous cement, active weathering is essentially limited to a thin surface layer. Nevertheless, prolonged weathering can produce deep solutional features even in quartzites.

The removal of the more soluble constituents in sandstone, and the precipitation of siliceous and iron-rich materials, may result in case hardening of the surface layer. Some case-hardened surfaces develop distinctive polygonal patterns of cracks, known as 'elephant skin'. The cracking seems to be the result of fatigue in the case-hardened layers and of changing surface stresses, similar to the crazing of glazed pottery surfaces. Fire is an important agent in the breakdown of sandstone surfaces in well-vegetated areas. Near Sydney, such surfaces are likely to be exposed to fire once every ten to twenty years. Fire moving rapidly over sandstone produces minor spalling and granular disintegration, while intense fires may cause spalling to depths of 2 cm. Sand blasting of sandstone surfaces during high winds has been overrated, especially as a cause of CAVERNOUS WEATHERING in humid areas, but is very important in arid lands. Shattering by crystal pressures developed in fractures due to freezing is particularly important under cold climates, especially in quartzites, which, though very resistant to abrasion, are brittle rocks that can withstand only slight strain deformation before they rupture explosively.

Initial classifications of sandstone landforms at the regional scale were based on the presumed controlling effect of climate. Climate is undoubtedly the dominant control in the amazing wind eroded YARDANG landscape of the hyper-arid Boukou region of the Sahara. Pseudokarst assemblages, like the impressive array of deep dolines and caves etched into the quartzites of the Roraima region of Venezuela, are certainly best developed in the humid tropics. But solutional features occur extensively on sandstones outside the tropics. Furthermore, some of the most common types of sandstone terrain, such as rounded slickrock relief, and the angular towers and great cliffs of ruiniform assemblages, occur under various climates. The variable properties of the rock, rather than climate, may thus be of primary importance in the shaping of sandstone landforms.

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R.W. YOUNG

SAPROLITE

Saprolite is weathered rock in which the fabric of the original rock is preserved. The word derives from the Greek sapros (σαπρos or σαπροσ) meaning rotten and was coined in 1895 by Becker. The term is generally applied to chemically altered rocks where all or part of the primary minerals are changed to new-formed minerals. Most often saprolite is referred to granitic rocks, but the term applies to the weathered component of any rock type.

The body of *in situ* weathered rock referred to as saprolite may comprise a number of zones (horizons, layers) depending on the relative rates of weathering, erosion and the composition and hydromorphic characteristics of the regolith. Figure 141 summarizes data on saprolite and various terms used by various authors for various parts of weathering profiles and their saprolite component.

Starting from the bottom of a weathering profile *saprock* is the part of the profile closest to the weathering front (Figure 141). Saprock is rock that has begun to weather, but only about 20–30 per cent of the primary minerals are chemically altered. This is hard to estimate and a preferable definition is that the material requires a sharp hammer blow to break it. The zone of saprock may contain boulders, beds or other masses of unweathered bedrock. At the weathering front, which may be very sharp or gradual, rocks change from fresh to partly weathered, alteration of feldspars to clay minerals is seen and ferromagnesian minerals release Fe^{2+} that is oxidized to red-yellow coloured Fe^{3+} oxihydroxides or oxides.

The form of the WEATHERING FRONT may be relatively flat or very irregular, the latter being more common. Its shape is primarily dependent on the nature of the rocks being weathered. In massive igneous bodies the form is generally more regular – but with isolated sub-spherical boulders or corestones of fresh rock in the saprock – than that in dipping sedimentary rocks or metamorphic rocks. The number and orientation of joints, cleavage and bedding planes that form the initial conduits for the weathering solutions mainly control this. Plates 111–115 are examples of different types of saprolite and weathering fronts.

Moving up profile the saprock gradually gives way to saprolite where the majority of labile (more easily chemically weathered) minerals are altered and replaced by new minerals formed by the chemical weathering. The most common minerals in saprolite are clays, iron oxihydroxides and oxides and any minerals largely resistant to weathering (e.g. quartz, magnetite or pre-existing clay minerals). The clay minerals generally gradually change from smectite (Si/Al = < 1) lower in the profile, where drainage is generally poorer than higher where the dominant clay is kaolinite (Si/Al = 1). Kaolinite forms, as drainage of the weathering solutions is more efficient.

Above the saprolite is collapsed saprolite or the 'mobile zone'. In this zone (Figure 141) the

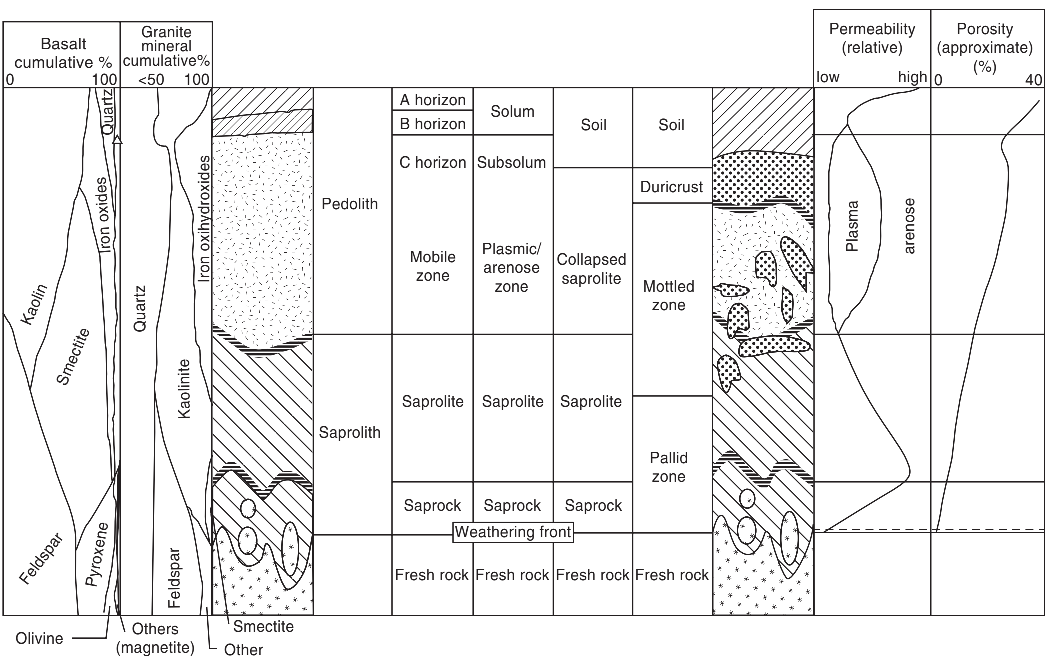

Figure 141 Some of the various terminologies used to describe weathering profiles including saprolite. On the left are two sketch examples of how mineralogy varies through the saprolite and on the right is an example of how the hydraulic properties of the weathering profile and saprolite vary with depth. (modified from Taylor and Eggleton 2001)

Plate 111 A small erosional remnant of a 'typical' weathering profile developed in felsic gneiss on the Yilgarn near Kalgoorlie in Western Australia. It is about 3 m high and shows ferruginous cap about 0.25 m developed over a 0.5 m thick mottled saprolite that grades downward into 3 m of gneissic saprolite. The saprolite is composed of relict quartz grains, and kaolinite with minor haematite. The composition of the mottled saprolite is much the same with increased quantities of haematite. The ferruginous crust is mainly haematite with quartz and minor goethite (photo Ian Roach)

Plate 113 A very deep weathering profile formed in ultra-mafic rocks at Marlborough in central Queensland, Australia. About 40 m of the profile is exposed, but no fresh rock occurs in this pit. The total profile depth is up to about 100 m. The bulk of the material in the photo is saprolite composed of clay minerals, mainly nontronite and talc with Fe^{3+} oxides, goethite and haematite and secondary silica as chrysoprase and chalcedony. The upper part of the profile is transported debris that in places has filled karstic channels in the upper saprolite. The hill is capped by siliceous duricrust derived from silica mobilization during the weathering of the ultra-mafic rocks

Plate 112 A deep weathering profile on the Yilgarn Craton of Western Australia showing similar features to those described in Plate 111. The profile here is some 30 m thick and at the base of the pit an irregular weathering front between fresh and weathered felsic granites can be observed. The original rock structures (joints) are clearly preserved in the saprolite in this profile

Plate 114 A 10 m section through a 20 m weathering profile formed in Ordovician intermediate volcanic rocks at Northparkes in New South Wales, Australia. This profile is unusual in that the pallid or bleached saprolite occurs above the mottled zone in the right of the photograph. It comprises mainly kaolinite in the bleached zone with minor secondary quartz and calcite and gypsum. The lower mottled zone has similar mineralogy with the addition of haematitic mottles. These mottles have been dated as Carboniferous (Pillans *et al*. 1999). The saprolite has been partly eroded and covered in part by palaeochannel sediments (left of photo) that are themselves weathered significantly and now form sedimentary saprolite with large haematitic and goethitic mottles. These mottles date from the Miocene. The whole profile is overlain by up to 2 m of Quaternary alluvium with a red-brown earth formed in it

Plate 115 A 7-m thick section through Proterozoic quartzite unconformably overlain by saprolite formed from Cretaceous labile sandstones and Quaternary transported cover at Darwin, Australia. The Proterozoic rocks are fresh. The Cretaceous has been completely transformed to a quartz/kaolinite saprolite and the whole sequence overlain by ferruginous lag gravel, sand and clays. It is interesting that the upper surface of the saprolite is karstic and that the overlying lags, etc. have filled the karstic channels

saprolite has been chemically eroded (weathered) to such a high degree that the original rock fabric is no longer self-supporting and it has collapsed. This process of collapse may also be enhanced by processes of bioturbation (e.g. tree roots, termites) and/or pedoturbation (e.g. wetting and drying, and illuviation). It is also from this point in the profile that the REGOLITH may move down slope (Plate 116) under the influence of gravity. Also within this zone, quartz sand and clays may begin to separate into clots of clay surrounded by sand (Taylor and Eggleton 2001: 161) by processes of pedoturbation.

Plate 116 Overturned Proterozoic metasediments moving down a slope of about 1.5° at the Mary River, Northern Territory, Australia

Plate 117 Close-up photograph of mottles on a wave cut platform at Darwin, Australia. The hammer provides scale. These mottles are up to 15 cm across and are composed of haematite, providing the red colour, cementing a saprolitic matrix of kaolinite and quartz. The intervening bleached saprolite is identical, except that it does not contain haematite

Secondary overprints may affect the appearance of the weathering profile and the saprolite. The most widely observed modification is the formation of iron oxihydroxide aggregations called mottles. Mottles (Plate 117) may extend throughout the profile, but are mostly observed in the upper collapsed saprolite. Another common feature in profiles developed mainly on felsic rocks is a bleaching or the removal of Fe-oxihydroxides from the lower parts of the saprolite and saprock. This zone is referred to by many as a pallid zone, particularly in what are described by some geomorphologists as a 'lateritic profile'.

Figure 141 (p. 909) shows idealized examples of how some physical and mineralogical properties of saprolite change through a profile. One point of interest here is the commonly observed addition of quartz to the upper parts of saprolite developed from basalt, which of course contains no quartz when fresh. This indicates the addition of 'foreign' material to saprolite profiles may occur either by overtopping of the saprolite or by other deposits from aeolian accession. This is a very common feature of most saprolite profiles but is most readily observable over weathered mafic rocks.

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GRAHAM TAYLOR

SASTRUGI

Sharp irregular ridges, mounds or dunes. They form on ice sheets, ice caps, sea ice and tundra (typical of Antarctica and Greenland), and are composed of ice and compacted snow. Originating from the Russian word *zastrugi*, they are formed by aeolian erosion and the deposition of drifting snow (Gow 1965), and typically are 1–2 m long and 10–15 cm in height (though exceptional cases can reach 1.5 m in height and hundreds of metres long). Sastrugi align longitudinally with the predominant wind direction, making it possible to infer the prevailing wind direction at the time of sastrugi development from their configuration. They often form after a blizzard on the hard ice surface, becoming larger and harder as the blizzards blow across them throughout the winter months. Sastrugi are usually found in the lee of obstacles but are also known to exist in open conditions.

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STEVE WARD

SCABLAND

A scabland is an erosional landscape formed by a catastrophic flood and is generally applied to the effects of Jökulhlaups. It was first introduced by Bretz (1923) to describe the erosion and stripping of the Columbia Plateau Basalts by floods from Glacial Lake Missoula in eastern Washington, USA. Bretz adopted a term that had been used by local farmers to describe the 'scabby' terrain: 'The terms "scabland" and "scabrock" are used in the Pacific Northwest to describe areas where denudation has removed or prevented the accumulation of a mantle of soil, and the underlying

Table 41 Bedforms in the Channeled Scabland

	Scoured in rock	Scoured in sediment	Depositional
Macroforms (scale controlled by channel width)	Pool and riffle sequence, Quadrilateral residual forms in channel Anastomosis	Large-scale streamlined residual forms	Longitudinal bars (a) Pendant bars (b) Alternate bars (c) Expansion bars Eddy bars
Mesoforms (scale controlled by channel depth)	Longitudinal grooves Pot-holes Inner channels Cataracts	Scour marks	Large-scale transverse ripples (giant current ripples)
Microforms	Scallop pits	Not preserved	Small-scale ripple stratification (restricted to slack water facies)

Source: From Baker (1978a)

rock is exposed or covered largely with its own coarse, angular debris' (Bretz 1923: 617).

Channeled Scabland

The formal physiographical region known as the 'Channeled Scabland' is located in the northern portion of the Columbia Plateau in eastern Washington, USA and comprises an area of approximately 40,000 km^2. It is a spectacular channel complex eroded deeply into loess and basalt bedrock. The large flood discharges spilled over pre-flood divides into adjacent valleys and produced the effect of channels dividing and rejoining to form anastomosing (see ANABRANCHING AND ANASTOMOSING RIVER) complexes. These divide crossings are several hundred feet above valley floors.

A typical scabland complex includes erosional and depositional forms. Baker (1978a) has adopted a hierarchical classification of bedforms for the Channeled Scabland (see Table 41). The erosional landforms include grooves, pot-holes, rock basins, inner channels and cataracts. Bretz *et al*. (1956) ascribed several scabland features to differences between various basalt flows. Cataracts, such as Dry Falls, formed as one group of basalt flows was stripped from underlying resistant flows. Where they were exposed by the floods, the columnar jointed basalts exerted a strong joint control and the plucking action by floodwater yielded boulders >30 m diameter. The most spectacular of the depositional forms are the streamlined channel deposits, some superimposed by giant current ripples 0.5 to 7 m high and 18 to 130 m in chord length. They are composed predominantly of gravel and boulders. Slackwater deposits accumulated in low velocity areas including re-entrants to the major valleys and in pre-flood tributary valleys.

The scablands were formed by discrete outbursts from a range of sources. These included an enormous subglacial reservoir that extended over much of central British Columbia (conservative estimates of water volume are 10^5 km^3) (Shaw *et al*. 1999) and Glacial Lake Missoula. This lake impounded 2,184 km^3 of water during its maximum extent (O'Connor and Baker 1992). The last major period of scabland flooding is placed approximately between 18,000 and 13,000 years BP. Facies analysis of sedimentary sequences suggest that there may have been as many as forty floods (Waitt 1985) each separated by decades or centuries. Shaw *et al*. (1999) propose that there were fewer floods and that many of the variations in the sedimentary sequences can be ascribed to pulses within a flood caused by the input of multiple sources of floodwaters during these long duration flows (up to 1,000 days).

High water marks along the scabland channels have been used to reconstruct the maximum flood stages and water surface gradients. These include eroded channel margins, depositional features, ice rafted ERRATICS and divide crossings. Discharges as large as $21.3 \times 10^6\,\mathrm{m^3\,sec^{-1}}$ were conveyed through the channel scabland (Baker 1978b). Some constricted channels reached velocities as high as 30 m sec^{-1}. These high velocities were

Plate 118 Portion of MOC image M2101914 which is centred near 7.89ºN, 153.95ºE, pixel resolution is 4.4 m. (Malin *et al.* 2001). This image shows anastomosing channel pattern and multiple streamlined forms in a flood channel emanating from a fissure. For detailed description see Burr *et al.* (2002)

possible because of the combination of great flow depth (60 to 120 m) and very steep water surface gradients 2 to 12 m/km.

On Mars, data from orbiting satellites have detected stripped zones on the floors of outflow channels in the Chryse Basin. These anastomosing complexes are 100 km wide and flow over 2,000 km across the planet's surface. They are usually initiated from collapsed zones. Other examples show multiple and asynchronous flows emanating from geothermal fissures in recent Martian history (see Plate 118) (Burr *et al.* 2002). By analogy to the scablands on Earth, it is generally accepted that the Martian outflow channels were also formed by catastrophic floods. Martian outflow channels include a distinctive assemblage of scabland landforms: regional and local anastomosing patterns, expanding and contracting reaches associated with flow constrictions, streamlined hills, inner channels with recessional headcuts, pendant forms (bars or erosional residuals) on the down current sides of flow obstacles, longitudinal grooves, irregular 'etched' zones on channel floors and scour marks around obstacles (Baker 1982).

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MARY C. BOURKE

SCANNING ELECTRON MICROSCOPY

The Scanning Electron Microscope (SEM), sometimes used in association with Energy Dispersive Spectrometry (EDS), has been used for studying the surface textures (and chemistry, with EDS) of sediments (especially quartz grains) since 1962. The use of the SEM has had many implications for geomorphology, including the determination of the origin of depositional landforms, the provenance of sediments, the energy of environments and processes of diagenesis and weathering and their development through time. Examples of the use of the SEM include the separation of till from glacilacustrine and glacifluvial grains within the glacial sedimentary environment, studies of the origin of fine silt and clay particles in the geological column, examination of aeolian and other environmental fracture–abrasion mechanisms in the field and the laboratory, and analysis of grain modification under different weathering regimes. A full discussion of grain textures associated with aeolian, fluvial, mass wasting, glacial, tectonic, impact, weathering and diagenetic processes is provided by Mahaney (2002).

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A.S. GOUDIE

SCHMIDT HAMMER

A concrete test hammer originally designed by E. Schmidt in 1948 for carrying out *in situ* tests on the hardness of concrete. The instrument measures the distance of rebound of a controlled impact on a rock surface. Because elastic recovery (the distance of repulsion of an elastic mass upon impact) depends on the hardness of the surface, and hardness is related to mechanical strength, the distance of rebound (R) gives a relative measure of surface hardness or strength.

In the Schmidt Type N hammer (which weighs 2.3 kg) the energy of impact (0.224 mkg) is obtained by releasing a spring-controlled plunger. The R value is shown by a pointer on a scale on the side of the instrument (range 10–100). This value represents the rebound distance as a percentage of the forward movement.

The Schmidt Hammer Type N is light and portable and allows *in situ* tests to be made in the field. By enabling quantitative comparison of the hardness of materials it provides a useful tool for geomorphologists. Among its uses have been the description of *nari* (calcrete) profiles, case hardening on tropical karst surfaces, and various types of aggregate resources (Day and Goudie 1977). It has also been much used to assess degree of weathering and the ages of geomorphic features upon which weathering phenomena occur (Ballantyne *et al.* 1989; McCarroll 1991).

Schmidt hammer rebound values have been found to correlate well with other measures of rock strength, including Young's Modulus and Uniaxial Strength (Katz *et al.* 2000)

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A.S. GOUDIE

SCREE

The terms scree and talus are synonymous, the former being preferred in England and the latter (equivalent to the French word for slope) used predominantly in North America and elsewhere. Both terms describe accumulations of loose, coarse, usually angular rock debris at the foot of steep rock slopes. The terms are used to describe both the landform and the material of which it is composed. The debris accumulations forming screes must be of sufficient thickness to develop a characteristic morphology independent of the underlying slope. Simple debris veneers only a few particles thick are termed 'debris mantled slopes' (Church *et al.* 1979).

Scree slopes occur in a wide range of environments but most significantly in environments where physical weathering processes dominate. The production and/or accumulation of debris must be greater than its subsequent weathering or removal. The coarseness of most scree deposits makes them resistant to subsequent erosion: they are often stable long-lasting elements of the landscape and may be preserved as fossil forms.

The basic characteristics of scree slopes depend primarily on the morphology (and thereby geology) of the flanking cliffs and the geomorphic processes involved in their development. Although the dominant process is usually assumed to be ROCKFALL, scree modification and accumulation may be the result of several different processes acting singly or in combination and several distinctive types of scree may be recognized. The plan morphology of screes depends on the form of the cliffs supplying the debris and the morphology of the surface on which the debris accumulates. Relatively simple cliffs, straight in plan view, tend to produce sheet (straight) rockfall talus slopes lacking significant lateral variation in their characteristics. As cliffs become more dissected, deposition is focused below couloirs (gullies) leading to the development of cones. As well as funnelling rockfalls, couloirs channel other rockwall processes (surface stream flow, snow avalanches (see AVALANCHE, SNOW), etc.) down the cliff, sometimes resulting in significant modification of the scree below. Therefore the cones developed below dissected cliffs are rarely simple single-process forms. In alpine environments debris cones can be significantly modified by snow avalanche activity. DEBRIS FLOW generation may also occur during heavy rainstorms when drainage from the cliff zone is focused by gullies onto finer grained materials at the head of the scree.

Rockfall screes result from the accumulation of discrete rockfall events over long periods of time. The basic characteristics are fall sorting (a logarithmic increase in mean grain size downslope) and a straight slope, often with a well-developed basal concavity. There is continuing debate about whether these straight slopes reflect the angle of repose (see RESPOSE, ANGLE OF) of coarse cohesionless material at about 35° (Carson 1977). Measured profiles of the upper part of many rockfall screes range between 32° and 37° but locally reach 40°. The degree of basal concavity varies with the length of slope and the characteristics of the basal zone.

Fall sorting on rockfall scree slopes, though not universal, primarily results from two mechanisms. Larger boulders have greater momentum and therefore tend to travel further downslope. Second, the frictional resistance (roughness) of the surface over which the boulder slides, rolls or bounces is defined by the relationship between the size of the moving boulder and the irregularities of the surface (boulders and voids) over which it is passing. Big boulders are only effectively trapped in large 'holes' or when they impact other large boulders. The degree of sorting depends on the slope length, cliff height and the size and shape of dominant particles. Locally random effects or differences in particle shapes may result in the absence or anomalous patterns of sorting. On scree slopes modified by snow avalanches loose surface material is swept from the upper slopes to the end of the avalanche track at or beyond the base of the scree, in extreme cases forming AVALANCHE BOULDER TONGUES. These screes show little size sorting on upper slopes but a rapid increase in mean grain size towards the base despite the presence of an unstable scattering of loose rock debris on the surface of larger clasts. Avalanche modified screes have strong basal concavities. Many scree cones have gullies formed by fluvial activity at their head and may have significant debris flow forms (levees, channels and terminal lobes) extending across the scree slope. In extreme cases these have been termed 'alluvial talus'. Scree-like forms produced by the breakup of single large rockfalls generally lack fall sorting and have more complex long profiles. In alpine areas large multiprocess scree cones may display complex surface characteristics associated with the interaction of these processes (Plate 119).

Most scree slopes, despite their coarse debris veneer, have considerable quantities of interstitial fine material at depth or exposed on the higher parts of the slope. They are not exclusively formed of coarse cohesionless material as many early models assumed. The upper parts of the scree slopes undergo 'talus' creep which is the aggregate movement caused by rockfall and other impacts, FREEZE–THAW CYCLE or frost heave activity, percolating flows, animals, etc. Loose material on very steep sections may undergo dry avalanching but such failures are usually small. Little if any talus creep seems to occur on coarse basal screes.

Plate 119 Multiprocess screes, Bow Lake, Alberta, Canada. Complex sheet and scree cones showing the varying influence of alluvial, debris flow and snow avalanche activity on the detailed morphology of these landforms predominantly created by rockfall. The basal area of these cones show evidence of permafrost creep and incipient (or arrested) rock glacier forms

Screes are most frequently studied in periglacial environments. At some of these sites a permanent snow (firn) patch may develop at the base of the slope. Debris landing on this icy surface slides to the base accumulating as a ridge (PROTALUS RAMPART or nivation ridge). In alpine areas, thick talus accumulations may develop PERMAFROST. Subsequent deformation and creep of this rock/ice mixture leads to ROCK GLACIER development.

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SEE ALSO: frost and frost weathering; geomorphological hazard; grèze litée; hillslope, form; hillslope, process; mass movement

BRIAN LUCKMAN

SEA LEVEL

Sea level is the divide between the marine and terrestrial realms; above it, the world is dominated by air, erosion and creatures that contend with gravity; below lies a submerged world dominated by sedimentation and neutrally buoyant animals. Relative to the landmasses, the position of sea level has fluctuated throughout the geologic past, owing to changes in both the quantity of ocean water and the geometry of the ocean basins. Although the total quantity of water in the hydrosphere probably has changed little since Archean times, the fractions held in land reservoirs – glaciers, lakes, ground water and, in particular, continental ice sheets – have fluctuated significantly. For example, if the present Antarctic ice sheet were to melt, sea level would rise by about 55 m; at the height of the last ice age, which is only one of many such events to have punctuated Earth history, sea level was about 120–130 m below its present position.

The proportions of land and sea are basically determined by the fact that continents, being composed of rock lighter than oceanic crust, stand about 4.5 kilometres above the ocean floor (in contrast, hot and oceanless Venus has no features resembling great ocean basins). However, owing largely to slow variations in the rates of seafloor spreading and plate tectonics, the average depth of the ocean basins has varied throughout geological time and shorelines have periodically advanced and retreated across the continental shelves. Figure 142 illustrates examples of observed sea level change on different time scales, from about a year to 10^8 years. On the longer timescale, sea level changed globally with amplitudes up to several hundred metres, largely owing to plate-tectonic changes in ocean basin geometry (Figure 142a). On timescales of tens to hundreds of thousands of years, periodic exchanges of mass between the ice sheets and oceans caused sea-level changes of tens to over a hundred metres in amplitude (Figure 142b).

Global changes of sea level caused by changes of the volumes of seawater or the ocean basins are referred to as *eustatic*. Superimposed on these global signals are more regional and local changes. At decadal, annual and shorter intervals, meteorology – and tide-driven changes become important and vary from place to place (Figure 142c). Over longer times, the relative positions of land and sea are affected by uplift or subsidence of the coastal zone. Observations vary substantially from site to site, even over relatively short distances such as in Scandinavia, where sea level at Ångerman has fallen nearly 200 m in the past 9,000 years while at Andøya the level 9,000 years ago was near its present position. In southern England, levels have risen slowly over the past 7,000 years, but along the Australian margin they have fallen by a few metres during the same interval (Figure 143).

Many factors that contribute to changes in sea level are linked. When ice sheets melt, the resulting sea-level change is spatially variable because the Earth's surface deforms under the changing ice and water load, and because the gravitational potential of the Earth–ocean–ice system also changes. The combined deformation–gravitational effects are referred to as the *glacio-hydro-isostatic* contributions to sea level, and it is they that cause the spatial variability illustrated in Figure 143.

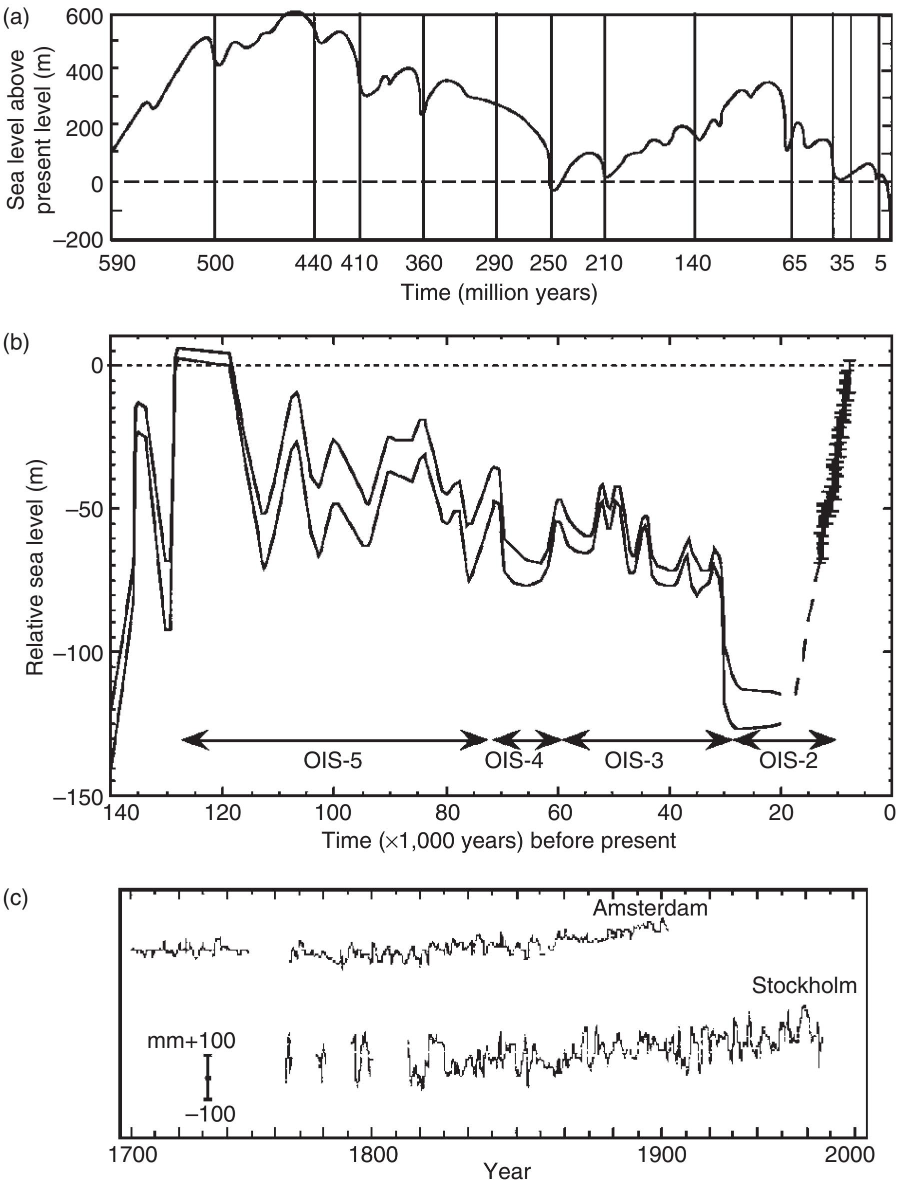

Figure 142 Sea-level variation at three timescales. (a) ~10^8 years, inferred from seismic sequence stratigraphy (Hallam 1992; Haq *et al.* 1988). The higher frequency changes reflect both global and local effects; the large, slow changes reflect continental breakup and changes of ocean ridge systems. (b) ~10^5 years of relative sea level at Huon Peninsula (HP), Papua New Guinea, driven by changes in northern continental ice sheets. Bars show marine oxygen isotope stages (OIS) discussed in text. (c) 10^0–10^2 years, measured by tide gauges from Amsterdam and Stockholm (a secular fall of ~4 mm/year has been removed from the Stockholm record): these changes are primarily of climatic origin and the apparent small rise starting AD ~1880 may reflect global warming (from Lambeck and Chappell 2001)

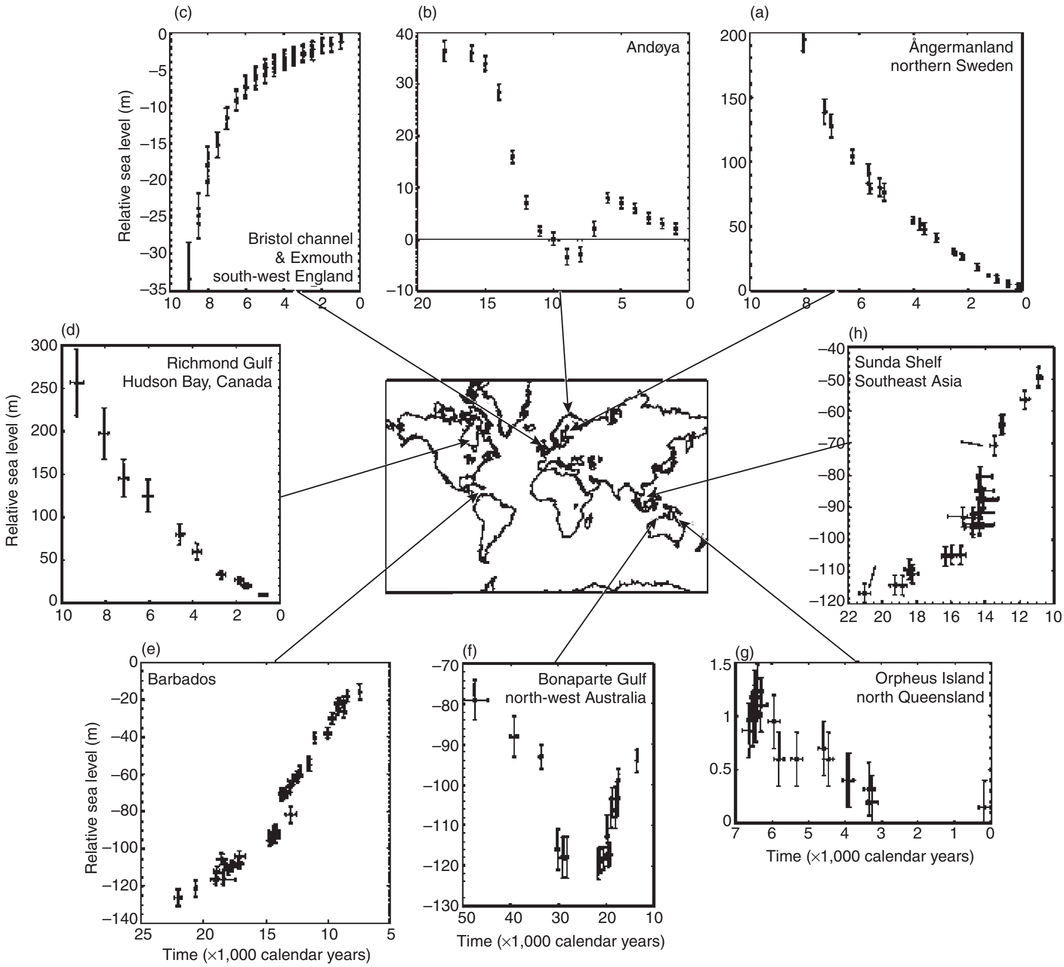

Figure 143 Observed variability of sea-level change in last 20,000 years from tectonically stable areas or sites where the tectonic rate is known and has been removed. (a) Ångerman, Gulf of Bothnia, Sweden. (b) Andøya, Nordland, Norway. (c) South of England. (d) Hudson Bay, Canada. (e) Barbados. (f) Bonaparte Gulf, north-west Australia. (g) Orpheus Island, north Queensland, Australia. (h) Sunda Shelf, Southeast Asia. (Note: scales differ from graph to graph). (Data sources given in Lambeck and Chappell 2001).

Isostatic warping and subsidence also occurs in sedimentary basins, in response to the accumulation over millions of years of sediments many kilometres deep. Furthermore, *tectonic* movements drive mountain uplift, enhancing landscape denudation and leading to rapid accumulation in sedimentary basins, and through coastal uplift or subsidence also contribute to changes of relative sea level.

Knowledge of the complex history of relative changes of sea level has applications in fields as diverse as understanding climate changes, analysing the structure of petroleum-bearing sedimentary basins, and determining deep-earth properties such as the viscosity of the Earth's mantle. Once comprehensive sea-level models are developed, it becomes possible to test hypotheses about the migrations of flora and fauna across shallow seas that are now covered by the ocean. Finally, to understand future sea-level rise under atmospheric greenhouse conditions, the background 'natural' signal must be known. The success of the outcomes of the various sea-level studies depends very much on the ability to separate the different contributions – eustatic, isostatic and tectonic – in the observational record.

Observational evidence

Evidence for historical sea-level changes comes from tidal marks and gauges, usually in ports and harbours, as well as buildings and other structures in littoral towns such as Venice. For the geologic past, the evidence occurs mostly in the form of sediments and biohermal reefs that were formed in coastal and nearshore situations. Upper Quaternary sea-level changes are usually pieced together from raised or submerged shoreline features, including shallow-water coral reefs (Figure 144). Further back in time, relative sea-level changes can be deduced from sedimentary basins, in which eustatic variations are registered in cyclic sediments (cyclothems) (Figure 145), and by alternating transgressive and regressive sediment tracts. Using the methods of *sequence stratigraphy* based on seismic and drillhole or outcrop data, the locus of coastal-zone sediments can be traced throughout a given basin sequence, allowing the course of sea-level changes relative to the basin to be identified (Hallam 1992; Haq *et al.* 1988).

A relative sea-level curve for a given area can be established from age–height data for a series of ancient shorelines, or other *indicator deposits* such as shallow marine sediments for which the depth of deposition relative to sea level is determinable. Terrestrial deposits within a sequence, such as peats and floodplain sediments, may usefully indicate levels not reached by the sea. Various methods are used to establish the ages. For deposits less than ~40,000 years old, *radiocarbon* dating is used widely, although *uranium-series* dating is preferable in the case of coral formations and has a greater time range, extending to ~0.5 Myr. Thermoluminescence (*TL*) and optically stimulated luminescence (*OSL*) methods are increasingly used for dating Upper Quaternary shoreline deposits, and amino acid racemization has proved to be useful despite its low precision. Sea-level indicator deposits also are correlated to marine oxygen isotope records, described later, that have a chronology based on slow variations in the Earth's orbit, which affect the seasonal receipts of solar radiation and acted as an ice-age pacemaker. *Orbital chronology* rests on astronomical observations and has been extrapolated several million years into the past. Finally, the dating of older deposits and sedimentary basin sequences generally rests on *geomagnetic reversal*

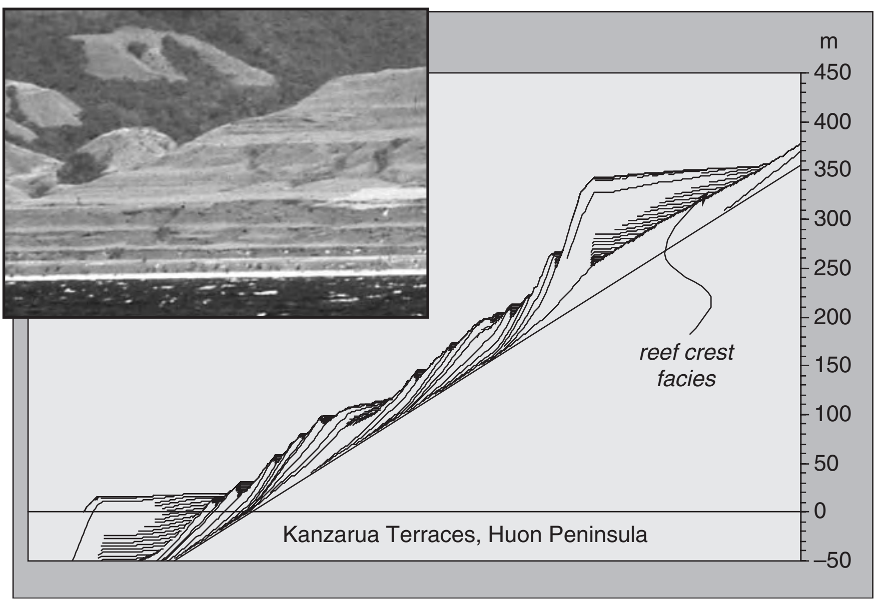

Figure 144 Typical expression of Late Quaternary sea-level changes superimposed on tectonically rising terrain: coral terraces Huon Peninsula, Papua New Guinea. The top terrace at right of the inset photo is 350 m above sea level and formed at the beginning of the Last Interglacial, ~127,000 years ago. Each of the smaller reef structures within the downstepping stratigraphic sequence was formed during a sea-level oscillation within the last glacial cycle. Subtracting the effect of uplift yields the sea-level curve shown at Figure 142b

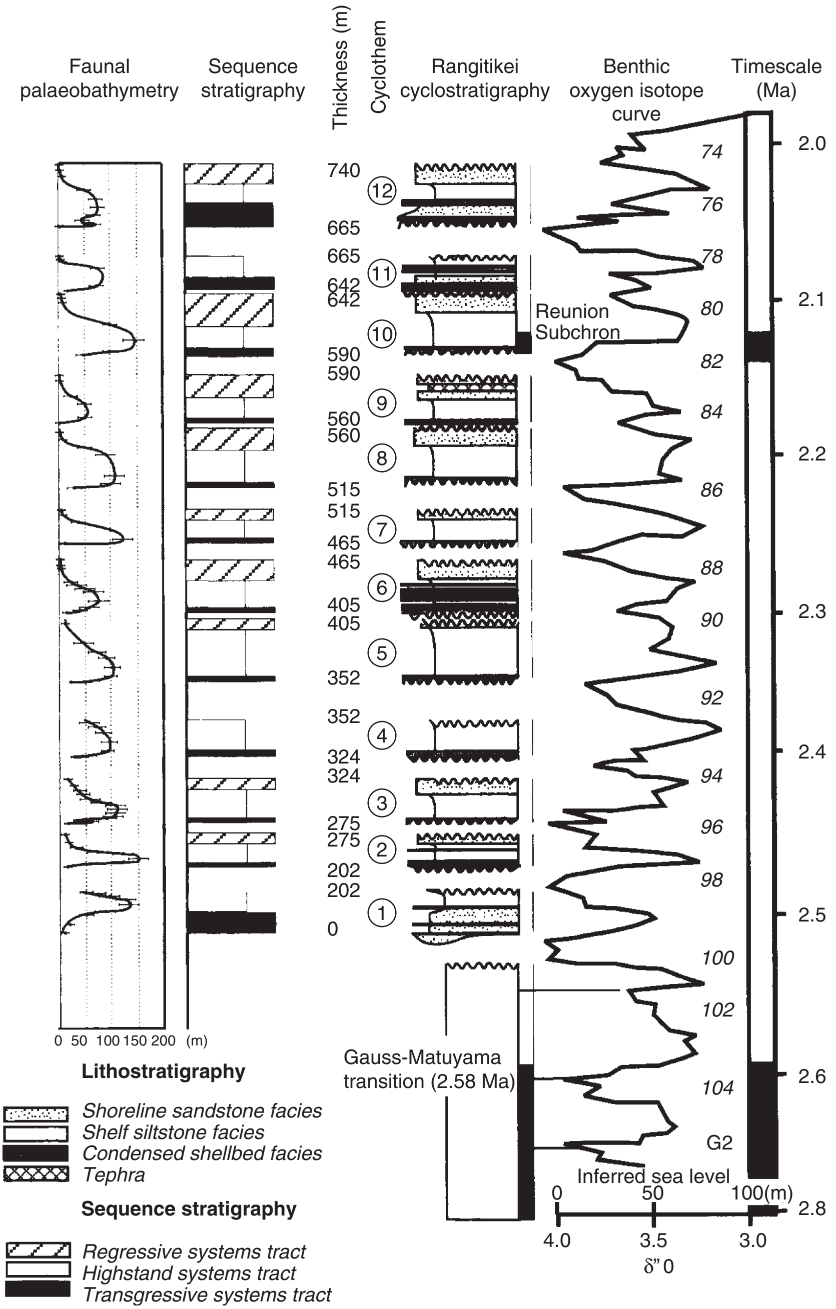

Figure 145 Late Pliocene sea-level changes inferred from shallow marine cyclothems in New Zealand. Centre columns show 12 sedimentary cycles; left column shows cyclic variations of water depth at the site of sedimentation, inferred from fossil marine invertebrates. Correlations to global marine oxygen isotope cycles (Shackleton *et al*. 1995) is shown at right, tied to magnetic reversal chronology (from Pillans *et al*. 1998)

chronology or on microfossil-based correlations with standard stratigraphic sequences tied to this chronology, which in turn is secured by potassium-argon dating methods.

Separation of eustatic, isostatic and tectonic components

RAW SEPARATION OF UNIFORM UPLIFT OR SUBSIDENCE

A first step towards separation of the tectonic, isostatic and eustatic contributions is subtraction of any obvious vertical crustal movements from a relative sea-level curve. More often than not, this is done by assuming that the sea level for some reference deposit in the record is known and also that the rate of uplift (or subsidence) was constant throughout the duration of the record. With these assumptions, the local sea level S represented by a deposit of age t that accumulated at depth d below sea level and now is height H above present sea level, is given by

$$S = H + d - Ut \quad \text{with} \quad U = (S_r - H_r + d_r)/t_r \tag{1}$$

where H_r, d_r and t_r are height, depositional depth and age of the reference deposit, and S_r is the sea level at its time of formation. For Upper Quaternary studies, the local height of the Last Interglacial shoreline is widely used to determine uplift rate U, as evidence reviewed below suggests that sea level then was little different from that of today. However, for every study, each variable in (1) has an error term arising from dating errors and uncertainties in field relationships. Moreover, in assuming a constant rate of uplift, this approach neglects reversing vertical movements that arise from the global glacio-hydro-isostatic response to advancing and retreating icesheets. Figure 142b shows the 'uplift-free' sea-level curve derived by this method from the coral terraces illustrated at Figure 144.

Similar principles are used to derive sea-level changes from sedimentary basin sequences, although here the vertical movement is downwards. In terms of sedimentary facies, individual cyclothems often are very similar from bottom to top of a thick cyclothemic sequence (Figure 145), implying fairly uniform subsidence of the basin.

GLACIO-HYDRO-ISOSTASY

When ice sheets melt, to a first approximation the sea level rises by an amount $\Delta\zeta_e(t)$ related to the land-based ice volume V_i (using the notation of K. Lambeck: see Lambeck and Chappell 2001),

$$\Delta\zeta_e(t) = -(\rho_i/\rho_o)\int(1/A_o(t)\ dV_i/dt)\,dt \tag{2}$$

where $A_o(t)$ is the ocean surface area (which changes as sea level rises or falls) and ρ_i, ρ_o are the average densities of ice and ocean water, respectively. $\Delta\zeta_e(t)$ is the ice-volume equivalent sea-level change (or *equivalent sea-level change*), which equals eustatic change if no other factors contribute to changes in ocean volume. The relative sea-level change $\Delta\zeta_{rsl}(\varphi, t)$ at position and time t, ignoring tectonic displacements, is

$$\Delta\zeta_{rsl}(\varphi, t) = \Delta\zeta_e(t) + \Delta\zeta_i(\varphi, t) + \Delta\zeta_w(\varphi, t) \tag{3}$$

where $\Delta\zeta_i$ and $\Delta\zeta_w$ are the glacio- and hydro-isostatic contributions. Both are functions of position and time. The water depth or terrain height, expressed relative to coeval sea level, is

$$h(\varphi, t) = h(\varphi, t_0) - \Delta\zeta_{rsl}(\varphi, t) \tag{4}$$

where $h(t_0)$ is the present-day (t_0) bathymetry or topography at φ. Both isostatic terms in (3) are functions of Earth rheology as well as of fluctuations in the ice sheets over time.

In formerly glaciated areas, the glacio-isostatic term $\Delta\zeta_i(\varphi, t)_i$ dominates during and after deglaciation, and leads to uplift at a rate that can exceed the global eustatic rise, so that sea level locally falls (the Ångerman result: Figure 143). Rebound is smaller near the ice margin and although it may dominate initially, the global sea-level rise becomes important later. When all melting has ceased, the residual rebound leads to falling local sea level (the Andøya result: Figure 143). During ice sheet growth, mantle material beneath the loaded area is displaced outward and broad bulges develop around the perimeter, which subside when the ice melts, leading to slowly rising local sea level after melting has ceased. Much further from the ice, the water load becomes the dominant cause of planetary deformation (the hydro-isostatic contribution $\Delta\zeta_w(\varphi, t)$), producing subsidence of the seafloor and adjacent margins. The effect is most pronounced at continental margins far from the ice sheets, such as the Australian coast (Figure 143), and once melting has ceased, sea levels continue to fall at a slow but perceptible rate. The amplitude of this postglacial 'highstand' effect can vary by several metres from site to site.

Isostatic corrections and a global eustatic curve can be derived from local sea-level curves, using a rheologically appropriate Earth model to predict surface deformation in response to changing ice and water loads, with the ocean surface remaining a gravitational equipotential surface at all times. The sea-level signal at sites far from the former ice margins approximates the equivalent sea-level function to about 10–15 per cent and the isostatic contribution is mainly from water loading, which is insensitive to the details of the ice sheets, provided that the total ice volumes are correct to within about 10 to 20 per cent. Hence, through an iterative procedure, it becomes possible to estimate changes in ice volumes V_i from observed sea-level changes $\Delta\zeta_{rsl\ (obs)}$, using (2), and (5), below:

$$\Delta\zeta_{rsl} = \Delta\zeta_{rsl\ (obs)} - (\Delta\zeta_i + \Delta\zeta_w) \tag{5}$$

The use of local sea-level curves from widely separated places allows Earth rheology parameters and models of ice distribution to be evaluated. Recent models include deformation of the basins over time, movement of grounded ice across the shelves, modification of sea level by the time-dependent gravitational attraction between the solid Earth, ocean, and ice, and the effect of glacially induced changes in Earth rotation on sea level.

Sea level through the last glacial cycle

Sea-level data for the last glacial cycle are more plentiful than for earlier periods and, at Huon Peninsula, Papua New Guinea, provide a near-complete relative sea-level curve (Figure 142b), which has been used for reconstructing ice-equivalent sea level for the past 140,000 years (Lambeck and Chappell 2001). Results indicate that ice melting has varied since the Last Glacial Maximum (LGM), with two periods of rapid sea-level rise from ~16,000 to 12,500 and from 11,500 to 8,000 years ago, separated by the Younger Dryas (YD) cold episode when sea level seems to have risen less rapidly. By 7,000 years ago, the northern ice sheets except for Greenland had gone and ocean volume approached its present level, but Antarctic ice melting may have since contributed a few metres of equivalent sea level.

The Last Interglacial, when sea level was similar to the present, ended about 118,000 years ago with rapidly falling sea level, associated with the growth of northern continental ice. Cyclic sea-level changes from 118,000 to ~60,000 years reflect the effect of the 20,000-year orbital precession cycle on the ice sheets, although other fluctuations also appear. These occur repeatedly about every 6,000 years from ~60,000 to 30,000 years, with amplitudes of 10–15 m, and each rise apparently coincides with a major episode of ice-rafted sediment deposition recorded in the North Atlantic, suggesting that the rise was caused by a large, rapid discharge of continental ice.

Oxygen isotopes and long sea-level records

Long, continuous records of oxygen isotopes in calcareous foraminifera preserved in deep-sea sediments have become standard records of Quaternary sea-level and temperature changes. The oxygen isotope ratio in foraminifera, conventionally expressed as $\delta^{18}O$ (the ‰ difference of $^{18}O/^{16}O$ in a sample from an international standard), depends on the temperature and isotope ratio of the seawater in which they lived. Furthermore, the seawater isotopic ratio is related to the size of polar ice sheets, because ^{18}O is preferentially removed from atmospheric water vapour as it makes its way poleward, causing the icecaps to be depleted in ^{18}O relative to seawater by ~25–55 ‰, varying with atmospheric temperature. Thus, foraminiferal isotopes $\Delta\delta^{18}O_f$ respond to ice volume changes ΔV_i according to

$$\Delta\delta^{18}O_f \approx \delta^{18}O_i\ \Delta\ V_i/V_w + C_T\ \Delta T + \varepsilon \tag{6}$$

where ΔV_i is relative to present day V_i, V_w is the present volume of seawater (mean $\delta^{18}O$ of modern seawater = 0 ‰$_{SMOW}$), C_T is the coefficient of temperature-dependent isotope fractionation for calcite (-0.23 ‰ $^{o}C^{-1}$), ΔT is temperature change, and ε represents any local change of seawater $\delta^{18}O$ not related to ΔV_i. (Equation 6 is approximate because the mean isotopic composition of the ice sheets $\delta^{18}O_i$ is assumed constant, but the uncertainty here probably is smaller than the effects of ice volume and temperature.) Finally, the first term on the RHS of (6) can be expressed in terms of equivalent sea level $\Delta\zeta_e$: comparison of isotopes and sea levels for the last glacial cycle indicates that $\Delta\delta^{18}O_w/\Delta\zeta_e \sim -0.009$ ‰ m^{-1}.

The numerical value of $\delta^{18}O_f$ becomes increasingly positive under a fall of both sea level and temperature, which typically happens under a major ice-sheet advance. Hence, marine oxygen isotopes provide composite records of sea level and temperature: for example, Figure 145 illustrates typically close correspondence between sea level and isotopic cycles. In a longer time frame, Tertiary isotope records reveal both progressive cooling and the onset of ice-driven sea-level cycles

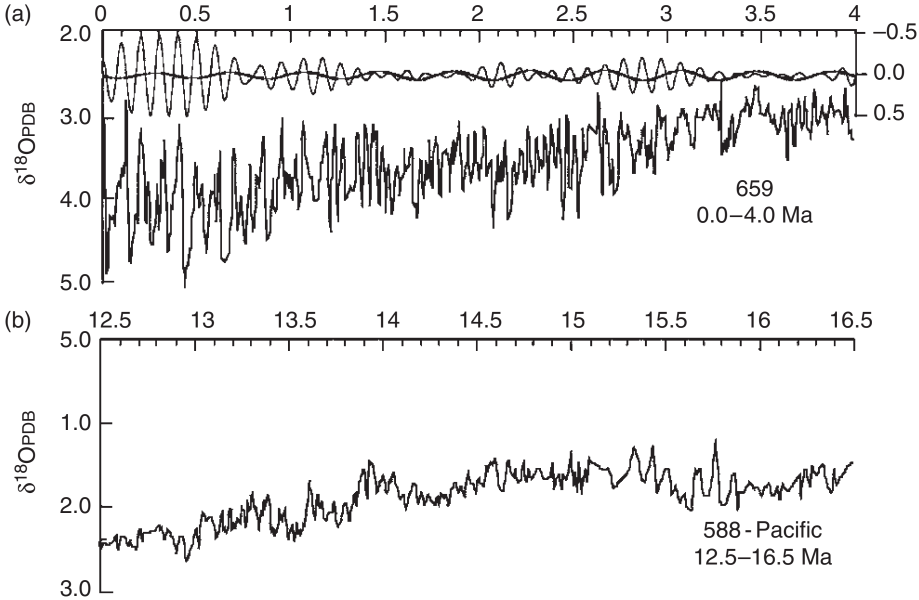

Figure 146 Contrasting records of marine oxygen isotopes from 0–4 Myr (Plio-Pleistocene) and 12.5–16.5 Myr (Miocene). The increase of amplitude in δ^{18}O cycles around 2.6 Myr represents the onset of large fluctuations of ice sheets and sea level, which become even more pronounced around 1 Myr, when they adopt a characteristic period of ~100,000 years. The upper curve in (a) shows the changing amplitude of the 100,000 year cycle (after Zachos *et al.* 2001)

in the Pliocene (Figure 146). As the history of ocean temperature becomes increasingly well determined, through trace-element analysis of microfossils and other techniques, the sea-level contribution in such records is now being isolated (see Zachos *et al.* 2001).

Shoreline reconstructions

Once a global eustatic curve $\zeta_e(t)$ is established, the course of shoreline changes through time can be predicted. Provided that the present-day shallow water bathymetry is known with high resolution, water depths for any region at any time within the range of the eustatic curve follow from (4), and the palaeo-shorelines at time t correspond to the contours $h(\varphi, t) = 0$. Thus, it becomes possible to examine the migrations of shorelines through time for intervals for which sufficient observational data exist to constrain the isostatic variables. Predictions of shorelines since the time of the LGM have been published for both global and regional reconstructions, which can provide useful insights into the interpretation of prehistoric sites. As Lambeck (1996) has shown, for example, the interpretation of post-Palaeolithic archaeology of the Aegean is intimately linked to reconstructions of shoreline changes, which were a powerful factor in trade and changing human activities in the region.

Geomorphic consequences of sea-level changes

Rising sea level at the end of the Pleistocene Period led to widespread geomorphologic changes in coastal regions. Under the influence of ice-age low sea levels, today's coastal valleys tended to become incised well below present sea level, only to turn into traps for floodplain aggradation when sea level rose, as ice sheets retreated. The alternation of incision and aggradation doubtless was repeated in each glacial cycle, throughout the Quaternary, leading to development of broad floodplains on thick sediment valley-fills, which in tectonically stable regions often extend hundreds of kilometres inland. Near-coastal limestone regions under the influence of low sea levels developed vadose karst systems below present sea level, including stream passages and speleothem formations. At coastlines, a range of landforms were the result of the postglacial sea-level rise, from drowned valleys through sand-barrier basins to estuarine and deltaic plains, the particular form depending on sediment supply, tidal range and wave energy. In tropical seas, coral reefs re-established after being reduced by subaerial erosion during glacial-age lowstands.

In tectonically stable terrain, the geomorphic expression of any given Quaternary sea-level cycle tends to overprint the record of earlier cycles. In contrast, tectonic uplift results in flights of coastal terraces built at times of high relative sea level (e.g. Figure 144) that often pass inland into flights of river terraces. However, even though a terraced river valley may grade to present sea-level, not all its terraces necessarily relate to sea level high-stands. Where continental shelves are broad, a river carrying sufficient fluvial sediment, when sea level was low and the coast far seawards of its present position, may have aggraded to a level higher than the recent floodplain, which thus becomes inset below a Pleistocene lowstand terrace. Given the tendency for all such features to become degraded and overprinted, the geomorphic legacy of past sea levels – while very sharp for recent events – becomes increasingly blurred with time. In contrast, the sedimentary record remains relatively intact.

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SEAFLOOR SPREADING

Seafloor spreading is the process by which new oceanic crust is generated at mid-ocean ridges, the long linear belts of elevated seafloor that lie at the centres of most ocean basins (see SUBMARINE LANDSLIDE GEOMORPHOLOGY). The term was coined by Dietz (1961) for the idea (jointly proposed with Hess 1962) that new crust is formed by magmatic intrusion along the crests of mid-ocean ridges, and that conveyor belts of crust thus formed move symmetrically away from the ridges, driven by convection currents in the underlying mantle. Their suggestion provided the first viable mechanism by which continental drift might take place. Acceptance of the concept came only after the demonstration by Vine and Matthews (1963) that linear magnetic anomalies, which are symmetrical on either side of a mid-ocean ridge, record reversals of the Earth's magnetic field as newly formed crust cools and moves laterally away from the axis. The seafloor spreading hypothesis was the direct catalyst for the development of the concept of PLATE TECTONICS, the grand unifying theory of Earth sciences. This posits that the Earth is capped by a number of rigid plates, containing both continental and oceanic crust, that deform only at plate boundaries.

The large-scale motion of the Earth's plates is now believed to be driven primarily by the pull of dense slabs of oceanic lithosphere as they are subducted into the mantle at convergent margins (see WILSON CYCLE). Seafloor spreading at divergent oceanic plate boundaries may therefore be regarded as an essentially passive process: plates are pulled rather than pushed apart. Separation of the oceanic plates induces upwelling of the convecting, plastically deforming mantle asthenosphere. As it rises up and decompresses the mantle peridotite starts to melt. This generates a basaltic liquid which separates from its host and rises upward to feed a magma chamber beneath the ridge axis, thence solidifying to form ocean crust.

The rate at which seafloor spreading occurs varies from place to place along the 60,000 km long global mid-ocean ridge system, from less than 1 cm yr^{-1} at the Gakkel Ridge (in the Arctic Ocean) and parts of the Southwest Indian Ridge, to 16 cm yr^{-1} at the East Pacific Rise (west of Peru). The oldest surviving ocean floor lies in the north-west Pacific Ocean and is of Jurassic age (approximately 180 Ma).

Seismic refraction experiments indicate that the ocean crust is generally 6–7 km thick and has a layered internal structure. A seismic discontinuity (the Mohorovicic discontinuity or 'Moho') separates it from rocks with typical mantle velocities below. The seismic layering has been correlated directly with the lithological layering observed in ophiolites, which are regarded as on-land fragments of ocean crust and shallow mantle. This has led to the conventional view of ocean crustal structure as a simple, uniform 'layer-cake' internal structure composed (from top to bottom) of: basaltic lavas, usually with lobate, pillowed morphology indicative of submarine eruption; a parallel 'sheeted' swarm of dolerite dykes that fed the lava flows; and coarse-grained gabbros and ultramafic (olivine-rich) plutonic rocks that crystallized slowly in the magma chamber.

In early models for the generation of ocean crust these magma reservoirs were envisaged as huge bodies the height of the entire lower crust and tens of kilometres wide beneath spreading axes. However, seismic reflection experiments at the fast-spreading East Pacific Rise in the late 1980s showed that no such magma body exists: purely molten material is instead restricted to a thin lens (probably of the order of 100 m thick) in the middle crust, overlying a much larger region in the lower crust that appears to be a partially molten mush composed of crystals with small amounts of interstitial melt. At slower spreading rifts such as the Mid-Atlantic Ridge even this thin magma lens is normally absent, implying that the lower crust there probably freezes solid in between melt delivery events from the mantle, and that magma supply in these environments may be relatively reduced (e.g. Sinton and Detrick 1992).

Ridges may be offset by several hundred kilometres by transform faults, across which lateral motion occurs as seafloor spreads in opposite directions on either side. Smaller scale offsets or discontinuities of the spreading axis between transform faults define the boundaries between individual spreading segments, which may be regarded as elongate volcanoes more or less aligned along the ridge crest (Macdonald *et al.* 1988).

The morphology of the region around the ridge crest is very much dependent upon spreading rate. Whereas fast-spreading ridges are characterized by relatively smooth seafloor with an elevated ridge crest, slower spreading ridges (less than ~5 cm yr^{-1}) are instead marked by a very rough seafloor and an axial valley that may be more than 2 km deeper than the surrounding walls. Mantle peridotite, now altered by the action of seawater to serpentinite, is commonly recovered in these axial valleys. The rough seafloor results from extensional faulting, which helps to accommodate separation of the plates when magma supply to the ridge is low. Some of the fault planes have very low dip angles and can accommodate tens of kilometres (in excess of a million years' worth) of displacement on a single structure. These structures, termed 'detachment faults', provide a mechanism by which mantle and lower crustal rocks may be exhumed onto the seafloor. Mantle rocks may also be exposed at some slow-spreading ridges because magma supply to the ridge axis was so low that a continuous layer-cake magmatic crustal layer was never produced in the first place (Cannat 1993). In places, therefore, the seismically defined crustal layer may be composed partly or completely of serpentinite, which has a velocity much lower than fresh peridotite but similar to that of basalt or gabbro. Ocean crustal structure, particularly at slow-spreading ridges, is now understood to be far more heterogeneous than originally suggested on the basis of the seismic refraction experiments. Fast-spread ocean crust may have a more regular 'layer-cake' architecture.

Total crustal production by seafloor spreading at mid-ocean ridges is estimated at 18 $km^3\,yr^{-1}$, generating a thermal flux equivalent to ~50 megawatts per kilometre of spreading ridge worldwide. This heat is extracted from the newly formed crust primarily by hydrothermal convection: seawater descends through cracks in the crust, heats up and is eventually vented back into the water column as 'black smoker' fluid. This fluid is hot (up to 400 °C) and rich in metals that have been stripped from the basaltic crust. Volumes are such that the equivalent of the entire world ocean is believed to circulate through the crust every seven million years or so. Hydrothermal circulation therefore plays an essential role in regulating the composition of seawater on geological timescales.

Fluid circulation and cooling persists away from the ridge crests. It causes the lithosphere – the mechanically rigid plate – to increase progressively in thickness as it moves away from the ridge. As the uppermost mantle cools below ~1,000 °C it ceases to be able to flow in the convecting asthenosphere and moves with the overlying crust, in effect being attached or welded to the base of the ocean crust. Mature ocean lithosphere is up to 100 km thick, all but the uppermost few kilometres being rigid mantle.

On a broad scale the ridge crest and surrounding seafloor is elevated relative to the abyssal plains because of the lower density of hot, partially molten asthenospheric mantle flowing upward beneath the ridge crest. Moving the lithosphere away from the region of upwelling and thickening it causes the seafloor to subside (see ISOSTASY). The rate of subsidence is proportional to the square root of the age of the crust and is independent of spreading rate, explaining why the width of the elevated region around the East Pacific Rise is far broader than that around the Mid-Atlantic Ridge.

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CHRIS MACLEOD

SEDIMENT BUDGET

A sediment budget is a quantitative accounting of the rates of production, transport and discharge of detritus in a geomorphic system such as a DRAINAGE BASIN, coastal BEACH, offshore zone, hillslope, river channel, GLACIER, or any landscape unit around which boundaries can be drawn. Sediment budgets apply the principle of conservation of mass to geomorphic systems, an approach that became popular in the 1970s. Sediment budgets provide a tool for research geomorphologists to judge the relative importance of sediment sources, storage sites and transfer processes, including how they change over time. Sediment budgets are also useful tools for resource management when it becomes important to distinguish human impacts on geomorphic systems from those that would have occurred without human interference (Reid and Dunne 1996). Studies have been reported that use sediment budgets to document the effect of agriculture, forestry, road construction, urbanization, DAMs, wildfires and mining. Sediment budgets have been used to construct a more complete picture of the distribution of sediment sources within a drainage basin (e.g. Marston and Dolan 1999) and the heavy metals that can be associated with the sediment (e.g. Marcus *et al.* 1993).

Four basic steps must be followed to construct a sediment budget (Lehre 1982): (1) define the boundaries of the geomorphic system; (2) identify the processes and sites of EROSION, transport and storage (deposition) in the geomorphic system, including the linkages between them; (3) quantify the contribution of each over space and time; and (4) set up an accounting sheet that balances the sediment production, sediment yield (see SEDIMENT LOAD AND YIELD), and storage. The first step depends on the ability to recognize the boundaries of the geomorphic system to be studied. Most sediment budgets have been prepared for drainage basins and sandy beaches, both of which have readily recognized boundaries. However, sediment budgets have also been attempted for KARST and glacier systems, where subsurface passages transport significant amounts of sediment that is difficult to trace and measure. Sediment budgets generated by wind in the form of sandstorms and DUST STORMs have been rare, although this component has been used to explain the discrepancies between inputs, change of storage and outputs – a dangerous practice unless all transfer and storage components have been measured with great accuracy (Hill *et al.* 1998). The second step requires training, experience and expertise with the full range of field, lab and office techniques in geomorphology (e.g. aerial photography and historical maps) to discern evidence of the range of possible processes and storage sites. A flow chart may help to visualize the processes, storage sites, outputs of sediment and linkages between them (Figure 147). The third step requires measurement of these processes using an appropriate choice of techniques. Direct measurement of processes is beset by problems, although controlled field experiments can help. In some cases, ^{137}Cs measurements have been used in combination with more traditional techniques to compile tracer-based sediment budgets (e.g. Walling *et al.* 2002). Geomorphologists might also use predictive equations, examine soil profiles, undertake photogrammetric measurements from aerial photographs, or conduct lab measurements of soil and sediment characteristics. Changes in sediment storage can be derived from geophysical surveys, morphometric modelling or field surveys using anchor chains.

One of the classic sediment budget studies in understanding the importance of sediment storage was undertaken by Trimble (1983). He compiled a sediment budget for the 360 km^2 Coon Creek watershed in Wisconsin for two periods: 1853–1938 when soil erosion was severe because of poor land management, and 1938–1975 following the implementation of SOIL CONSERVATION practices. Sediment yield at the mouth of the WATERSHED was essentially identical for the two periods, from which one might conclude that soil conservation practices failed to have the desired effect. However, when one examines the sediment budget data, it becomes apparent that upland erosion had been reduced by 26 per cent. Sediment yield figures remained high even after soil conservation measures had been implemented because sediment that had been stored in tributary valleys and the upper main valley during the early period had been mobilized during the latter period. This removal from storage offset the reduction in hillslope erosion, with the net effect that overall sediment yields remained essentially unchanged. One of the main lessons of erosion studies has been learned from Trimble's study: sediment yield values alone are not a good indicator of spatial and temporal changes in erosion within a watershed.

Reneau and Dietrich (1991) used a sediment budget approach in the southern Oregon Coast Range to demonstrate that a steady state condition exists between sediment production from hillslopes and sediment yield from the watershed. The authors suggest that erosion rates were spatially uniform and perhaps even diagnostic of a landscape in geomorphic equilibrium. In earlier work elsewhere in the Oregon Coast Range, Dietrich and Dunne (1978) found that the rate that bedrock in the Rock Creek watershed was converted to soil, with the associated increase in volume by a factor of 1.17, was equivalent to the rate of DENUDATION calculated for the watershed over the same time interval. Thus, they conclude that reduction in watershed relief by denudation was balanced by the increase in volume of hillside materials; relief is not changing over time, without accounting for tectonics.

Graf (1994) compiled a fluvial sediment budget for the northern Rio Grande drainage system in New Mexico for the express purpose of tracking deposits likely to contain plutonium. Early development of nuclear weapons at Los Alamos National Laboratory had led to discharge of highly concentrated plutonium wastes to arroyos that had carried the plutonium adsorbed onto sediment into one of the major waterways of western North America. Graf coupled extensive and detailed field mapping with simulation modelling of plutonium movement to demonstrate the importance of large FLOODs in delivering pulses of plutonium downriver. Fortunately, the concentrations of plutonium in stored sediments are generally low, but do vary considerably. With large floods, concentrations will decline over time to approximate levels from atmospheric fallout; without large floods, the portion of plutonium that is associated with bedload will persist. Graf was also able to demonstrate the significance of plutonium storage behind Cochiti Reservoir along the Rio Grande, a site that is now closed for fishing and swimming.

Madej (1987) showed how the residence time of stored sediment could be calculated as part of fluvial sediment budget studies if one knows the sediment storage, expressed as cubic metres per metre length of valley sediment, and the sediment transport rate, expressed in units of cubic metres per year. When the storage is divided by the transport rate, the residence time is derived, in units of years per cubic metre. This strategy was used by Madej to calculate the residence time of sediment in various reaches of Redwood Creek in north-west California. Comparing the residence time to the rate of sediment delivery allows one to determine whether the system is at/near geomorphic equilibrium or likely to experience progressive change over time. If a reach receives

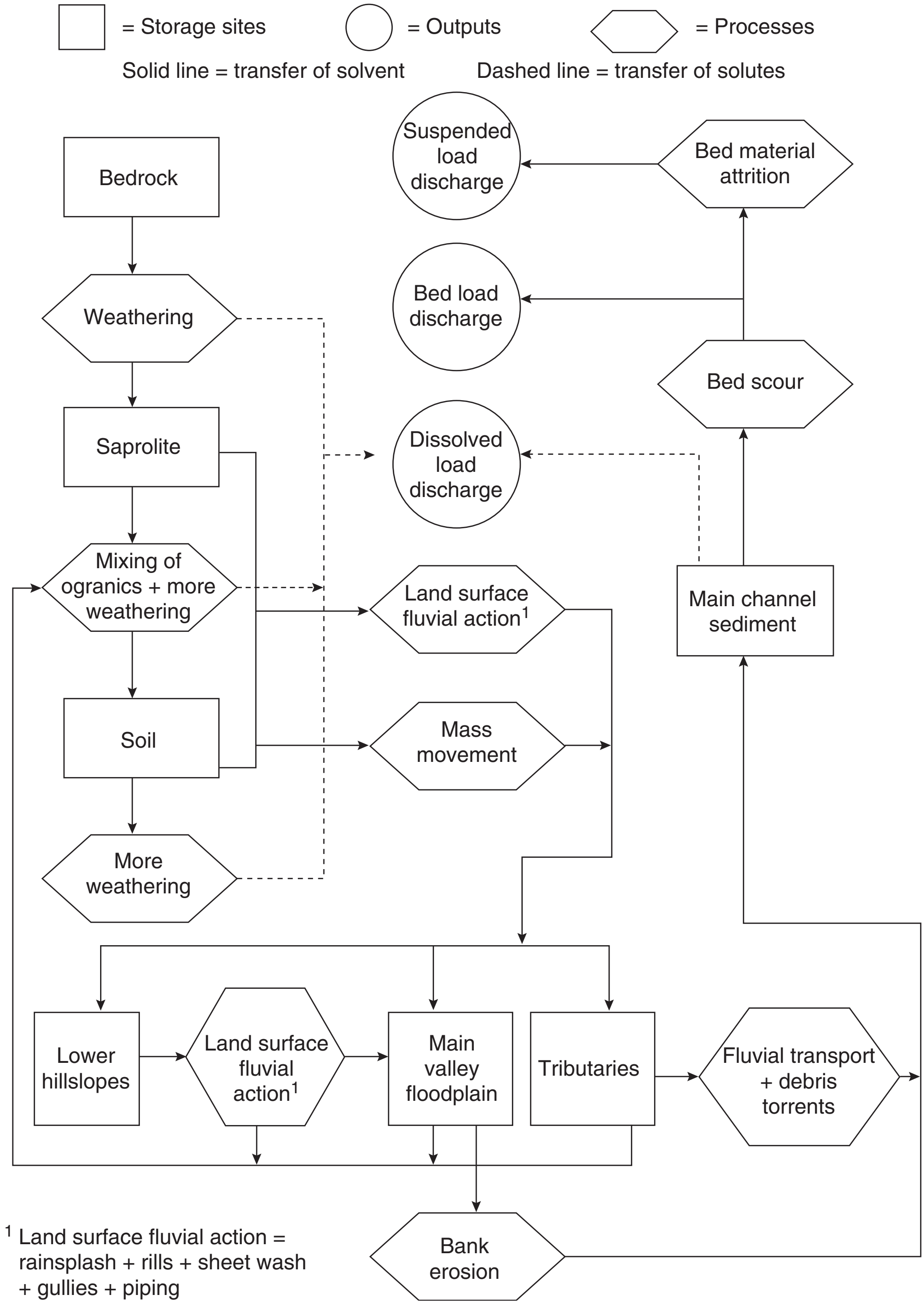

Figure 147 Flow chart for a drainage basin sediment budget (after Dietrich and Dunne 1978)

a volume of sediment from storms with a recurrence interval of fifty years and that sediment is likely to reside in the stream for a hundred years, the stream will experience progressive AGGRADATION. The recovery of Redwood Creek from catastrophic floods and associated MASS MOVEMENT was found to vary depending on the residence time, which in turn varies with landscape position in the valley floor. Indeed, sediment budgets have been used to compare the importance of frequent, low magnitude events with infrequent, catastrophic events in the transport of sediment (e.g. Springer *et al*. 2001).

Sediment budgets have proved especially useful in examining changes in beaches over time. Consider a sediment budget for a sandy beach between two rocky headlands. Sediment sources could include eroding cliffs (see CLIFF, COASTAL), onshore transport, marine erosion of beach material, supply from dunes (see DUNE, COASTAL), subsurface erosion, fluvial input, BEACH NOURISHMENT and LONGSHORE (LITTORAL) DRIFT. Sediment storage could occur on the beach and adjacent inland dunes, as well as in offshore banks, SPITs and bars (see BAR, COASTAL). Sand could be lost through offshore transport, longshore drift, aeolian erosion (see AEOLIAN PROCESSES), and dredging. The effects of various shoreline protection measures can be effectively measured with this approach (Cooper *et al*. 2001). One noteworthy study demonstrated that the loss of sand along beaches of southern California could be attributed to the construction of sediment detention basins in the mountains surrounding the Los Angeles Basin. When it was discovered that valuable recreational beaches were being deprived of river-delivered sand and diminishing in size, artificial nourishment was required at great expense (Cooke 1984).

Sediment budgets are utilized over a wide range of spatial scales, from small plot studies (Duijsings 1987) to continental-scale assessments. New techniques are being developed to construct more accurate sediment budgets. The sediment budget approach is being applied to an ever-expanding variety of geomorphic environments and for discerning the effects of human activities in these settings. Because sediment budgets deal with the sources, storage, throughflow and outputs of sediment in a geomorphic system, they have been characterized as a fundamental method in understanding cascading process systems.

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SEDIMENT CELL

A *sediment cell* is a section of the coastal zone where the sediment inputs, throughput and outputs may be considered part of a closed system. Given the vast range of coastal environments and the nature and sources of sediments, a wide range of sediment cells exist, both in form, and temporal and spatial scale. A more general definition is, therefore, a coastal environment where the input, accumulation and output of sediments are part of an interrelated flow of sediments, some of which may be derived from, and/or exported to, adjacent sediment cells.

Komar (1998) discusses sediment cells using the concept of coastal SEDIMENT BUDGETs, based on the example of the Southern California littoral cells. Each cell has a sediment source (river, cliff erosion, etc.), longshore transport, and finally sinks which include submarine canyons, loss to dunes and transport to downdrift cells. Each cell therefore contains a cycle of sediment input, transport and sedimentation, the latter either as accumulation within the cell, or loss onshore, offshore or longshore. He presents a more generalized budget of littoral sediments which includes sediment credit, debit and a balance, the latter resulting in positive beach deposition or negative beach erosion.

Once established cells can undergo *temporal variation* in response to changes in the boundary conditions. This can include a change in the sediment budget, as has occurred on many coasts following the postglacial marine transgression, as shelf sediment supplies have become exhausted. It could also be human induced through construction of dams or shoreline structures which interrupt or stop sediment supply. It can also be produced by changes in the driving processes, such as wave climate, wave refraction and climatic conditions, all of which can change both the magnitude and direction of sediment transport within the cell.

Cells can also reach *equilibrium*, when there is essentially zero sediment transport and zero change to the sediment budget over a given time span. This can occur on swash aligned beaches where the wave crest is in equilibrium with the shoreline alignment. On a drift or current aligned shore it would infer zero longshore sediment transport, which is possible but unlikely. On a graded shore the sediment texture is arranged so that it is in equilibrium with the prevailing processes so that no further regrading is required. Swash aligned beaches may take the form of a log spiral (see LOG SPIRAL BEACH) or zeta shape, in locations where waves are refracted around a headland and the shoreline aligns to the spiral of the refracted wave crest. So long as there is no littoral drift they may be considered an equilibrium or closed cell. However on many coasts such systems do release sediment downdrift through pulse boundaries.

Davies (1980) approaches sediment cells from the concept of a sediment store, which has input, throughput and outputs, as well as internal biological supply and loss from attrition. The nature of the store or cell is dependent on the nature, scale and rate of the boundary components and internal dynamics.

Carter (1988) provides the most advanced treatment of what he calls a ‘coastal cell’, within which is a recognizable compartmentalization of the sediment budget. He goes on to say that this is easy to identify where the cell is restricted to a bay, estuary, river mouth or pocket beach, but many coastal cells have leakages longshore to adjacent systems, onshore to dunes and estuaries or offshore to the inner continental shelf. The cell boundaries may therefore be free or fixed (Table 42). *Fixed* boundaries include impermeable morphological structures such as headlands, shoals, inlets, river mouths. *Free* boundaries are more transparent and therefore less recognizable, as they may result from a change in wave field and direction of transport, rather than a distinctive morphological feature. Both boundary types may either cause the cells to ‘divide’, ‘meet’ or ‘pulse’. The *divide* boundary occurs at the updrift limit of shoreline erosion, while the *meet* boundary occurs at the downdrift limit of sediment deposition. The *pulse* boundary permits sediment exchange between cells.

Sediment cells can also vary considerably in size and relation to neigbouring cells. Cells may be either independent (e.g. small pocket beach) with fixed boundaries and no leakages, or be nested or cascaded within a series of interconnected (leaking) subcells (e.g. deltaic systems or series of interconnected beaches).

Sediment cells can be identified as a *morphological* unit, particularly when they have fixed boundaries, e.g. an embayed or pocket beach or estuary. They can also be identified based on *sediment texture*, either through the presence of one sediment type, e.g. a coastal dunefield, or

Table 42 Sediment cell definitions and budget

Sediment cell – a closed and balanced sediment budget		
Boundaries		
fixed	morphological structure (e.g. headland, inlet, river mouth, foreland)	
free	dynamic divide induced by processes, e.g. change in direction of littoral drift	
Boundary types		
divide–		updrift, limit of erosion or sources of sediment
meet–		downdrift, limit of sediment deposition, sink
pulse–		leakage and exchange across boundary
Sediment inputs		
External		
		Terrestrial – river supply, cliff erosion
		Biological – carbonate detritus
		Updrift longshore transport – from neighbouring cell
		Headland bypassing via dunes or subaqueous sand pulses
		Onshore transport from inner shelf, esp. during sea level transgression
Internal		
		Biogenic production
		Chemical production, e.g. ooids and cements
		Beach nourishment
Sediment outputs		
Onshore		barrier – dunes
		– overwashing
		estuary infilling (flood tide deltas)
Longshore		longshore sediment transport
Offshore		inner shelf sand bodies
		submarine canyons
Internal		attrition, solution, cementation (e.g. beachrock)
		beach mining
Sediment balance		
Inputs–outputs = cell erosion, deposition or stable		

Sources: After Komar (1998), Davies (1980) and Carter (1988)

a gradation in the sediment texture resulting from selected longshore transport of size factions. Finally they can be defined by the rates and scale of sediment *transport* within the system, with the boundaries having little or no transport.

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SEE ALSO: log spiral beach; longshore (littoral) drift; sediment budget

ANDREW D. SHORT

SEDIMENT DELIVERY RATIO

The rate of sediment yield at a specified point in a channel network, expressed as a fraction of the rate of erosion in the contributing catchment, is termed the sediment delivery ratio. Sediment delivery ratios are widely used to adjust SOIL EROSION estimates to account for deposition of sediment as it is transported from its point of origin to, and through, the stream network.

Sediment delivery ratios have been widely used to estimate stream sediment loads from erosion rates predicted by the Universal Soil Loss Equation (USLE) and its successor, the Revised USLE (RUSLE). These empirical equations are designed to predict gross rates of erosion at the soil surface. Because the USLE and RUSLE were developed from studies of small test plots, they define 'erosion' as the movement of soil particles from one location to another – but, importantly, not necessarily from their point of origin to a stream channel. A fraction of the sediment mobilized by surface erosion will be intercepted (for example, in densely vegetated zones or low-gradient footslopes) before it reaches the channel network. Of the sediment that reaches the channel network, a further fraction will be deposited on the floodplain or stored in the channel. The proportion that is delivered to a sampling point in the channel network – rather than intercepted on the soil surface, deposited on the floodplain, or stored in the channel – is the sediment delivery ratio.

Sediment delivery ratios are commonly estimated from the measured sediment yield (from sediment gauging methods or accumulation in a sediment trap) at a given point in the channel network. This is then divided by the estimated rate of erosion in the surrounding catchment (derived from the USLE/RUSLE or, in some cases, direct field measurements). Thus sediment delivery ratios will not only reflect sediment interception, storage and deposition, but will also reflect any errors made in estimating sediment yields or rates of surface erosion; both are subject to significant uncertainties (Meade 1988; Trimble and Crosson 2000).

Sediment delivery ratios reported in the literature range from over 100 per cent to less than 1 per cent. This variability arises from differences in geomorphic characteristics between catchments, as well as from variations in erosion rates and sediment yields through time at any individual site. Sediment delivery is often highly episodic, and measurements of sediment yield – even when averaged over decades – can be significantly higher or lower than long-term rates of sediment supply to the channel network (e.g. Clapp *et al*. 2000; Kirchner *et al*. 2001).

Nonetheless, systematic relationships have been observed between sediment delivery ratios and catchment morphology and processes. Sediment delivery ratios tend to be higher in catchments where channel slopes and valley sideslopes are steep, and where relief and drainage density are high. Conversely, sediment delivery ratios tend to be lower where sediment sources are far from channels, or are separated from them by sediment-trapping zones (typically characterized by low gradients and dense vegetation). Sediment delivery ratios also tend to be lower where sheet and rill erosion predominate, and higher where gully erosion predominates, because gullies tend to be more directly connected to the channel network.

Sediment delivery ratios also generally decrease as drainage area increases, ranging from roughly 30–100 per cent in 0.1 km^2 catchments to roughly 2–20 per cent in 1,000 km^2 catchments (e.g. Novotny and Olem 1994). This is consistent with the fact that as one moves downstream, channel and valley gradients typically become gentler and floodplains and footslopes typically become wider. All these trends provide greater opportunities for sediment storage, both on hillslopes and in the fluvial system. As one might expect, the lowest sediment delivery ratios are typically observed where rivers emerge from steep mountain fronts and flow out across broad depositional basins.

Sediment yield predictions are often generated by combining USLE or RUSLE estimates of surface erosion with sediment delivery ratios plucked from the literature. This approach is problematic, because sediment delivery ratios vary widely and are not always consistently defined. The denominator of the sediment delivery ratio is sometimes the gross rate of sediment mobilization on the soil surface (as in the USLE/RUSLE); in this case the ratio reflects sediment interception en route to the channel as well as net deposition and storage in the channel network. Alternatively, the denominator is sometimes the rate of sediment supply to the channel network (excluding sediment interception during overland transport, but including sediment production from channel incision or bank erosion); in this case the sediment delivery ratio reflects only the transmission efficiency of the fluvial system. Because sediment delivery ratios may be conceptually defined or operationally measured differently from one study to the next, they should be interpreted with caution.

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JAMES W. KIRCHNER

SEDIMENT LOAD AND YIELD

Normally, sediment discharge is reported in units of mass per unit time, kilograms per second, megagrams per day or megagrams per year. A megagram is equivalent to a tonne, so frequently loads are reported in tons per day or tons per year. Sediment concentration is reported in milligrams per litre or grams per cubic metre of water. The size of particulate material transported by rivers ranges between fine clay and colloidal particles of less than 0.5 micrometres in diameter to large boulders moved during flood events. A distinction is commonly made between bedload and suspended load. This is a distinction based on mode of transport. Bedload consists of the coarser sediment particles that roll, hop or slide along the bed, with more or less continuous contact with the bed, while the suspended load consists of sediment particles that are supported in the flow by the upward components of turbulent currents. The total sediment load is the sum of the bedload and the suspended load. Engineers prefer the distinction between bed-material load and wash load where sediment particles that are finer than those contained in the bed are called wash load and sediment particles that are the same size range as those found in the bed are called bed-material load. The implication is that bed-material load is primarily a function of transporting capacity, boundary shear stress or force available to initiate movement, whereas wash load is commonly a function of variable sources of supply. In either case, there is overlap between the two categories. Church (1983) recommends 1 mm as the appropriate boundary value between bedload and wash load; Walling (1988) recommends 0.2 mm as the cutoff between bedload and suspended load.

Fluvial sediment transported as bedload consists almost entirely of inorganic material, except in steep, forested watersheds where large organic debris is a highly significant component. The inorganic bedload normally resembles the local bedrock in terms of mineral composition except in glacially disturbed watersheds. The finer sediment transported in suspension commonly incorporates a proportion of organic material. The mineralogy of these sediments may bear little resemblance to the local bedrock because of selective detachment and transport processes, chemical weathering processes which disintegrate the rock and secondary redistributed glacigenic sediments. Also the size distribution of the suspended sediment will be considerably enriched in clay-sized particles and organic matter when compared with the source material. Enrichment in fine material has important implications for transport of contaminants which tend to be adsorbed onto finer particles as these finer particles have greater specific surface areas and cation exchange capacities.

Sediment transport capacity load or non-capacity load is a classification of the load of sediment being carried by a stream. If a stream has excess or unsatisfied capacity it is said to be carrying a non-capacity load; if it is carrying as much material as available stream energy permits, it is carrying a capacity load. The enormous capacity for the transport of fine sediment is exemplified by rivers carrying hyperconcentrated flows. Almost all streams carry a non-capacity load of fine sediments. For particles larger than some critical diameter, streams carry a capacity load. A commonly accepted critical particle diameter is around the silt/sand boundary (or about 0.063 mm).

Sediment transport formulae are still evolving, especially bedload formulae. Bedload transport commences when the rate of water flow reaches a magnitude adequate to exert a critical tractive force. Du Boys introduced a formula for movement of bedload of the form

$$Q_b = C_b \cdot \tau(\tau - \tau_c)$$

where Q_b = bedload transport per unit width per unit time, C_b = a special sediment parameter, τ = shear stress, and τ_c = critical shear stress for start of material movement.

Suspended load transport can be considered as an advanced stage of bedload transport by which particles in saltation are caught by the upward component of the turbulent velocity and are kept in suspension. The fundamental equation of sediment suspension by fluid turbulence is

$$c\omega = -\beta \cdot \varepsilon \cdot dc/d\gamma$$

where c = concentration of suspended load in dry weight per unit volume, ω = settling velocity of the particles, β = sediment transfer coefficient, ε = exchange coefficient for suspended load, and γ = vertical distance from the stream bed.

There are problems of application because ε varies through the vertical, and the expression for the variation in suspended sediment concentration with depth is fairly complex. Nevertheless, suspended sediment load can be predicted more accurately than the bedload.

It is also possible to combine bedload and suspended load transport formulae in one expression for total load. The fundamental reference is Einstein (1950) but there are numerous papers still appearing in *Water Resources Research* debating improvements in these formulae.

Milliman and Syvitski (1992) admit that no one knows exactly how much sediment is discharged to the ocean but they suggest that it is probably of the order of 20 billion tonnes per year. This is 50 per cent higher than the estimate by Milliman and Meade (1983) and is accounted for by the fact that the contribution of small mountainous rivers was overlooked in the earlier estimate. On the other hand, it seems probable that, because of the proliferation of dam construction during the second half of the twentieth century, that a smaller total is in order; they also note that an unknown amount is stored in subaerial parts of sinking deltas. The suggestion is that prior to widespread farming and deforestation, the export of sediment was about 10 billion tonnes per year.

Sediment yield is defined as the total mass of particulate material reaching the outlet of a drainage basin per year. Specific sediment yield is the sediment yield per unit area and, making suitable adjustments for sediment specific gravity and density of packing, this figure can be converted into a depth of sediment removed from the whole surface of the basin per year. A unit of measurement known as the Bubnoff (Fischer 1969), which is equivalent to 1 mm of denudation per 1,000 years has been found to be useful for regional comparisons, though it should always be borne in mind that variable thicknesses of sediment are removed from different parts of each basin and the Bubnoff is only an average statistic. Documented values range from 0.09 to over 13,000 Bubnoffs (0.26 to 36,000 $\mathrm{t\,km^{-2}}$). The two extreme cases are from Finland and Taiwan respectively. The total sediment yield should include both bedload and suspended load, but only rarely are measurements of bedload available. In such a case, 10 per cent is normally added to the suspended load. In steep mountainous basins, this is clearly an underestimate and in lowland rivers, this is an overestimate.

Where the sediment load of a river is deposited in a lake or reservoir it may be possible to estimate the total yield quite accurately. The sediment yield from a drainage basin is commonly only a small proportion of the gross erosion in the basin. Much of the eroded material is deposited before it reaches the outlet and the ratio of sediment yield to gross erosion is termed the sediment delivery ratio. The magnitude of the sediment yield is a function of climate, topography, sediment availability, lithology, vegetation cover and land use.

The sediment budget is defined as a method of accounting of sources, sinks and redistribution pathways of sediments in a unit region over unit time (Swanson *et al.* 1982). The method provides a bridge between studies of fluvial sediment transport and of the associated deposits in storage in the basin. The transport of sediment out of the basin is then seen as a residual term after deposition and storage of the sediment within the basin. In general, the storage term gets larger with increasing size of basin. Sediment on its way from source to outlet of a basin gets side tracked in a number of ways, not only into lake, floodplain and channel storage, but also into other hillslope locations. It is in this context that the concept of virtual velocity of sediment becomes important (Church 2002). Virtual velocity can be defined as the velocity of sediment through a reservoir and is simply the inverse of residence time per metre of reservoir length. Residence time per metre equals the mapped volume of sediment per metre divided by the bedload discharge rate.

It is estimated for the United States that the amount of sediment delivered to the oceans is only about 10 per cent of the total amount eroded off the uplands and 90 per cent is stored between the erosion sites and the sea. They provide illustrative data on the importance of short-term scale, decade to century scale and longer timescales of storage. The short-term pattern on the lower Mississippi River shows sediment being deposited and stored on the river bed at lower flows and being resuspended and flushed out to sea at higher flows. The decade to century scale effect is illustrated by the movement of hydraulic mining debris through the Sacramento River system in California. The slug of sediment (or wave according to Gilbert 1917) created by gold mining during 1855–1885 took almost a century to reach the sea and caused major changes in river bed levels from temporary storage during that period. A classic study by Trimble (1983) examines the long-term effects of sediment storage resulting from forest clearance and accelerated erosion in the uplands of Coon Creek basin in Wisconsin in the 1850s. During 1853–1975, 80 million tons of sediment were eroded off the uplands and delivered to storage sites whereas only 5 million tons were transported out of the basin. In British Columbia, Church and Slaymaker (1989) estimate that sediment stored in paraglacial fans, valley fill and floodplains will require several tens of thousands of years to be transported to the sea.

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SEE ALSO: sediment rating curve; sediment routing

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SEDIMENT RATING CURVE

A sediment rating curve represents the relation between suspended (see SUSPENDED LOAD) sediment concentration (or discharge) and water discharge at a stream measurement station. It is used for estimating suspended sediment discharge averaged over a period of flow record. The relation may concern 'instantaneous' values or may relate average values over (e.g. daily) time intervals.

In theory, suspended sediment concentration should increase with water discharge because the associated increase in turbulence increases the capacity of the river to carry suspended sediment. In practice, the concentration of suspended sediment, particularly the silt-clay fractions, tends to be more influenced by the sediment supply to the channel from hillslope and riparian erosion processes, which can be 'patchy' in space and time. During runoff events, the relation can vary due to sediment exhaustion and also to phase lags between the sediment and water peaks passing the measurement station. The relation can also vary due to seasonal controls on the supplies of sediment and water, to changes in basin land use, or following extreme hydrological events.

Recognizing that there is no unique relation between suspended sediment concentration and water discharge, a sediment rating curve thus aims to model the conditional mean sediment concentration (as a function of water discharge). This is estimated by sampling a series of concurrent measurements of water discharge, Q, and discharge-weighted sediment concentration, C. Ideally, these samples should be collected over a wide range of water discharge, at rising and falling stages, over all seasons, and over many years so that all the factors inducing variance in the relation are represented in an unbiased fashion.

The traditional approach to fitting a rating curve has been to plot C against Q on log-log graphs. These plots accommodate the large ranges of Q and C observed in rivers, the data-scatter tends to be homoscedastic (i.e. independent of discharge), and the underlying relation often shows a simple power form $C = aQ^b$ (a and b are empirical coefficients) which is linear on a log-log plot and easily modelled with linear regression methods. Note that by using log data, the regression procedure models the geometric conditional mean, rather than the desired arithmetic conditional mean, thus a correction factor is required when transforming the rating curve back from log values (Ferguson 1986).

The power law approach should be applied with caution (e.g. Walling and Webb 1988), since large errors can arise because the least-squares curve, while appearing to fit the overall dataset well, may fit poorly at high discharges, and it is these discharges that usually transport the bulk of the sediment load. In such cases, other curve-fitting techniques such as non-linear regression, Locally Weighted Scatterplot Smoothing (LOWESS), or simply segmented curves based on subsets of the data perform better.

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D. MURRAY HICKS

SEDIMENT ROUTING

The process through which sediment is transported downstream following a specific path or route. The sediment is fluvial sediment, which includes both bedload and suspended load. The path or route, of the sediment may be the course of the natural channel, an artificial canal or a restored channel. Sediment routing is closely related to sediment transport. Where sediment transport may be concerned with the details of sediment movement, sediment routing defines the path of that sediment on a channel reach or watershed scale.

Sediment routing models are used in measuring the SEDIMENT BUDGET for a watershed. The routing model quantifies sediment sources and sinks in the watershed, and how much and how fast that sediment is transferred downstream by river reaches. One of the first steps in measuring the sediment budget in a watershed is the creation of a map of the paths travelled by the sediment.

Sediment routing is affected by landscape changes. Where roads are built in forests, there is an immediate effect on the route travelled by the fluvial sediments. Where roads cut across channels, the path of the water and sediment often changes to follow the road instead of continuing down the natural channel. The result may be either deposition on the road surface or increased road surface erosion. Either way, the path of the channel has been changed, altering the sediment routing processes in the watershed.

Artificial canals are built to route sediment through a city without causing damage. Trapezoidal concrete channels route both water and sediment through residential areas without causing flooding or other damage during high flow events. These types of canals are most common where towns are situated next to mountains, for example, Los Angeles, California and Albuquerque, New Mexico. When a large flow event occurs, the canals are filled with material flowing off the mountain side.

Sediment routing is an important consideration when planning a river restoration project, and restorations are often undertaken when the transport of sediment through the channel has been determined to be too slow or too fast, resulting in either erosion or sedimentation of the channel bed. Natural channels are altered to change the sediment route through the addition of sediment sinks and bedforms. Often, both the path and slope of the channel may be altered in an attempt to change the rates of sediment and water transport. In the case of the Florida Everglades, a meandering channel was straightened. The route that the sediment travelled went from a winding path to a straight line. Subsequent effects to the ecosystem were disastrous, and current work is attempting to re-create the original route.

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SEE ALSO: sediment budget

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SEDIMENT WAVE

A sediment wave is a transient zone of sediment accumulation in a river channel that is created by sediment input and does not originate solely from variations in channel topography. Similar terms are 'sediment slug' or 'pulse'. Sediment waves have a minimum spatial scale measured in channel widths and a minimum volumetric scale corresponding to major bars (see BAR, RIVER); they exist over a number of hydrographic events (Nicholas *et al*. 1995; Lisle *et al*. 2001). A sediment wave is not necessarily a body of bed material moving *en masse* downstream, but evolves as a disturbance in the interactions between flow, channel topography and the transport of bed material that can be contributed from any part of the basin, including the input immediately responsible for forming the wave.

Sediment waves evolve by various degrees of dispersion and translation, depending on the form and sedimentology of the channel. Dispersion is the spreading of the wave as its apex lowers and remains stationary and its trailing edge remains stationary or is accreted upstream. Translation is the downstream advancement of the wave, including its apex and trailing edge. A bed material wave that evolves only by dispersion is still regarded as a wave. Dispersion dominates the evolution of bed material waves in quasi-uniform, gravel-bed channels (see GRAVEL-BED RIVER), where the Froude number at significant sediment-transporting flows is generally high (<1); translation of waves can be important in sand-bed channels where the Froude number is low ($\ll 1$) (Lisle *et al*. 2001). Well-documented examples of these contrasting behaviours are provided by Sutherland *et al*. (2003) and Meade (1985). Wave material that is finer than ambient bed material can promote wave translation, but sand waves in steep, gravel-bed channels tend to evolve primarily by dispersion (Lisle *et al*. 2001).

The foregoing applies to quasi-uniform channels, but other features and processes in natural channels may promote wave translation. DEBRIS FLOWS, for example, translate sediment downstream during single events. The downstream spread of a sediment wave through a series of sedimentation zones (Church 1983), where deposition is locally enhanced by valley-scale topography, could manifest wave translation. Advancing zones of increased transport at the leading edge of a wave may activate sediment stored in unstable reaches and propagate a more pronounced wave downstream (Wathen and Hoey 1998).

The relative dominance of dispersion and translation is important for river resources. Dispersion of sediment inputs attenuates but prolongs sediment impacts downstream, whereas translation propagates pronounced sequences of impact and recovery.

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SEE ALSO: sediment routing

THOMAS E. LISLE

SEDIMENTATION

The term sedimentation refers to the settling of solids from suspension in a fluid. In geomorphology the fluid is typically water (fluvial, lacustrine or marine sedimentation) or air (aeolian sedimentation). The fundamental process of sedimentation is described by Stokes' law which describes the settling of spherical particles from a still fluid.

This is expressed as $V = (2gr^2)(d_1 - d_2)/9\eta$ where: V is the particle fall velocity (cm s^{-1}), g is acceleration due to gravity (cm sec^{-2}), r is the radius of the particle (cm), d_1 is the density of the particle (g cm^{-3}), d_2 is the density of the fluid (g cm^{-3}) and η is the viscosity of the fluid (dyne sec cm^{-2}). In natural systems this simple relation is complicated by the non-spherical nature of most sedimentary particles, and by the fact that particles are typically deposited from a moving fluid column. In a moving fluid the vertical component of turbulent eddies transfers momentum to the particle that may exceed the velocity of gravitational settling so that the particle remains suspended. As flow velocities decline successively finer particles settle out of the column so that theoretically the typical sedimentary structure associated with sediments deposited from a waning flow is a normally graded bed which fines upwards. In practice, the nature of the source sediment, temporal and spatial variability in flow, and post depositional modification, significantly complicate the nature and interpretation of deposits. For a good review of controls on sedimentation and the interpretation of deposits see Leeder (1992).

As a consequence of the difficulties of developing physical models of sediment transport and deposition, geomorphologists have developed empirically based generalizations to describe sedimentation in particular systems. Examples include the Hjulström curve (Hjulström 1935) which predicts entrainment and deposition threshold velocities for sediments of varying size in fluvial systems. More work has focused on defining conditions for the initial entrainment of sediment than subsequent sedimentation. (See for example Shields (1936) diagram or work by Bagnold (1941) on aeolian entrainment.) Threshold velocities for deposition are lower than those for entrainment due to the role of inertia and bed packing in limiting initial motion. For example bedload sediments in Turkey Brook, England show depositional thresholds that are only 35 per cent of entrainment thresholds (Reid and Frostick 1994).

Geomorphological studies of sedimentation can be divided into those which aim to analyse and model contemporary sedimentary processes, and those which draw on this understanding to interpret past environments and processes from sedimentary evidence. The former, in the tradition of the work by Shields and Bagnold, constitute a central part of modern process geomorphology. The insights of the latter approach, often classified as Quaternary science or historical geomorphology, are equally necessary for the proper understanding of contemporary landscapes.

Within geomorphology a broader definition of sedimentation is generally accepted which includes the deposition of sediments from beneath glaciers and ice sheets (glacial sedimentation) and deposition by mass movement processes on slopes. Sedimentation in glacial environments is complex, ranging from purely glacial deposition of sediment, where there is a close link between style of deposition and the dynamics of the glacial ice, through waterlain tills deposited below floating ice, to extensive fluvial and aeolian sedimentation of glacially derived material. A good review of glacial sedimentation is given by Hambrey (1994).

In the broader sense referred to above, sedimentation is taken to be synonymous with deposition of sediment and is consequently central to the study of a wide range of depositional landforms and environments. Sedimentation is the process which defines the end point of the sequence of sediment EROSION, sediment transport and sediment deposition. As such, accumulation of sediment in sedimentation zones provides an integration of upstream/upglacier/upwind geomorphic action. Where system stability allows that the locus of sedimentation remains constant over time the resulting depositional sequence provides a valuable stratigraphic archive of changing rates of sediment flux. These records are of particular importance in determining process rates over geomorphologically significant timescales, extending beyond the few years of most monitored datasets. Lake sediments are a good example of a stable locus of sedimentation and numerous studies of lake sedimentation have inferred long-term catchment sediment yields from lake sediment volumes and chronological control on stratigraphy (e.g. Desloges and Gilbert 1998). In appropriate contexts lake sediments can be used not only to measure sediment flux, but also to identify changes in style of sedimentation which indicate change in the balance of sediment delivery processes. For example, Menounos (2000) used distinct coarse layers in the sediments of a high alpine lake in Colorado to reconstruct the frequency of debris flow activity in the catchment.

Church and Jones (1982) used the term sedimentation zones to describe river reaches where sediment characteristically accumulates, which are separated by reaches where sediment transport dominates. Sedimentation zones are large-scale stores of sediment within the landscape system. The concept can be to some extent generalized. Sedimentation is not uniform across landscape systems but is typically concentrated in zones where the landscape morphology is such that sedimentation is favoured. In the strict definition of sedimentation these areas are those where lower fluid velocities or physical retention of sediment promote sedimentation, for example lakes, or areas of preferential aeolian sedimentation around major topographic obstacles. A good example of the latter are the sand dune systems of Great Sand Dune National Monument in Colorado where sands deflated from alluvial surfaces accumulate against a downwind mountain front (Plate 120). Taking the broader definition of sedimentation, breaks of slope, which promote the accumulation of slope deposits, may be identified as areas of preferential sedimentation. In alpine environments for example cirque basins can be identified as morphological units characterized by local sedimentation.

One common usage in the literature is accelerated sedimentation referring to increased rates of sedimentation usually as a consequence of human action. Accelerated sedimentation is the necessary consequence of enhanced erosion upstream due to human modification of land use, or of increased deposition associated with human modification of the fluid flow. The most common impacts include mining and intensification of upland agriculture. The classic study of mining impacts was carried out by Gilbert (1917) on the impacts of hydraulic gold mining in the Sierra Nevada. Between 1853 and 1884 over 1 billion m^3 of sediment was produced by mining leading to major fluvial aggradation of upland valleys. This sediment continued to be reworked through the Sacramento valley throughout the twentieth century (James 1999). Xu (1998) records a 25 fold increase in sedimentation rates in the Lower Yellow River in China spanning the past 2,200 years associated with a 40 fold increase in population upstream and consequent intensification of agriculture and engineering works on the river. Similarly, O'Hara *et al.* (1993) suggested that significant increases in lake sedimentation rates in a Mexican highland lake were due to over-intensification of pre-Hispanic agricultural practices.

Plate 120 Alluvial sediments in the foreground provide the sediment source for aeolian sedimentation of the dune systems against the mountain front, Great Sand Dune National Monument, Colorado

Phases of accelerated sedimentation are not confined to fluvial systems. Severe desertification in north–central China is occurring as a result of human modification of land use leading to accelerated aeolian sedimentation due to encroachment of dune systems (Fullen and Mitchell 1994).

In general the most dramatic increases in sedimentation rates across a range of different environments, are typically seen within the past 200 years, associated with mechanization and increased rate of land use change in response to rapid population growth.

A related term often used in this context, particularly with reference to infill of river channels and reservoirs, is siltation. The strict definition of siltation is the sedimentation of silt-sized particles but the term is also used more generally to refer to the infill of channels and basins with fine-grained sediment. In areas of rapid erosion reservoir siltation can significantly reduce reservoir capacity and represents a significant economic cost (Palmieri *et al.* 2001) necessitating erosion control measures in the catchment. Fine-grained sedimentation causes particular problems in upper reaches of salmon streams (Hartman *et al.* 1996) as the sedimentation clogs the gravel interstices and inhibits spawning of the returning fish. In areas of the Pacific North West of North America where commercial logging coexists with economically important salmon runs the fine-grained sedimentation associated with logging-related slope failures is a major source of land use conflict.

Long-term changes in sedimentation rate also occur naturally without direct human impact. One of the main drivers is climate change, either directly through impact on weathering and erosion rates, or indirectly through climate driven vegetation change (e.g. Evans 1997; Xu 1998). One major shift in sedimentation rates characteristic of formerly glaciated areas is a peak in sedimentation associated with deglaciation. PARAGLACIAL sedimentation, conditioned by the former presence of glacial ice, commonly occurs at rates of at least an order of magnitude above subsequent equilibrium rates of sedimentation (Hinderer 2001).

Increases in sedimentation rate at a point have two fundamental causes. One is changes in the nature of erosion so that the sediment load is increased and rates of sedimentation increase without necessary changes in the style or location of sedimentation. The second main cause is associated with anthropogenic changes to the nature of the landscape system which change the balance of sedimentation and sediment transport for a particular location. An understanding of sedimentation and sedimentary processes are important to the geomorphologist in many ways. There are many practical applications in controlling and mitigating anthropogenically induced sedimentation, and the sediments are valuable archives of past rates and styles of sedimentation. Fundamentally, however, sedimentation is necessarily linked to erosion (the source of the sediment) and it is the balance of erosion and sedimentation across space and over long timescales which determines the geomorphology of contemporary landscapes.

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SEE ALSO: alluvial fan; bar, river; dune, aeolian; erosion; glacial deposition; paraglacial

MARTIN G. EVANS

SEISMOTECTONIC GEOMORPHOLOGY

Seismotectonic geomorphology is the study of landforms produced by earthquakes. It combines the results of seismotectonic, geomorphic and palaeoseismological research. Palaeoseismology deals with the age, frequency and size of prehistoric earthquakes (Wallace 1981). The palaeoseismic record includes strong (M >6.5) and very strong (M >7.8) earthquakes, since geological effects of moderate or weak earthquakes are rarely preserved in the near-surface zone. Seismic activity is associated with active faulting. Faults are considered active when they show potential or probability of future displacements in the present tectonic setting, or may have displacement within a future period of concern to humans, i.e. they have ruptured during the Holocene (active faults) or the Quaternary (potentially active faults). Active faults are usually segmented, each segment showing a different history of movement. Large earthquakes of a characteristic size repeatedly rupture the same part or segment of a fault.

Earthquakes producing recognizable surface deformation are called morphogenic earthquakes, whereas deposits or landforms formed during an earthquake are described as coseismic, as opposed to delayed-response features that follow the seismic event. Geomorphic features occurring both on-fault and off-fault form either primary (resulting from coseismic slip) or secondary (produced by earthquake shaking, like rockfalls or deformed tree rings) evidence of seismicity (McCalpin 1996). Sediments deformed during seismic shaking are called seismites.

Evidence of present and past earthquakes include deformation of the ground surface along seismogenic faults (fault scarps, fissures, sag ponds, offset stream valleys, shutter ridges, folded terraces, deformed alluvial fans, river reversals, fractured cave structures, displaced beach ridges, coral platfroms, delta plains or wave-cut notches), large-scale features of sudden uplift or subsidence above plate-boundary faults (warped river terraces, elevated shorelines, drowned tidal marshes, emerged or subsided coral reefs), as well as stratigraphic or geomorphic effects of ground shaking or tsunamis far from the seismogenic fault (i.e. landslides, slumps, rockfalls, liquefaction features like mud volcanoes or sand-blow deposits).

The primary geomorphic evidence of seismic activity associated with *normal faulting* are fault scarps (see FAULT AND FAULT SCARP). These scarps range in size from mountain fronts up to 1 km high, cut on bedrock, to centimetre-scale scarplets in unconsolidated sediments. Simple (single-event) fault scarps are formed almost instantaneously, and attain heights from a decimetre to a few metres per event. In normal or reverse faulting such scarps face in the direction of slip, whereas during strike-slip faulting they face in different directions. At the base of recent fault scarps, closed basins or sag ponds may develop. Some scarplets, called earthquake rents, reverse scarplets or cicatrices, parallel the base of the scarp but face uphill. Horsts and grabens, as well as unpaired normal faults creating half-grabens (fault-angle depressions) are also very common. Coseismic displacement on normal faults is characterized by greater subsidence of the hanging wall as compared to the size of uplift of the footwall. Scarp degradation is affected by both lithologic and (micro)climatic factors. In semi-arid climate, the free face in unconsolidated sediments becomes completely destroyed in a timespan ranging from one day (1983 Borah Peak earthquake, $M_s = 7.3$) to one to two thousand years (Crone *et al.* 1987; Wallace 1977). Fractured bedrock scarps will gradually degrade to an angle of repose that can be maintained for one million years or so. Scarps formed by more than one earthquake are called compound, composite or multiple-event scarps. During each earthquake, a colluvial wedge is shed from the scarp, being subsequently covered in interseismic periods by soils that can be dated by different techniques. The fault-induced incision into the upthrown block after faulting can create tectonic terraces that diverge downstream and abruptly terminate against the fault scarp.

Thrust earthquakes occur frequently in fold-and-thrust belts (e.g. Algeria, 1980), at convergent continental plate boundaries (Iran, 1978; Armenia, 1988) or in regions near transpressive bends of strike-slip faults. During thrusting, hanging wall uplift usually exceeds footwall subsidence, and the area affected by coseismic deformation depends on the size of displacement, fault geometry and the rigidity of the deformed crust. The size of displacement diminishes towards the thrust tip, hence, the hanging wall is usually folded adjacent to that tip. Many $M < 7$ earthquakes may not be accompanied by any geomorphic expression. Typical landforms produced by thrusting are thrust fault scarps which are usually more sinuous and irregular as compared to other fault types, and are composed of short, disconnected segments, or produce a continuous but zigzag-like trace on the scale of metres. These scarps show varied morphology due to mixed low-angle faulting and folding. Seven to eight types of thrust fault scarps have been distinguished. Steeply dipping reverse faults in bedrock form simple thrust scarps, whereas those in unconsolidated sediments result in hanging wall collapse scarps. Low-angle thrust faults produce pressure ridges of shape depending on surface material rheology and the magnitude of slip. In more cohesive materials, pressure ridges have smoother fronts and may display backthrusts or represent low-angle pressure ridges, but as thrust displacement decreases below 1 m, all pressure ridge types merge into a single type of a small moletrack. An increasing oblique component of slip produces *en echelon* pressure ridges or oblique tension gashes in pressure ridge front. Surface thrusting deforming flat terrains is usually expressed as a wide gentle warp of fluvial/marine terrace surfaces. Thrust faulting in bedrock produces an overhanging scarp, but in unconsolidated deposits such overhangs collapse very quickly, creating a free face and a steep debris slope. Numerous degraded reverse fault scarps are asymmetric, with the steepest part of the scarp lying downslope of the scarp midpoint, whereas the normal fault scarps typically show symmetric cross profiles. High escarpments formed by repeated thrusting are usually obscured by a long chain of landslides. Thrust faults are difficult to recognize where cutting high relief and irregular topography, whereas broadly distributed thrusting, surface warping and folding cannot be detected unless planar geomorphic features are deformed.

Active folding is a coseismic process related to faulting on a blind thrust or other dip-slip fault at depth. It can also induce seismicity due to flexural slip during folding. Geomorphic manifestations of surface folds are deformed fluvial channels and terraces. Hanging wall ramp folds can generate distinct facets at the top of scarps, resembling those of normal fault scarps. These facets, however, may be independent of the palaeoseismic history of the fault. Some propagation folds formed at thrust tips may also produce multifaceted scarp profiles.

Major seismogenic *strike-slip faults* are associated either with plate boundaries or are located within intraplate settings, at the boundaries of continental microplates. Landforms made by (palaeo)earthquakes along active strike-slip faults include: linear valleys (they can be created by simple deflection of streams along the fault trace, even without brecciation), offset, beheaded or deflected valleys and streams, offset ridges, sags and sag ponds (related to downwarping between two strands of the fault zone), shutter ridges (formed where a fault displaces ridge crests on one side of the fault against gullies on the other side; usually occurring where a fault breaks through the pre-existing pressure ridges), pressure ridges (small warped areas formed by compression between multiple traces in a fault zone), topographic benches (elevated, flat, gently warped or tilted areas, usually formed due to the displacement between several fault segments or splays in the fault zone), fault scarps of minor to moderate height, and small-scale horsts, grabens and pull-apart basins. The fault trace is frequently composed of a wide zone of alternating tension gashes (extensional) and moletracks (compressional) that strike obliquely to the general fault strike. Landforms typically used to estimate palaeoseismic offset are: fluvial terraces, stream channels and alluvial fans. A minimum slip rate, e.g. of the San Andreas fault, California, has been estimated based on offset alluvial fans at 10–35 mm/yr (Keller *et al.* 1982). The maximum single-event stream offset of 9.5 m was produced by the 1857 Ft. Tejon, San Andreas fault, earthquake of M = 8 (McCalpin 1996).

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SEE ALSO: crustal deformation; fault and fault scarp

WITOLD ZUCHIEWICZ

SELF-ORGANIZED CRITICALITY

Self-organized criticality is an approach to understanding non-linear systems initiated by Per Bak and colleagues, as explained in his book *How Nature Works* (Bak 1997). Self-organized criticality is one of a whole host of linked new approaches, including CHAOS THEORY, complexity theory (see COMPLEXITY IN GEOMORPHOLOGY) and FRACTALS, which aim to provide better explanations of the complex behaviour of non-linear natural systems.

Self-organized criticality is used to explain the behaviour of many complex natural systems which seem to evolve into a poised or critical state, far from equilibrium. Per Bak uses the helpful analogy of a self-organized system being like a sandpile created by dropping grains of sand onto a flat surface. During the initial stages of development of the sandpile, predicting the behaviour of the pile is relatively simple and depends upon the physical properties of the individual grains. As the pile grows, however, avalanches start to occur, whose behaviour is complex and cannot be predicted by the characteristics of the individual grains. At this point the system becomes a complex, self-critical one – whose behaviour can only be understood by considering the properties of the whole pile (a holistic approach) rather than from a reductionist description of the behaviour of individual grains. There are, however, many concepts of self-organization used in science which have subtly different meanings and which are based on different interpretations of natural systems (as elucidated for geomorphology by Phillips 1999). Geomorphologists have used a range of these concepts.

In recent years geomorphologists have had a great interest in ideas such as self-organized criticality which may help explain many of the complex landscapes we see around us. Why and how, for example, do regular patterns such as river networks, rills, stone polygons, beach cusps and dune systems develop? Reductionist approaches to these questions have hoped that studying the basic physics of processes operating at the microscale could provide a general answer. However, such approaches have not often proved able to successfully link process and pattern across different scales. Could such patterns instead be seen to be examples of self-organized criticality, where regular patterns have emerged out of the complex behaviour of smaller scale processes? Many geomorphologists have used cellular models to investigate such systems, in which simple rules are applied to describe the interactions of neighbouring cells. As the models are run, patterns at a larger scale emerge from these simple rules.

Several examples illustrate the use of cellular models to investigate self-organized criticality in geomorphic systems. Werner and Fink (1993) investigated the formation of beach cusps, finding that they could be simulated from a cellular model based on the interaction of water flow, sediment transport and morphological change. Similarly, and at a larger scale, Rodriguez-Iturbe and Rinaldo (1997) use models for flow and sediment transport to develop fluvial networks. The resultant networks have fractal and multifractal properties and are seen by Rodriguez-Iturbe and Rinaldo to be the product of self-organizing processes. In a recent study, de Boer (2001) has built a cellular model to simulate the long-term evolution of a fluvial landscape. Using simple rules to model sediment transport between adjacent cells, a record of total sediment yield is created which has complex magnitude and frequency properties which cannot be predicted from the simple, local rules. Thus the sediment dynamics of the modelled landscape are an emergent property of the entire system resulting from the local interaction of individual cells. For geomorphologists, the insight that complex behaviour may be the result of internal interactions rather than external forcings is highly important, and complicates our search for the geomorphic imprint of external forcings such as climate change.

However, despite these promising model simulations it is as yet unclear whether self-organized criticality really exists in natural geomorphic systems. The behaviour of modelled systems cannot simply be applied to natural systems, for which we do not always have enough data and in which external forcings may also play a role. Werner (1999) suggests that hierarchical models might be better suited to modelling complex natural landform patterns which self-organize in temporal hierarchies.

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HEATHER A. VILES

SENSITIVE CLAY

The basic idea of sensitivity is that the structure of the clay soil system has an effect on the properties, so that once the structure is destroyed (by remoulding) a new set of properties is observed. This is probably true in just about all clays although in heavily overconsolidated (see OVERCONSOLIDATED CLAY) systems the effect will be negligible; but in the QUICKCLAYS structure is all-important and when it is destroyed most of the strength properties disappear. The high sensitivities are of interest in applied geomorphology because of associated ground failures and landslides.

The sensitivity of a clay is the ratio of the undisturbed strength to the remoulded strength. It appears to have been first defined by Karl Terzaghi in 1944. The next development is due to Skempton and Northey (1952) and it is this paper which launched the scientific study of sensitivity. There had been Scandinavian experiences with sensitive clays for many years but the Skempton and Northey paper marks a critical beginning. They divided the sensitivity range:

About 1,	insensitive clays
1–2,	low sensitivity
2–4,	medium
4–8,	sensitive
>8,	extra-sensitive
>16,	quickclays

Much of the discussion about sensitivity in clays concerns the mechanism by which it arises. There has been considerable discussion on this topic and many factors influencing sensitivity have been proposed (see Mitchell and Houston 1969); the factors have been nicely ordered by Quigley (1979).

Factors affecting sensitivity

Factors producing high undisturbed strength and high sensitivity:

1 Depositional flocculation, including low electro-kinetic potential, high sediment concentration, divalent cation adsorption.
2 Slow increase in sediment load.
3 Cementation bonds, including carbonates and amorphous sesquioxides

Factors producing low remoulded strength and high sensitivity:

1 High water content (greater than the liquid limit), little consolidation.
2 Low specific surface of soil grains, high silt content or high rock flour content in the clay fraction. High primary mineral, low clay mineral content.
3 High electro-kinetic (zeta) potential, low salinity via leaching, organic dispersants, inorganic dispersants, high monovalent cation adsorption relative to divalent cations.
4 Low amorphous content.
5 Low smectite (EXPANSIVE SOIL clay) content.

The list is fairly comprehensive, but it includes major and minor factors. It appears that there are probably two basic populations of sensitive clays: those with relatively low sensitivities, which contain clay minerals to a significant degree; and those with high sensitivities which tend to lack clay minerals. These high sensitivity materials are dominated by primary minerals with particle sizes in the clay or very fine silt ranges.

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SEE ALSO: quickclay

IAN SMALLEY

SERPULID REEF

Serpulid reefs are built by marine polychaetes which secrete hard, calcareous tubes. The reefs built by *Ficopomatus enigmaticus* (Fauvel) develop reefs which are greater than 3 m thick and up to 20 m in diameter (Fornós *et al*. 1997). Individual worm tubes are typically 100 μm thick, 4–5 mm in diameter and up to 100 mm long, with tubes being comprised of calcium carbonate interspersed with a mucopolysaccharide matrix (Rouse and Pleijel 2001). Serpulidae have a near worldwide distribution; reefs have been found in the fossil record from the Cretaceous to the Recent and they are important in palaeoenvironmental reconstruction. The global distribution of the species has increased during historical times, due to international shipping, with several species colonizing non-native habitats and industrial structures such as docks.

The reefs typically develop in brackish and marine conditions, in intertidal and shallow subtidal zones, on hard rock substrata or shell fragments. Occasionally they extend onto, and in rare circumstances they colonize, soft substrates such as mud, sand or wood. Serpulid reefs tend to develop in three phases: (1) individuals coat the surface of hard substrata; (2) sinuous, random growth and extension of colonies; and (3) development of reef structures which are primarily controlled by water turbulence and dominant current direction. Morphological forms vary from fringing reefs along rocky shorelines, to subtidal microatolls and dense patch reefs (Fornós *et al*. 1997). Where serpulid reefs develop on soft substrata they often serve an important functional role, by providing the only hard substrate for a range of species.

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LARISSA NAYLOR

SHEAR AND SHEAR SURFACE

Shearing of material occurs when it is subjected to a stress or confining pressure, acting in a particular direction, that exceeds the strength of the material. The material fractures along a plane of least resistance, which may be curved, and the mass on one side of this shear surface is displaced so that movement takes place in opposite directions on either side.

In geomorphology, shear failure of materials is most commonly encountered in MASS MOVEMENTS. In a hillslope, material near the ground surface (soil or other sediments, *in situ* weathering products, bedrock, etc.) is subjected to shear stress as it is acted on by a downslope component of gravitational force. This is resisted by the shear strength of the material, which acts in the opposite direction. If the stress exceeds the strength, a shear surface will form (within one material or between different materials depending on relative strengths across all possible planes of weakness) and the mass above the surface will move downslope. (See HILLSLOPE, PROCESS; LANDSLIDE; SLOPE STABILITY.) Fault planes within the Earth's crust can also be regarded as shear surfaces: the rocks on one side of a fault move past the rocks on the other side in response to differential stresses.

Another condition that can cause shearing is when a material is confined at depth by stresses acting in perpendicular (x, y, z) directions, and compression by the stress in one direction exceeds the strength of the material. A shear surface forms as the material is forced to the side in opposite directions either side of a plane of least resistance.

The nature of a shear surface and any displacement along it will be determined by the properties of the material(s), the magnitudes and directions of the stresses, the topographic or geophysical context, and the scale at which the shear surface is considered. A smooth plane of movement at a large scale may comprise an irregular shear surface at a smaller scale. There may be considerable frictional resistance to movement, especially if the irregularities resemble interlocking teeth. At this extreme, the shear stress will only cause movement if it is sufficient either to force the two sides far enough apart to enable movement, or to cause shearing of the asperities thus further smoothing the overall surface.

At the other extreme, very small stresses within fine-grained materials may cause individual plate-like clay mineral particles to line up with the prevailing stress direction. Over time these 'microshears' can extend, join up and form a shear surface, which is smooth at the microscopic scale (see SLICKENSIDE).

Some materials will not form a discrete shear surface, instead developing shear zones where complex displacements occur throughout measurable thicknesses, especially in materials that display plastic behaviour under stress such as certain clay formations, or highly heterogeneous sediments.

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SEE ALSO: riedel shear

ALAN P. DYKES

SHEET EROSION, SHEET FLOW, SHEET WASH

In a general sense, the term sheet flow is used to refer to any flow of water of more or less homogeneous depth and velocity over a surface without any clear channel development. The term is used to describe flow over alluvial floodplains, coastal plains, beach surfaces as well as flow over hillslopes without noticeable channel incision. Here, we will only refer to the use of the term sheet flow in this latter sense. Sheet flow over hillslopes therefore contrasts with flow in clearly defined channels (RILLS and gullies (see GULLY)).

Sheet erosion is often defined as the removal of a thin layer of surface soil by water erosion processes without the development of noticeable channels (Plate 121). The absence of channels (see CHANNEL, ALLUVIAL) does not imply that sheet erosion is completely uniform. The presence of roughness elements such as rock fragments, clods or vegetation tussocks will still lead to variations in flow depth and velocity, even under steady flow conditions. In order to stress the absence of channels the term interrill flow (i.e. flow on areas between or without rills) is sometimes preferred. Local variations in flow depth and velocity do not necessarily lead to channel development: as Smith and Bretherton (1972) pointed out, no channels will be formed by OVERLAND FLOW when the influx of sediment into a concentrated flowline by processes such as splash and creep equals or exceeds the local transporting capacity of the flow.

Sheet erosion mainly occurs under conditions where the soil surface is insufficiently protected by vegetation cover against drop impact and crusting (see CRUSTING OF SOIL): it is therefore frequently encountered on freshly ploughed arable land, on overgrazed rangeland and on natural hillslopes in arid and semi-arid environments.

If a soil surface is only affected by sheet erosion this will result in a gradual lowering of the whole soil surface. In many landscapes, sheet erosion often occurs simultaneously with rill and gully erosion, with gully erosion dominating in HILLSLOPE HOLLOWS (concavities), while sheet and rill erosion mainly affect convex and rectilinear slope segments. When hillslope erosion mainly occurs in channels, the much more rapid lowering of the channel beds will in principle result in a dissection of the soil surface. This need not always be the case: on arable land, rills formed due to erosion are frequently obliterated by soil tillage. The repeated occurrence of sheet and rill erosion will in this case also lead to a gradual removal of a 'sheet' of topsoil over the whole field surface. In natural conditions rill beds often stabilize after a certain time period due to bed ARMOURING: rills then become conveyers of sediment eroded on interrill surfaces rather than important sediment

Plate 121 Intense sheet erosion on arable land in central Belgium, caused by a very intense rainstorm. Height of pole: *c*.1.5 m

sources. In the long term, this situation may also result in a gradual removal of the topsoil over the whole surface. The rest of this section concentrates on sheet erosion *sensu stricto*, i.e. water erosion that occurs without the formation of noticeable channels.

In the case of sheet erosion soil detachment mainly occurs by raindrop impact. The detached sediments may then be tranported by various processes: raindrop splash, raindrop-induced flow transport and flow transport. The hydraulic properties of the sheet (or interrill) flow are an important control on these processes and are therefore discussed first.

Sheet flow on hillslopes is generally characterized by flow depths between 0 and 20 mm and velocities $< 0.5\,\mathrm{m\,s^{-1}}$. In general, experimental studies in the field and laboratory show that classic approaches for the prediction of flow depths and velocities which are based on the estimation and prediction of the Darcy–Weisbach friction factor or Manning's n are only moderately succesful (Abrahams *et al*. 1994). The key reason for this is that the interaction between roughness elements and flow hydraulics is fundamentally different in sheet flow. On natural surfaces where individual roughness elements are more or less randomly distributed, flow resistance tends to increase with increasing discharge until roughness elements are fully submerged. A further increase of discharge leads to a rapid decrease of flow resistance (Lawrence 1997). On tilled surfaces roughness elements are not randomly distributed and the correct prediction of flow characteristics sometimes requires separate calculations for each flowpath (Takken and Govers 2000). As the latter requires very detailed information on surface topography, simplified approaches are often used.

Raindrop detachment is basically controlled by the balance between rainfall erosivity and the soil's resistance to splash detachment. Various rainfall characteristics have been proposed as indicators of the rainfall's EROSIVITY. Rainfall kinetic energy is undoubtedly most frequently used to assess rainfall erosivity. However, there are indications that parameters that give relatively more weight to drop diameter, such as the product of rainfall momentum and drop diameter, are somewhat better predictors of splash detachment (Salles *et al*. 2000). Splash detachment is also strongly controlled by the presence and the thickness of a water layer on the surface: most studies report an exponential decline of splash detachment with water depth, with splash detachment becoming negligible when the depth of the water film exceeds *c*. one drop diameter (Torri *et al*. 1987). The resistance of a soil to splash detachment is determined by the cohesion of the soil and the weight of the soil grains. Minimum resistance values are observed for fine sandy soil materials as these materials consist of very fine grains yet they are almost totally cohesionless (Poesen 1985).

If a significant slope gradient is present, the redistribution of the detached material will result in a net downslope transport of soil. The rate of downslope transport increases approximately linearly with slope gradient. However, as splash distances are of the order of several decimetres, splash transport is generally negligible compared to (raindrop-induced) flow transport, except in the case of short, very steep slopes (e.g. unprotected terrace risers).

Although sheet flow has a limited capacity to detach (cohesive) sediment, it is in most cases the main transporting agent in sheet erosion. Two modes of sediment transport can be distinguished: (1) flow transport, whereby the flow itself is transporting the sediment and (2) raindrop-induced sediment transport whereby sediment is brought into the flow by raindrops impacting soil surface and the suspended sediment is consequently transported downslope. Raindrop-induced sediment transport flow is most important for relatively coarse (sand-sized) material and for low-energy flows with relatively low flow depths (Kinell 2001).

Sheet erosion is slope-dependent, primarily because the transporting capacity of the sheet flow increases with slope gradient. The increase of sheet erosion intensity with slope gradient is more or less linear, all other factors being constant. In practice other factors such as grain size distribution of the surface sediments, crusting and vegetation characteristics are strongly slope dependent, so that the relationship between slope and sheet erosion rates may take a completely different form. An interesting debate exists with respect to the effect of slope length on sheet erosion rates: recent experimental evidence suggests that the rapid increase of erosion rates per unit surface area which is often observed when plots of limited length are compared cannot be extrapolated to greater slope lengths. The increase of erosion rates with increasing slope length for short slopes is due to the fact that in the first metres below the divide the sediment load in the sheet flow is supply-limited and therefore sediment transport increases rapidly with slope length. Further downslope, sediment transport becomes limited by the transporting capacity of the flow. This leads to a much slower increase of sediment transport with slope length and consequently also to a much slower increase or even a decrease of the erosion rate per unit area (Rejman and Usowicz 2002).

Considering the primary role of raindrop impact in detaching and transporting sediment in sheet flow, it is no surprise that sheet erosion is strongly dependent on rainfall and rainfall energy: this is, amongst others, reflected in the fact that rainfall erosivity in the Revised Universal Soil Loss Equation which is designed to describe both sheet and rill erosion equals the product of rainfall intensity and rainfall kinetic energy during a 30-minute period.

In general sheet erosion decreases strongly with increasing cover of the soil surface by vegetation or other non-erodible elements such as rock fragments. Most studies, such as the one by Hussein and Laflen (1982), report an exponential decline, whereby a soil cover of *c*. 30 per cent reduces erosion to less than 50 per cent of the value for a bare surface. This strong, non-linear response is due to the fact that the presence of cover has an impact on various aspects of the sheet erosion process (increased infiltration, protection of the surface cover against splash, reduced flow velocities, etc.). The adequate management of soil cover is therefore the most important management strategy in order to reduce sheet erosion.

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GERARD GOVERS

SHEETING

Some rocky massifs are divided by flat-lying or gently arcuate partings that at many sites are more inclined than the land surface, and plunge steeply (up to 70°). These fractures are known as sheeting or EXFOLIATION. Even though the term sheeting suggests thin layers, some are 10 m or more thick. Sheeting has been observed to depths of 100 m or more and is well developed in granitoids but also in other quite different rocks (dacite, rhyolite, sandstone, conglomerate and limestone). There are two interpretations of sheeting that fall into two categories: exogenetic and endogenetic. The exogenetic explanations are INSOLATION WEATHERING, CHEMICAL WEATHERING and offloading or pressure release, all of them imply dilation by rock volume increase.

The endogenetic explanations are all concerned with tectonic stresses. Some authors consider the sheet structure formed during the emplacement and later cooling of the granite mass and propose the same origin for the shape of the associated dome. But sheeting is well developed in sedimentary and volcanic rocks that were never emplaced and even in granites the magnetic foliation contemporaneous with the emplacement of magmatic rock is clearly discordant with the sheeting.

It has also been suggested that sheeting is associated with big thrust-shearing planes which would affect the granite as well as the other rock types where sheeting is habitually observed. This explanation is the best one, because it serves for all kinds of rocks affected by sheeting.

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SEE ALSO: pressure release

JUAN RAMON VIDAL-ROMANI

SHIELD

The continental nuclei where Archean and Proterozoic rocks, typically burittle, rigid, granitic, gneissic and associated rocks, outcrop. They form extensive, flat, relatively stable areas (cratons) which have been relatively undisturbed since Precambrian time, except for gentle warping. The major shields are Canadian, Fennoscandian, Angaran (north-east Siberia), African, Brazilian, West Australian and East Antarctic. The shields are bordered by stable platforms, which are continental areas floored by shield extensions and covered by a relatively thin layer of sedimentary rocks.

Because they are ancient areas, they have often experienced bevelling of their rocks by erosion and are characterized by extensive erosional plains. Some shield areas have been fashioned in part by glacial erosion, as in Canada and on the Baltic Shield, but ancient saprolites may also be extensively developed (Lidmar-Bergström 1995). Because of the presence of limited relief and resistant rocks, shields tend to be areas with low rates of mechanical and chemical denudation (Millot *et al*. 2002).

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A.S. GOUDIE

SHINGLE COAST

The term 'shingle' has been used for at least 400 years in Britain and some Commonwealth countries, to describe sediments composed of mainly rounded pebbles, larger in diameter than sand (>2 mm) but smaller than boulders (<200 mm). Elsewhere terms such as gravel, stone, levées de galets, playas de cantos, schotterwälle and steinstrand are used. A generalized world distribution of shingle coasts is given in Figure 148. In many locations shingle is mixed with sand, silt, clay or organic debris, resulting in a 'mixed' sediment beach (e.g. Kirk 1980), but all shingle and boulder beaches can be regarded as different types of 'coarse clastic' beach (Carter and Orford 1993).

In general shingle coasts have received less scientific attention than sandy and muddy shorelines. In part, this reflects the fact that, at a world scale, they are much less common. However, in recent decades there has been an increasing awareness of the geomorphologic, ecological and engineering significance of shingle coasts in the contexts of sea-level change, flood defence and habitat conservation. Such coasts are now recognized as an internationally important, but disappearing resource (Packham *et al*. 2001).

Shingle coasts form in WAVE-dominated locations where suitably sized material is available. At a global scale they dominate high latitudes and those areas of temperate shores which were affected by Quaternary glaciation. They are locally important in some other temperate and low latitude areas where high relief landscapes of suitable geology occur near the coast, near the estuaries (see ESTUARY) of high-energy rivers, or where coral (see CORL REEF) is present. Elsewhere they are of limited importance. Isla and Bujalesky (1993) describe shingle coast locations in Argentina, McKay and Terich (1992) and Forbes *et al*. (1995) in North America and Shulmeister and Kirk (1993) in New Zealand.

At a regional scale, lithographic composition determines shingle availability and durability. Hard materials such as flint, chert, granite, quartzite and some metamorphic materials survive much longer at this clast size than sandstones, limestones or shells. Around Great Britain some 19,000 km of shoreline have an important shingle component, with almost 3,500 km of these coasts being pure shingle (Sneddon and Randall 1993/1994). Many of the

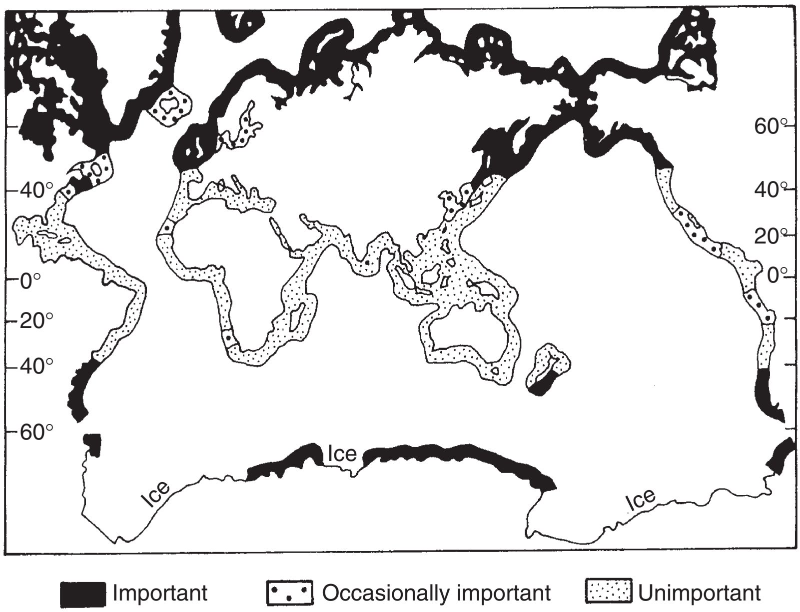


Figure 148 A generalized world distribution of shingle coasts (after Pye 2001)

shingle-barrier (see BARRIER AND BARRIER ISLAND) systems occurring on present-day coastlines were initiated during the Holocene (see HOLOCENE GEOMORPHOLOGY) marine transgression and are currently sustaining considerable morphological change as a result of increasing SEA LEVELS causing landward and longshore reworking of a finite sediment volume.

Shingle coasts can comprise several different landform types (Figure 149), which vary according to their history, mobility and oceanicity and therefore offer different habitats to vegetation, wildlife and humans (Pye 2001; Sneddon and Randall 1993/1994).

Fringing, or pocket BEACHes, are narrow strips of shingle coast in contact with the land along the top of the beach. These are usually subject to regular marine inundation. They frequently occur at the foot of sedimentary cliffs, such as chalk in southern Britain, but may also occur in front of coastal dunes or saltmarsh cliffs.

Embayment beach-ridge plains, or apposition beaches, are comprised of a series of relict storm beach-ridges and an active front ridge system which together partly or totally infill a previous embayment. Such systems may be hundreds of metres or even kilometres wide and can be transitional to CUSPATE FORELANDs or nesses.

Shingle SPITs are strips of shingle, which grow out from the coast where there is an abrupt change in the direction of the coastline. They commonly occur, therefore, along coasts which have an irregular plan. Spits often display recurved hooks along their length and at their distal ends, where the shingle is, or has been, subject to wave action from two or more directions. Indeed, in many cases, it is possible to trace the development of a spit's growth via recurved hooks, seen as lateral projections from the lee of the spit, which locate the position of the past distal points (Randall 1973; Plate 122). Paired spits are found at the entrance of several harbours on the south coast of England, including Pagham and Langstone. These may have originated as bars or TOMBOLOs, which have breached, but in other cases, independent growth of two spits may be due to bi-directional longshore drift.

On eroding coasts, shingle spits are transgressive and frequently overlay back-barrier marsh or lagoon deposits as at Shingle Street, Suffolk, and in some instances may be dissected to form barrier islands. Transgressive ridges, often composed mainly of shell-shingle, are well developed on the marsh coast of Essex. Similar features are also found in the Gulf of Mexico, where they are

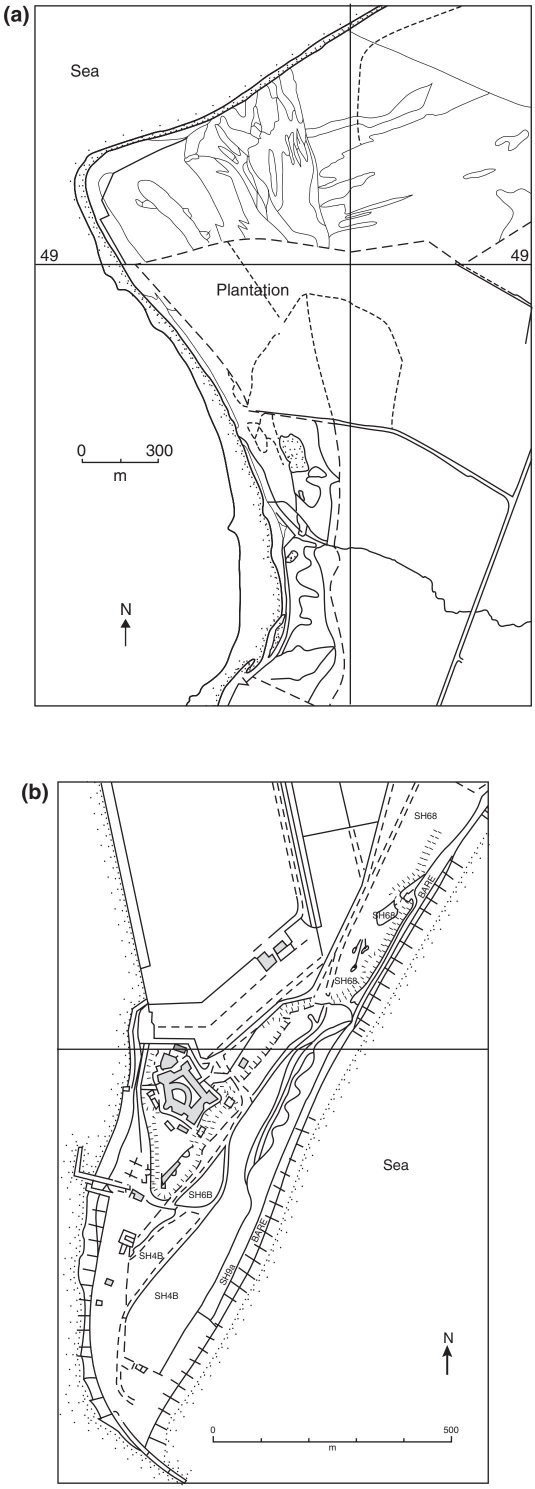
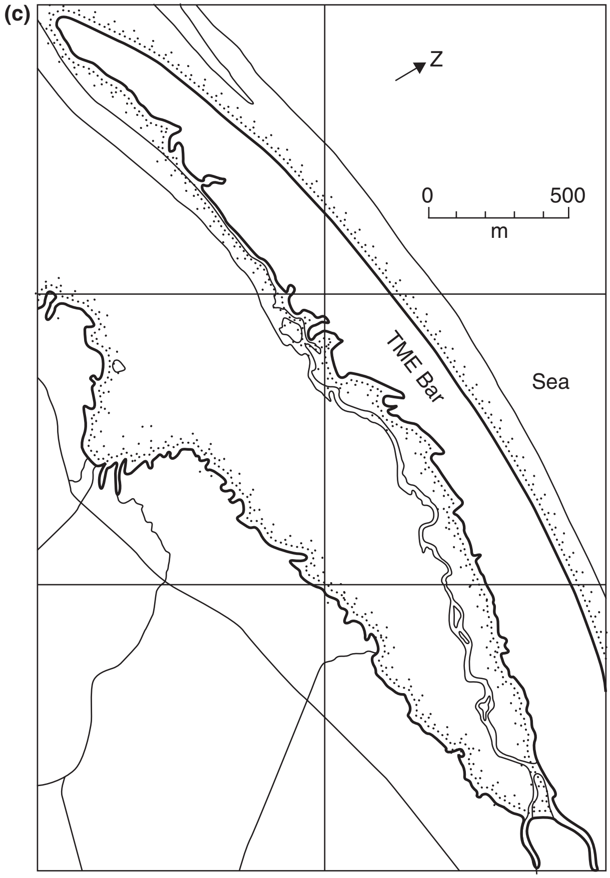


Figure 149 (a) A fringing beach at Llandulas in north Wales; (b) a shingle spit at Landguard Point, England; (c) a shingle bar at Culbin, Scotland; (d) an offshore barrier island, Scolt Head, England



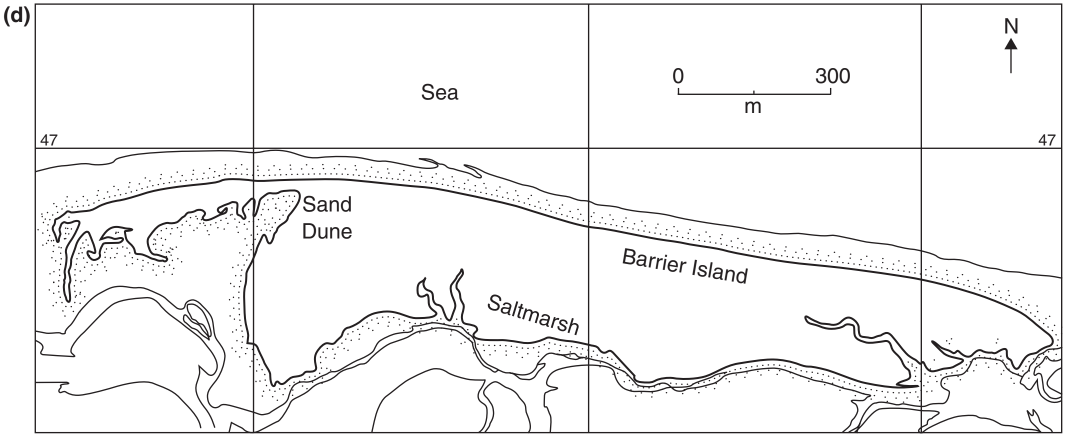


Figure 149 (Continued)

Plate 122 The shingle coast of Suffolk, UK. On the distal point of Orford Ness and on the mainland opposite there are recurved hooks enclosing lagoons. At the mouth of the estuary of the River Ore longshore drift of shingle sediments can be observed (photograph courtesy of Cambridge University Collection of Air Photographs)

known as CHENIER RIDGEs, and in Auckland Bay, New Zealand. Tombolo barriers, or bars (see BAR, COASTAL), are geomorphologically similar to spits, representing the extreme case where a spit has grown across an estuary or coastal indentation. This results in the formation of a lagoon behind the bar, which clearly affects the hydrology and ecology of the leeward slope. Chesil Beach in Dorset and Slapton Ley in Devon are prime British examples.

Rivers, which provide a source of shingle-sized sediment (see GRAVEL-BED RIVERS), may have prograded strandplains or deltas of shingle at their mouth. In Scotland, the Kingston Shingles are found at the mouth of the Spey (Sneddon and Randall 1993/1994) and South Island, New Zealand has particularly good examples, such as at the mouth of the Waitaki River.

At points of littoral drift convergence the formation of a second set of apposition ridges deposited at a different angle, will lead to the formation of a ness or cuspate foreland, a triangular mass of shingle such as Dungeness, Kent, in England, Rhunahaorine Point, Argyll in Scotland or Cape Canaveral in Florida. The Island of Rügen in Baltic Germany is effectively a cuspate foreland cut off from the mainland. Such features often support a terrestrial geomorphic system inland of the coastal ridges.

The final type of shingle formation is the offshore barrier island, formed where a large mass of shingle has been deposited offshore and which may act as the 'skeleton' for a coastal sand-dune (see DUNE, COASTAL) system. Culbin Bar, Morayshire and Scolt Head Island, Norfolk, are prime British examples.

Most shingle coasts have a steep upper beach slope and a relatively steep overall nearshore profile. Partial wave energy reflection results in the formation of edge-waves and rhythmic longshore features such as BEACH CUSPs. Shingle features are frequently of considerable ecological importance in terms of habitat diversity and play a vital geomorphologic role in determining the stability of adjoining 'soft' sediments of mudflats and saltmarshes. Unless the shingle coastal features are mobile, a partial vegetation cover is the norm. The middle and lower beach are usually kept bare by wave action, but upper beaches are vegetated. The rate and extent of plant colonization is dependent upon the degree of disturbance and shingle mobility, the presence or otherwise of a fine sediment matrix within the spaces between larger sediments and the hydrological regime of the shingle.

All shingle coasts contain a mixture of different sized sediments. Some are well sorted and consist entirely of pebbles, while others are poorly sorted and may also contain sand and/or boulders. Because there is frequently considerable temporal and spatial variation in shingle and mixed shingle/sand beaches (Kirk 1980; Schulmeister and Kirk 1993), accurate determination of average textural qualities is difficult.

Most coarse sediment coasts become coarser up-beach, because backwash and gravity can move larger clasts. Hence many locations have shingle only on the upper beach. Williams and Caldwell (1988) also comment on clast shape with discoid pebbles sorted preferentially on the upper parts of the beach with spheres and rods occurring nearer the sea. Sediment grading alongshore also occurs due to selective transport of finer sediments in the downdrift direction as at Chesil Beach. However, other sites show much more complex patterns as a result of bi-directional currents of varying magnitudes.

Shingle coast micro-relief dynamics depend upon spring to neap tidal patterns and wind and wave conditions. The upper 50–80 cm of sediment is frequently remobilized forming berms and cusps that change from one tidal cycle to another. More major changes occur seasonally as a result of spring to neap tidal fluctuations and especially at those times when storm-wave energy is higher.

The internal sedimentary architecture of shingle landforms reflects the process regime and net evolutionary trends of the structure (Randall 1973). Ridge external structures vary dependent upon whether they are vertically accreting but laterally stable, laterally migrating or developing on a seaward prograding plain. The depressions between ridge crests may be partly filled by washover and storm-tossed deposits, so that there is often a marked difference in average particle size and shape between ridge fulls (crests) and lows (Randall and Fuller 2001). Sediment grading may also occur as a result of longshore drift with selective transport of finer sediments downdrift. However, on many coasts sediment grading has been found to be complex in relation to seasonal variation in the longshore current regime (Pye 2001).

Sea-level rise has the tendency to move shingle landforms inland (Carter and Orford 1993; Forbes *et al.* 1995), but if sea-level rise is particularly rapid, shingle structures may be drowned *in situ* by overstepping. Normally, however, under moderate storm-wave activity, shingle is pushed to the top of the front-beach ridge, while in major storms the ridge is overtopped or breached, creating shingle aprons in the backbarrier area. As this pattern is repeated, so the ridge migrates landwards by rollover. Many of the major shingle formations present today formed in this way during the Holocene marine transgression, initiating at a time of lower sea-stand and reaching their present location by around 4,000 BP. Most current shingle features are relict or dependent upon erosional sediments rather than glacial debris. Hence, there is currently a shortage of sediment at the updrift end of many transport cells and increasing risk of OVERWASHING and breaching (Orford *et al.* 2001).

Traditionally in developed countries, shingle coasts have been heavily managed to retard erosion, drift and sediment cycling and, more recently, for habitat conservation. Management methods may include beach reprofiling or protection (with gabions or tetrapods) or the construction of GROYNEs and offshore breakwaters. Groynes have been used since the nineteenth century but frequently they have a negative effect downdrift by reducing longshore sediment availability. More recently beach nourishment has been seen as more cost effective and environmentally acceptable (Bradbury and Kidd 1998), but this, too, may change the natural geomorphologic character of the coast and may not be cost effective in the long term. Elsewhere gravel extraction, building developments, military activity and MANAGED RETREAT have markedly changed the landscape and landforms of shingle coasts.

At a world scale, large shingle structures are uncommon and under increasing pressure from development, mining and 'coastal squeeze' as a result of rising sea levels and coastal erosion. Most shingle structures were formed earlier in the Holocene Period and currently shingle supply is limited at the updrift end of coastal SEDIMENT CELLs. This results in the increased likelihood of breaching during STORM SURGEs. Naturally, shingle coasts are dynamic and tend to migrate landward but people prefer a static coast. For economic reasons some areas of shingle coasts have to be protected, but in less developed areas space should be left for natural dynamic coastal change. Wherever possible shingle structures should be left entirely alone, since in the majority of circumstances, geomorphologic change promotes environmental diversity (Randall and Doody 1995).

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SEE ALSO: coastal geomorphology; raised beach

ROLAND E. RANDALL

SHORE PLATFORM

Shore platforms are rock surfaces created by the erosion and retreat of coastal cliffs (Figure 150) (see CLIFF, COASTAL). Although geological and other factors are responsible for enormous variations in their morphology within fairly small areas, the distinction has often been made between subhorizontal, supra-, inter-, or subtidal platforms that terminate abruptly seawards in a low tide cliff, and gently sloping, largely intertidal, platforms, with gradients between about 1° and 5°, that continue below the low tidal level without a major break in slope (Plate 123). Subhorizontal platforms have generally been associated with Australasia, although they are common in much of the tropical and subtropical world, whereas sloping platforms have been described most frequently from the stormy waters of the northern Atlantic.

The inherent complexity of shore platforms and other rocky coastal systems has made it difficult to determine how they are formed or how they develop through time. The physical resistance of

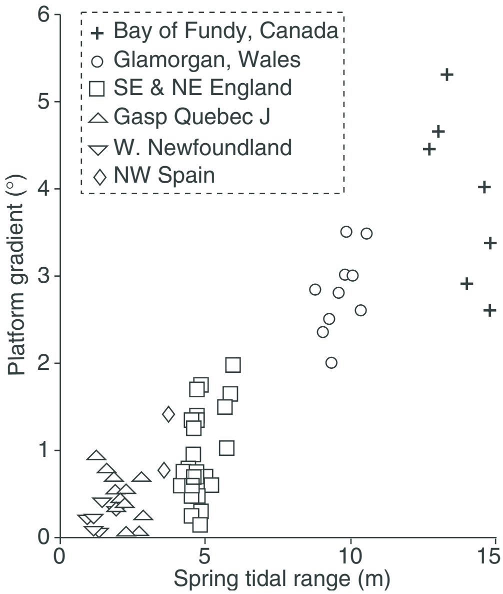

Figure 150 Shore platform in Liassic limestones and shales at Monknash, south Wales

Plate 123 Gently sloping and quasi-horizontal shore platforms

rocks depends upon their chemical composition, angle of dip, strike, bed thickness, joint (see JOINTING) pattern and density, degree of WEATHERING and a myriad of other factors. A wide range of mechanisms also operate on shore platforms, including WAVEs, tides, bio-erosional and bio-constructional organisms, frost, chemical and salt weathering, WETTING AND DRYING WEATHERING and MASS MOVEMENTs. The relative and absolute importance of these processes have varied through time, with changes in relative SEA LEVEL and climate, and rock coasts often retain vestiges of environmental conditions that were quite different from today.

Much of the debate over the past one hundred years has been concerned with the relative roles of marine and subaerial processes in platform development, and, more recently, on the relationship between platform morphology and tidal range. A number of mechanisms have been proposed for platform formation:

- Platforms are cut in weathered or unweathered rock by waves. This produces sloping platforms (wave-cut platforms) in macrotidal regions and horizontal platforms, at the 'level of greatest wear' in areas with a small tidal range.
- Old Hat platforms develop in very sheltered areas, at the level of permanent seawater saturation. Above this level, weak waves wash away the fine, weathered debris, exposing the top of the resistant, unaltered rock below.
- Platforms can be formed by differential wave erosion of cliffs consisting of weak, weathered rocks above the saturation level, and more resistant, unweathered rock below. Evidence is lacking, however, for the existence of a permanent intertidal level of saturation in coastal rocks.
- Horizontal platforms, often with ridges or ramparts at their seaward ends, develop by WATER-LAYER WEATHERING and other weathering processes levelling and lowering uneven, rough, sloping or subhorizontal platforms that were originally cut by waves.
- Horizontal and sloping platforms are the result of alternate wetting and drying, which is responsible for cliff erosion and platform downwearing.
- Some workers have proposed that frost and possibly sea ice produces shore platforms in cool climatic regions.

It is increasingly evident that shore platforms are the product of mechanical wave erosion, weathering and bioerosional activity, although their relative importance depends upon climate, geology, wave and tidal conditions, and stage of development. Air compression in rock crevices and other mechanical wave erosional processes are most effective on steep uneven platform surfaces. As platforms become wider, smoother and more gently sloping, mechanical wave erosion must become less effective because of wave attenuation and the lack of rock scarps or upstanding beds of steeply dipping rock, and weathering processes must therefore become, at least relatively, more important. MICRO-EROSION METER (MEM) data from a variety of environments suggest that shore platforms are being lowered by weathering at rates that are frequently between about 0.5 to 1 mm yr^{-1}. It is difficult to accept that these high rates can be sustained indefinitely, however, and there is some MEM data to support the contention that they must decrease through time as platforms are reduced in elevation and therefore experience progressively longer periods of inundation, shorter periods of exposure and less frequent cycles of wetting and drying.

There is a moderately strong global relationship between mean regional platform gradient and tidal range (Figure 151). For wave-cut shore platforms this can be attributed to the way that tides control the expenditure of wave energy within the intertidal zone. The strong correlation between

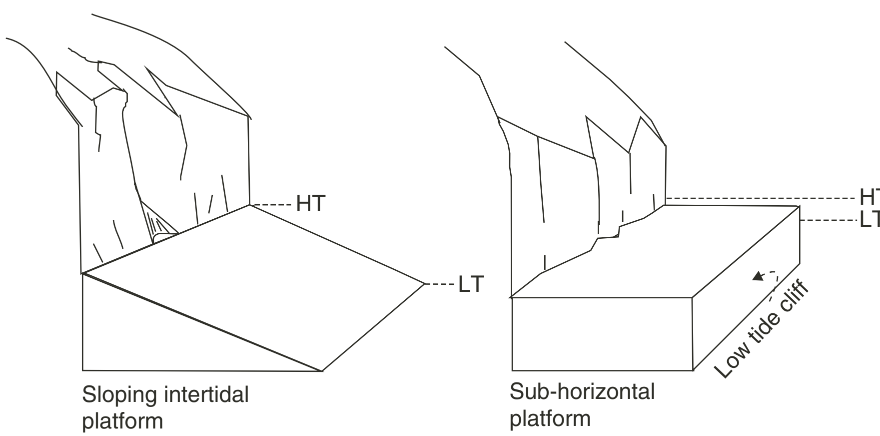

Figure 151 Relationship between mean regional shore platform gradient and spring tidal range

wetting and drying frequency distributions and tidal range also provides a possible explanation for the gradient–tidal range relationship in areas where weathering rather than wave action is dominant. Nevertheless, there is a basic problem with all theories that attribute platform formation entirely to subaerial or intertidal weathering, while relegating the role of waves to removal of fine-grained debris. This is concerned with the apparent lack of a mechanism to place a limit on maximum platform width, if wave strength and attenuation are not important factors.

Rock coast workers need to determine whether, or to what degree, shore platforms and related elements of rock coasts have been inherited from interglacial stages (see ICE AGES) when sea level was similar to today's. In hard, resistant rocks, the platforms often seem to be too wide to have developed in the few thousand years since the sea reached its present level, and they are frequently backed by ancient composite cliffs with glacial, periglacial or other terrestrial deposits, RAISED BEACHes and erosional ledges at their base. Although these coasts often lack datable materials, a variety of techniques has been used to show that at least in some areas, shore platforms, caves and other features have been inherited from one or more interglacial stage. It is particularly difficult to assess the possible role of INHERITANCE in areas of fairly weak rock, because platform width is less anomalous with regard to present rates of erosion than in more resistant rock areas, and because faster rates of erosion could account for the general lack of till covers, ancient beach deposits and structural remnants. Lacking evidence to the contrary, most workers have concluded that shore platforms in fairly weak rocks are entirely postglacial features. There is abundant evidence, however, of coastal inheritance from the last and older interglacial stages on the fairly weak rock coast of Galicia in northwestern Spain. Thick fluvio-nival and geliflucted slope deposits covered this coast during the latter part of the last glacial stage, and the ancient coast was then exhumed and inherited as sea level rose to its present position during the Holocene. The Galician evidence has important implications for the possible role of inheritance in the development of shore platforms in other areas (Trenhaile *et al*. 1999).

Modelling provides one of the few ways to study the long-term evolution of slowly changing rock coasts. The earliest MODELS were qualitative and structured within evolutionary cycles of erosion. More recent models are mathematical, but although field evidence suggests that most mechanical wave erosion occurs through water hammer, air compression in joints and ABRASION in shallow water, processes that are closely associated with the water surface, models have generally been concerned with submarine erosion in tideless seas. Nevertheless, a model has been developed that considers rates of wave attenuation and the long-term distribution of wave energy within the intertidal zone. This model has been used to study the long-term evolution of shore platforms with Quaternary changes in sea level on stable and tectonically mobile coasts. The model indicated that whether an ancient shore platform is subsequently inherited, modified or completely replaced by a contemporary platform depends upon the complex interaction of a multitude of factors that determine the erosive efficacy of the waves. It suggested that intertidal and subtidal surfaces trend towards a state of static equilibrium under oscillating sea level conditions, as attenuated waves become increasingly unable to continue eroding the rock, although they can be in a temporary, though possibly long-lasting, state of DYNAMIC EQUILIBRIUM. Most modelled surfaces were, at least in part, inherited from one, or in many cases more, interglacial stages when sea level was similar to today's (Trenhaile 2001). In future, platform models must also consider the effects of downwearing by weathering as well as backwearing by waves within the intertidal zone. In turn, reliable modelling is dependent on the acquisition of quantitative field data. Although the number of sites and the length of the records are quite limited, the micro-erosion meter has provided useful information on platform downwearing by weathering and corrasion. We are still unable, however, to measure the effect of joint block and large rock fragment detachment by wave quarrying and other mechanisms.

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ALAN TRENHAILE

SICHELWANNE

Crescent-shaped grooves formed by action of a glacier. The term is derived from Germany, meaning 'sickle tubs', but is part of a collective set of features known as plastically sculptured forms, or p-forms. Sichelwannen (plural) occur in crystalline rocks set in glacial environments, and range in size from 1–10 metres in length, 5–6 metres wide, and from millimetres up to several metres in depth. They are a transverse type of p-form, commonly displaying striations, and with the horns of the crescent shape pointing down-glacier. The origin of sichelwannen is unclear, and several methods of formation have been suggested in the literature. The most plausible manner of formation is erosion by high pressure subglacial meltwater. Similar forms to sichelwannen have been produced by water in less resistant rock, where up-slope topographic obstructions force subglacial channels to bifurcate, producing the characteristic horseshoe pattern on the stoss-side of the obstacle. Erosion by glacier ice has also been proposed, supported by the presence of striations on sichelwannen. However, advocates of a fluvial origin believe the distinctively patchy nature of the striations prohibits formation by ice.

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STEVE WARD

SILCRETE

A terrestrial geochemical sediment arising from low temperature near-surface physico-chemical processes operating within the zone of WEATHERING in which silica has accumulated in, and/or replaced, a pre-existing soil, sediment, rock or weathered material. Silcretes are defined as containing >85 per cent silica by weight, with some pure examples consisting of >95 per cent silica (Summerfield 1983). They can be distinguished from other silica-cemented sedimentary rocks such as orthoquartzites as they often exhibit a porphyroclastic, as opposed to even-grained, texture when viewed in microscopic thin-section (Watson and Nash 1997). Silcretes commonly consist of hard, silica-cemented quartz sand or brittle quartzitic material with a conchoidal fracture. The cement can consist of a range of silica minerals, of which opal, chalcedony, cryptocrystalline silica and quartz are the most widely documented. The presence of other minerals may affect the silcrete colour, with grey, brown and green varieties reported.

Although not as common as many other DURICRUSTs, silcretes are widely distributed in low latitude and other environments, particularly those areas which did not undergo Late Tertiary and Quaternary glaciation. They are found on every continent except Antarctica but are most widespread in inland and southern Australia, and in the Kalahari and Cape coastal regions of southern Africa. Other locations with significant areas of silcrete include Britain (where they are termed 'sarsens') Europe, the Sahara, Tanzania, USA, Uruguay and Brazil. Given their hardness and chemical stability, they are extremely resistant to erosion and often act as CAPROCKs and influence INVERTED RELIEF development. Silcretes most commonly occur as distinct horizons, but may also form a coating on rock outcrops or lenses within other duricrusts such as CALCRETE. Well-developed silcrete horizons are between 0.5 and 3 m thick, although thicknesses of >10 m have been recorded. A variety of terms have been used to describe profiles including massive, columnar, nodular, glaebular and mammilated, reflecting the numerous surface morphologies and modes of origin of many silcretes.

Silcrete development can take place via a variety of pedogenic and non-pedogenic processes, but all require a silica source, a means of transferring this silica to the site of formation, and a mechanism to trigger precipitation. The most significant source is CHEMICAL WEATHERING of silicate-rich rocks, particularly those containing clay minerals. Silica released in this way may then be available for transport in solution. Highly alkaline conditions, such as those found in arid zone lakes, can also lead to extensive dissolution of silicate minerals. Quartz is only weakly soluble in neutral pH terrestrial surface waters (around 10 ppm at 25 °C), but solubility increases dramatically above pH 9.0. Other silica sources include replacement of quartz during carbonate precipitation, dissolution of volcanic and other dust, and biological inputs from silica-rich plants and micro-organisms. Silica from these sources may be transferred in solution to the point of precipitation via lateral or vertical movements of ground water, pore water and surface water, with a range of local and far-travelled silica potentially contributing to silcrete formation. Silica precipitation may also be initiated by a variety of factors, of which the most important are evaporation, cooling, organic processes, absorption by solids, reactions with cations and changes in pH (particularly a shift to below pH 9.0 in alkaline environments).

Silcrete formation by pedogenic processes involves the accumulation of silica from downward percolating soil water during a succession of cycles of leaching and precipitation. Such silcretes commonly consist of a nodular base overlain by a columnar section containing illuviation structures, capped by a brecciated component. Silica mineralogy often varies throughout the profile, with more ordered forms of silica cement in the uppermost sections and less ordered forms towards the base (Milnes and Thiry 1992). Pedogenic silcretes may develop over large areas, are often semi-continuous and relate directly to palaeosurfaces. Non-pedogenic models embrace formation in a variety of settings, including zones of water table fluctuation or groundwater outflow, locations marginal to drainage lines, as well as lakes and seasonal pools. Silica precipitation in these environments is usually controlled by evaporation, pH shifts or organic processes, with the resulting silcretes forming localized sheets or lenses. Non-pedogenic silcretes are usually simple in terms of their macro- and micromorphology, although a range of silica cements may be present. Significantly, they are almost always devoid of the organized profile associated with pedogenic silcretes. Non-pedogenic silcretes may also form at considerably greater depths than pedogenic types and therefore do not normally represent a palaeosurface.

Perhaps the greatest controversy surrounding silcrete is the degree to which environmental controls determine formation, a critical issue if silcretes are to be used as palaeoenvironmental indicators (Nash *et al.* 1994). Silcrete has been suggested to form under climates ranging from semi-arid to monsoonal on the basis of geochemical, mineralogical and micromorphological evidence, and by comparison of the geographic and stratigraphic distribution of silcrete with contemporary and past climate. The situation is made more complex by the fact that most silcretes are relict. Contemporary silcrete formation has only been documented from one location, a hypersaline lake in the Kalahari (Shaw *et al.* 1990). Unfortunately, silicification at this site is driven by organic silica fixation so the resultant silcretes are not an ideal modern analogue. The only safe conclusion that can be made is that the presence of silcrete should not be considered diagnostic of specific environmental conditions. It is essential to establish the mode of origin of any silcrete and view formation within the context of other climate proxies before using it as a palaeoenvironmental indicator. Pedogenic silcretes, such as those in southern South Africa, may have taken hundreds of thousands of years to form and, as a result, have integrated climatic effects over considerable time periods. In contrast, some non-pedogenic groundwater silcretes, such as those in the Paris Basin, developed in a few tens of thousands of years under steady groundwater outflow.

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DAVID J. NASH

SINGING SAND

Two types of sand that emit audible sounds with a coherent wave pattern, on being sheared by wind or other mechanical means, have been reported throughout the world over the past century. They are known as squeaking or singing (found on beaches) in booming (found in sand dunes) sands. The exact mechanism by which these coherent sounds are produced from these sand materials is still unknown.

Many acoustical sands are well-sorted materials with sizes in the range of 100–500 microns, roughly rounded particles with a high content of quartz particles. However, booming sand materials are also found in Kauai, Hawaii with a high calcareous content.

Various material science techniques have been used to show that there is a thin rind-like layer on the particles of these acoustical sand materials. Direct Scanning Electron Microscopic examination of particles that have been sliced by grinding show this layer in both pure silica-gel singing materials and the calcareous Kauai sands. In the former case, the rind layer is composed of amorphous silica and in the latter, an alumino-silicate clay-like material. The rind width is about 5 microns.

Fourier Transform Infra Red (FTIR) analysis has shown that acoustical sand exhibit a broad absorption band in the range of 2,800 cm^{-1}. This band is due to clusters of water in the rind-like layer.

Etching of acoustical sands with hydrofluoric acid (HF) removes the surface layer and causes the sand to become silent. In the case of the Kauai sands, sonication is sufficient to remove this layer and this technique also silences the booming property.

Sand materials that do not sing can be made to do so by grinding the grains in a mill, periodically removing the fines and renewing the water, leaving a well-sorted and highly polished material which 'sings' and exhibits the characteristic FTIR 3,400 cm^{-1} broad band.

Finally, excess heating of these materials also removes this singing or booming sound.

Interest in finding the underlying mechanism of the production of this coherent wave phenomenon in granular materials is due to its possible use in a saser device for sonic hammering.

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DOUGLAS GOLDSACK AND MARCEL LEACH

SINUOSITY

Few natural rivers are straight for more than a few channel widths. Rivers that are not completely straight are sinuous, even if they are not clearly MEANDERING in the sense of having more or less regular oscillations in direction. Schumm (1963) introduced a quantitative definition of sinuosity as the distance along a river between two points A and B, divided by the valley distance between A and B. It is therefore a dimensionless ratio with a minimum value of 1.0 and a realistic maximum of around 3 to 4, after which neck cutoffs occur. It can be calculated for a single bend (when A and B are successive crossovers), a series of bends, or a longer reach.

The sinuosity of a modern channel, or a well-preserved palaeochannel, is readily determined from a map or aerial photograph. Some ambiguities of operational definition must be kept in mind when comparing values quoted by different authors. Is valley distance defined as a straight line, or does it allow for large-scale valley bends? If the river is braided (see BRAIDED RIVER), is the sinuosity computed for the centreline, the biggest channel, or the sum of all the channels as suggested by Richards (1982)?

Sinuosity can alternatively be assessed using variograms and FRACTAL concepts (Nikora 1991; Lancaster and Bras 2002), which can reveal any scale dependence as one moves from single simple bends to compound loops and multiple loops. The sinuosity of a fragmentary palaeochannel can be estimated from the variance of channel direction at places where this can be reconstructed (Ferguson 1977).

Evidently sinuosity varies spatially according to the straight, meandering, or other channel pattern of different reaches. Sinuosity can also fluctuate over years or decades as bends of an actively meandering channel grow and are cut off, and it may change progressively in the event of climate change, flow regulation, or other disturbance of the system.

Sinuosity has significance for fluvial processes and channel regime because it can be written not as a ratio of distances but of slopes: the valley slope divided by the channel slope. Since valley slope is largely inherited, an increase in the meandering tendency of a river leads not only to greater sinuosity but also reduced channel slope. This has consequences for velocity, shear stress, and BEDLOAD transport. Sinuosity is therefore regarded by many geomorphologists and other fluvial scientists as one of several channel properties that can adjust if the bedload supply to a reach is not the same as the transport capacity.

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ROB FERGUSON

SKERRY

A term that describes the low rocky islets common to many mid and high latitude coastlines. Many skerries may be covered at high tide and are subject to wave processes with the result that they may carry the signature of marine quarrying and abrasion. However, in spite of a marine influence, many skerries have also been shaped by past glacial erosion and, within the constraints of the geological structure of the host rock, may have inherited a moulded bedform or even roche moutonée shape. For example, many skerries in Finland, Sweden and Norway, particularly in sites sheltered from severe wave activity, still retain striations etched onto glacially moulded bedforms. This trait can also be seen in the rocky islets of formerly glaciated lakes such as in Loch Lomond in Scotland and in Lake Nasijarvi in Finland. On the coast, fields or chains of skerries frequently occur offshore of areally scoured surfaces that have been partly submerged by Holocene sea-level rise. Good examples occur in the Outer Hebrides of Scotland and in arctic Canada, and along the STRANDFLAT coasts of Norway, Sweden and Finland in the Baltic and in western Iceland.

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JIM HANSOM

SLAKING

The disintegration of a loosely consolidated material following the introduction of water or exposure to the atmosphere (Plate 124). Clays and shales (mudrocks) are especially prone to this form of failure, especially in the presence of saline waters. Materials with high Exchangeable Sodium Percentages (ESP), including some colluvia, may be susceptible, and slaking is an important process on many badland surfaces, including DONGAS. Various tests are available for determining the durability of slaking-prone materials (Czerewko and Cripps 2001) (see WETTING AND DRYING WEATHERING).

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A.S. GOUDIE

Plate 124 Materials with a high Exchangeable Sodium Percentage, such as this colluvial deposit from central Swaziland, are prone to severe erosion as a result of their propensity to slaking following rain events

SLICKENSIDE

A polished, striated rock surface on a fault or bedding plane caused by the frictional movement between one rock mass sliding over another. Slickensides are a common feature on fault planes, and though sometimes can be featureless, they commonly display prominent parallel ribbing or striation. These striations may be exhibited on mineral coatings, such as quartz and calcite, as well as on the rock itself, and can provide an indication of the direction of fault movement from their orientation (they form parallel to the direction of fault displacement). However, striations can often be erased by subsequent fault movement, and thus should only be considered as a record of the most recent fault movement. Additionally, from analysis of a suite of slickenside striations, an estimation of the magnitude of the *in situ* stress field can be established. The origin of slickenside striations is uncertain. Some may be grooves formed by outcropped resistant rock on the opposite fault block, or mineral lineations that grow with their long axis parallel to the direction of fault movement. Slickensides with striations often contain small steps oriented in one direction similarly to the striations.

The term slickenside also refers to natural crack surfaces along planes of weakness in soils, resulting from the movement of one mass of soil against another. This is commonly by the swelling and contraction of soils with high clay levels able to swell (e.g. montmorillonite). Slickenside also refers to the polished surface produced by the passage of a mudflow.

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SEE ALSO: fault and fault scarp

STEVE WARD

SLOPE, EVOLUTION

Landscapes change over time in response to the internal redistribution of sediment, usually with some net export of material to rivers or the ocean. The way in which landscapes and slopes evolve depends on their initial form, the slope processes (see HILLSLOPE, PROCESS) operating and the boundary conditions which determine where and how much sediment is removed. This discussion will mainly focus on two-dimensional slope profiles, but some of the aspects which can only be addressed in 3-D will also be discussed below.

Slope evolution can be described in conceptual development sequences, and much of the history of early twentieth-century geomorphology was concerned with championing alternative conceptual models (Chorley *et al.* 1973), under the banners of W.M. Davis, Walther Penck and others (see CYCLE OF EROSION). More recent approaches have focused increasingly on the application of MODELS in geomorphology, and it is instructive to compare the various development sequences in these terms, to understand the conditions under which each is most appropriate.

Although G.K. Gilbert was not primarily a theoretician, his work in the Henry Mountains (1877) repeatedly interpreted slope profiles in the context of the interaction between form and process, introducing the term DYNAMIC EQUILIBRIUM to describe this balance. He correctly attributed the convexity of divides to SOIL CREEP and similar diffusive processes; and the concavity of the lower slopes to SOIL EROSION processes.

W.M. Davis (1909) spent much of his life canvassing the concept of the Geographical Cycle, which described what he perceived as the 'normal' sequence of erosional landforms, strongly based on his experience of humid temperate conditions. The cycle assumed the rapid uplift of a low relief landscape, and its erosion during a period of tectonically stable conditions. Generally the landscape is taken as soil-mantled, with three life stages. In *youth*, rivers incise deeply into the landscape, and hillsides gradually encroach upon the original surface, parts of which may survive as an *erosion surface*. Slopes become *mature* when the original surface has been consumed, and the highest points begin to undergo appreciable erosion. During maturity, slopes decline in gradient everywhere, forming a connected or *graded* system conveying material to the lowest point. Eventually maturity gives way to *old age* as the surface is reduced to a low relief surface, or PENEPLAIN, which may still retain a few remnant hills, or *monadnocks*. This sequence of ever-reducing relief could be interrupted by relative falls in sea level, which might *rejuvenate* the landscape, perhaps creating flights of partial erosion surfaces which preserved the morphology of previous interrupted cycles. Davis, and his disciples like Johnson and Wooldridge, used the methodology of the geographical cycle to reconstruct former sea levels and the history of the landscape from the inferred remains of high level erosion surfaces and sequences of river and coastal terraces. The two key assumptions of Davis's approach were first that uplift was rapid and followed by long periods of tectonic stability, and second that hillsides were essentially soil-mantled.

Walther Penck (1953 [1924]), working in the much more tectonically active area of the Andes, considered that slope forms responded primarily to the rate of uplift, which he assumed to be a continuous, if episodic, process. In perhaps his central conceptual model, he considered that convex slopes were produced where the rate of uplift was accelerating, and concave slopes where uplift rates were falling. The normal situation, of steady uniform uplift, was associated with slopes of uniform gradient, and these would retreat parallel to themselves while the uplift continued. This approach differs from Davis's in two important respects: first in the assumption of active tectonics, and second in coupling evolution of the slope to conditions at the slope base. Both these assumptions are most relevant where slopes have little or no regolith cover, and all detached material is immediately removed.

Lester King (1953), working in the semi-arid climate of South Africa, proposed a morphologically intermediate conceptual model, in which the steep and rocky upper slopes retreat parallel to themselves, and undergo replacement by a PEDIMENT, which is a lower gradient surface with some regolith cover. Pediments gradually coalesce and eventually consume any residual mountain masses. The combination of rocky and regolith-covered slope elements is significant in this model, while the tectonic assumptions return to the stability advocated by Davis.

In comparing these conceptual models, one of the important distinctions concerns the presence or absence of regolith on the slopes. Where there is a deep regolith, slope processes are able to act at their full capacity, and removal is said to be transport-limited or flux limited. Where the regolith is thin or absent and slopes are steep, material is removed as fast as it is detached by weathering or entrainment, and removal is said to be weathering-limited or supply limited (see WEATHERING-LIMITED AND TRANSPORT-LIMITED). In this case the actual transport rate is well below the *transporting capacity* of the sediment transport processes. Slope evolution is radically different according to which of these two regimes is active, and there is the possibility of an intermediate regime, in which both detachment rates and transport capacities are important. The nature of the regime can be seen from the *travel distance* of material during a transport event. Where travel distance is short compared to the slope length, for example under soil creep, rainsplash, bedload transport or soil erosion, removal is transport-limited; whereas where travel distance is long, for example under rock fall, debris flows, washload transport or movement of solutes, removal is essentially weathering-limited.

Several conceptual models have also been proposed to describe the conditions and forms associated with equilibrium or quasi-equilibrium forms for slope profiles. The essence of these approaches is that the profile form is considered to be independent of the initial surface form which is eroded. J.T. Hack (1960) has been associated with a particular form of DYNAMIC EQUILIBRIUM, in which the landform remains essentially fixed in form as it erodes. This form may strictly occur only where balanced by equal and opposite tectonic uplift, but Hack argued that, during the stage of Davisian maturity, much of the Appalachian landscape eroded with little change in form except close to base level, so that dynamic equilibrium provided a working approximation to observed conditions. M.J. Kirkby (1971) proposed the concept of *characteristic forms*, in which slope profiles retain their form, with more and more subdued relief as they decline in elevation, corresponding to the transition from Davisian maturity to old age. This is seen as a quasi-equilibrium in which the profile form is characteristic of the ensemble of processes acting on it.

By comparing these qualitative concepts with simple slope models, the conditions for all these simplified forms can be compared in a quantitative way. For a slope profile, evolution is controlled by four sets of constraints. First, mass must be conserved; second, evolution takes place from an initial profile form; third, evolution is subject to boundary conditions, which define, for example, conditions at the top (divide) and bottom (stream) of a hillslope; and fourth, evolution occurs through sediment transport by one or, usually, more slope processes. The transporting capacity for each process varies in some way with topography, usually with distance from the divide and gradient; and removal is also subject to transport or weathering-limited conditions, which can be defined by a fuller specification of the slope processes.

The Mass Balance equation for a slope profile states that:

Input − Output = net increase in Storage

or in the simplest case of a simple slope profile from divide to stream:

$$\frac{\partial z}{\partial t} + \frac{\partial S}{\partial x} = 0 \qquad (1)$$

where S is sediment transport per unit width, z is elevation above an arbitrary datum, x is horizontal distance measured from the divide, and t is elapsed time.

The cases of weathering and transport-limited removal can be distinguished using a sedimentation equation:

$$\frac{dS}{dx} = D - \frac{S}{h} \qquad (2)$$

where D is the rate of sediment detachment, and h is the travel distance (the mean distance travelled by sediment in an event). The second term of the right-hand side is the rate of deposition. Where sediment detachment balances sedimentation, the flow is said to be at its travel capacity, $C = D \cdot h$.

Where the travel distance is small (in relation to the slope length), then removal is essentially transport-limited and $S = C$, so that equation (1) can be used to define slope evolution, with $\mathbf{C}$ replacing S. Where the travel distance is long ($h \gg x_0$), removal is weathering-limited and slope evolution is approximated by equation (2) with the second term on the right-hand side negligible. Between these extremes, it is necessary to retain both equations (1) and (2) to determine how slopes evolve.

The simplest boundary conditions are of a summit fixed in horizontal position (at $\boldsymbol{x} = 0$), and a stream fixed at some point $x = x_0$.

For several processes, capacity sediment transport rates can be written in the form:

$$C \propto x^m \Lambda^n \qquad (3)$$

For example simple but empirically acceptable formulations are shown in Table 43, although EQUIFINALITY allows some range of possible exponent values. Mass movements may also be included in a similar scheme, but with two threshold gradients: a lower threshold below which no movement occurs, and an upper threshold above which material will never stop. The lower threshold Λ_T corresponds to the maximum stable slope gradient under saturated conditions, and the upper threshold Λ_0 to the angle of repose for debris. A simple formulation then takes the form:

$$D \propto \Lambda(\Lambda - \Lambda_T);\ h \propto 1/(\Lambda_0 - \Lambda);$$
$$C \propto \Lambda(\Lambda - \Lambda_T)/(\Lambda_0 - \Lambda) \qquad (4)$$

with the capacity, $\mathbf{C}$, defined only in the range $\Lambda_T < \Lambda < \Lambda_0$, and travel distances generally of the order of the total slope length. By treating mass movement as a continuous process, the feedbacks produced by the size of an individual slide event are ignored, so that this formulation works best for small and shallow slides, and for rockfalls from cliffs.

Without entering into a full analysis of these equations, some results can be quoted here without proof. First, the conditions for downcutting at a steady uniform rate, T corresponding to a strict application of Hack's dynamic equilibrium, is that:

$$S = Tx \qquad (5)$$

This is necessarily true because, in the steady state, the slope processes must carry away all the material eroded upslope.

For any transport-limited removal process (i.e. S = C), equation (3) gives:

$$C = Tx \sim x^m \Lambda^n, \quad \text{or} \quad \Lambda \sim Tx^{(1-m)/n} \qquad (6)$$

Thus, for steady-state downcutting, hillslopes are convex (gradient increasing downslope) when the distance exponent $m < 1$, and concave if $m > 1$. This means that slopes become convex for soil creep, rainsplash, solifluction and tillage erosion, and become concave for rillwash. More realistically, a combination of processes is acting, for

Table 43 Exponents for some sediment transport processes in equation (3)

Process	Travel distance	m	n
Soil creep	Small	0	1
Rainsplash	(< 1 m)	0	1
Solifluction	(< 1 m)	0	1
Tillage erosion	(<1 m)	0	1
Rillwash	~10 m	2	1–2
Solution	≫1 km	1	0

example rainsplash and rillwash. Adding these terms the combined transporting capacity may be written as:

$$C = k\Lambda\left[1+\left(\frac{x}{u}\right)^2\right] \tag{7}$$

where u is the distance beyond which the rillwash term (the second term on the right-hand side) becomes greater than the rainsplash term (first term on RHS). Thus for $x<u$, rainsplash is the dominant process, and for $x>u$, rillwash is dominant.

Solving for constant downcutting as before,

$$\Lambda = \frac{Tx}{\left[1+\left(\frac{x}{u}\right)^2\right]} \tag{8}$$

In this case, gradient increases for $0<x<u$, and gradient decreases for $x>u$. In other words the constant downcutting form is convex where rainsplash is the dominant process and concave where rillwash is the dominant process (Figure 152).

This relationship is not, however, universal, and if downcutting decreases downslope, the concavity expands slightly into the rainsplash-dominated zone. A simple representation of Davisian decline can be made by assuming, instead of constant downcutting, a rate of downcutting which is directly proportional to height above the basal point. With this assumption, the whole hillslope must eventually erode to a flat uniform base-level plain, and equation (3) is replaced by:

$$-\frac{\partial z}{\partial t} = \frac{\partial C}{\partial x} = \alpha z \tag{9}$$

for an appropriate constant α. Although there is not always a simple analytical solution to this equation, the difference between this Davisian hillslope and the constant downcutting form is quite clear from Figure 153. In each case, the divide convexity begins the same, but, for the declining form, the rate of increase in gradient is less than for the constant downcutting form, and the concavity, where rillwash is present, is broader. It can be shown that the shape of both the declining and constant downcutting forms depends primarily on the combination of processes operating, and the effect of the initial form is progressively obliterated over time. This means that physical remains of former eroded land surfaces (erosion surfaces and terraces) only survive for a limited period of time before they disappear from the landscape, usually surviving longest in flat areas and along divides, where denudation is initially least.

Where travel distances are long, then removal of material is approximately *weathering-limited*. In this case slope development follows equation (2) above, with the final term negligible (because h is large), giving, when combined with equation(1):

$$-\frac{\partial z}{\partial t} = \frac{dS}{dx} = D \tag{10}$$

For the case of mass movements, using equation (3) and writing Λ as $-\frac{\partial z}{\partial x}$:

$$\frac{\partial z}{\partial t} = (\Lambda - \Lambda_{\mathrm{T}})\frac{\partial z}{\partial x} \tag{11}$$

The solution to this equation shows lateral retreat of steep slopes at a horizontal rate of $(\Lambda - \Lambda_T)$. This evolution of the landscape essentially describes a parallel retreat of the landscape

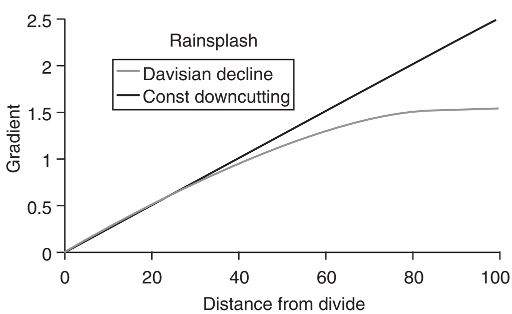


Figure 152 Slope evolution where rainsplash is dominant

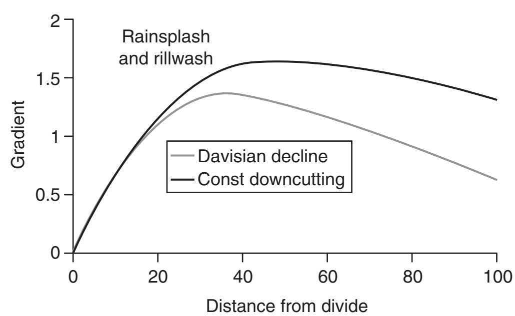


Figure 153 Slope evolution with rainsplash and rillwash

in a way which is close to the conceptual models of Penck and King. Although there is no space to fully develop this argument here, there is a strong association between *transport-limited* removal and Davisian decline of slopes, and between *weathering-limited* removal and lateral retreat of landforms. Because this distinction is closely linked to the presence or absence of a regolith deep enough to allow transport processes to operate at their capacity rates, there is also a general association between these two sets of conditions and both climate and tectonics. There is a link to climate because, in dry climates, there is little water flowing through the soil, consequently little leaching and slow bedrock weathering. Most weathering products are therefore removed by erosional processes before fine-grained and deep soils can develop. On the contrary, humid climates allow more rapid weathering, converting parent materials to fine-textured soils before it is eroded. Steep slopes increase erosion processes, but have much less effect on weathering rates, so that soil is thinner and stonier than on gentle slopes. Active tectonic uplift also plays an important role, by creating and maintaining steep slopes.

From this discussion it may be seen that the various qualitative conceptual models have been significantly shaped by the experiences of their authors. Davis, working in the humid-temperate areas of north-east USA, focused on transport-limited processes, and slope decline towards eventual peneplains. Penck and King, working in more tectonically active and more arid areas, saw the lateral retreat of steep slopes, generally with shallow regolith. It is clear, however, that, although many details are still poorly understood, a single set of principles can be widely applied, and takes in the early conceptual models as special cases.

Figure 154 provides cartoons of 'typical' slope evolution from a plateau with a steeply incised stream, but without further tectonic uplift as material is removed by the basal stream. It can be seen that elements of both transport and weathering-limited removal occur in both settings while slopes are steep, but that weathering-limited elements survive much longer under semi-arid conditions.

Although most of the features of a hillside may be described in a slope profile, the whole scale of the landscape is a problem which can only be addressed in three dimensions, defining the typical length of a single slope, and the DRAINAGE DENSITY of the landscape, which are related by the relationship:

$$\text{Mean slope length} = 1/(2\ \text{DD}) \tag{12}$$

Following the ideas developed by Smith and Bretherton (1972), it is argued that, where sediment transport processes increase more than linearly with catchment area, any small irregularity in the landscape will tend to grow with positive feedback until it develops into a valley. The threshold at which this occurs determines the drainage density of the landscape. A 3-D *landscape model* is able to demonstrate this behaviour. Near the divides, any small irregularities in the landscape become smoother over time, whereas downslope, some hollows grow into valleys. The form of the sediment transport equations, such as equations (3) and (4) above, determine the threshold of this valley instability. For example, if the combination of rainsplash and wash is put in the form (note that this is a different form from equation (7) above):

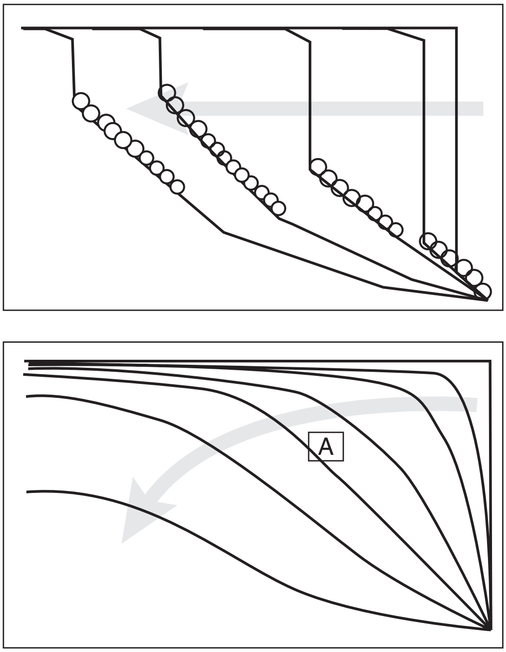
$$S = C = k\Lambda\left[1+\Lambda\left(\frac{x}{u}\right)^2\right] \tag{13}$$

then it can be shown that the critical distance for hollow enlargement is:

$$x = u / \sqrt{\Lambda} \tag{14}$$

Table 44 Conditions associated with transport-limited and weathering-limited removal

	Transport-limited	*Weathering-limited*
Regolith	Deep enough to allow transport processes to operate	Shallow and generally stony
Climate	Humid temperate	Semi-arid
Gradient	Gentle	Steep
Tectonics	Inactive	Active
Dominant erosion processes	Creep, rainsplash, rillwash	Mass movements
Ratio of weathering to erosion	High	Low



Semi-arid slope evolution
Upper plateau influenced by rainsplash etc. Steep escarpment (often defined by lithology) and boulder slope influenced by mass movements. This is the only section undergoing parallel retreat. Lower concavity modified by rillwash and stream incision.

Humid-temperate slope evolution
Upper plateau influenced by creep or solifluction. Until A, lower slope dominated by mass movements, at reducing rate as slope towards a landslide-stable gradient. After A, slope dominated by creep and rillwash, declining towards base-level peneplain.

Figure 154 Slope evolution from a plateau with a steeply incised stream under semi-arid and humid-temperate conditions

This expression suggests that drainage density should be greater in steeper areas, and this forecast is supported by empirical evidence, particularly Dietrich and co-worker's data from California (Dietrich and Dunne 1993). Thus the spacing of streams, and so the whole scale of the landscape, is also set by the balance between slope and stream processes, and valleys occur where the processes driven by water flow begin to predominate over processes driven mainly by gradient, such as creep or rainsplash. This view of the landscape indicates not only that drainage density varies over the landscape in relation to steepness, but that it evolves through time as relief changes through erosional lowering or tectonic uplift.

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MIKE KIRKBY

SLOPE STABILITY

Slope stability and its corollary slope instability, are defined as the propensity for a slope to undergo morphologically disruptive processes, especially landsliding (Plate 125) (see LANDSLIDE). Slow distributed forms of MASS MOVEMENT such as soil creep are generally not considered sufficiently disruptive to be included in this definition. From a hazard and engineering perspective, assessments of slope stability are focused on periods ranging from days to decades. However, slope stability may also be treated as a component of landform evolution and therefore its significance can only be judged by taking into account much longer periods of time.

In every slope, there are stresses which tend to promote downslope movement of material (shear stress) and opposing stresses which tend to resist movement (shear strength). In order to assess the degree of stability, these stresses can be calculated for a failure surface within the slope and compared to provide a FACTOR OF SAFETY (defined as the ratio of shear strength:shear stress). In a static slope, shear strength exceeds shear stress and the factor of safety is greater than 1.0 whereas for slopes on the point of movement shear strength is just balanced by shear stress and the factor of safety is 1.0.

While engineering codes of practice may specify a particular factor of safety for completed earthworks, it is not the most meaningful representation of slope stability. Two slopes that have the same factor of safety but large absolute differences in excess strength (i.e. strength minus shear stress) can be used to illustrate the limitations of the factor of safety. For example, the strength to stress ratios, in unspecified stress units for a slope (A) of 400/200 and for a slope (B) of 200/100 both yield a factor of safety of 2.0. However, slope (A) has an excess strength of 200 units while slope (B) has an excess strength of only 100 units. As excess strength is the quantity that must be entirely reduced (by reduction in strength or increase in shear stress) in order to produce failure, it represents the 'margin of stability' or inherent resistance to failure. Instability, however, is determined not only by the margin of stability of the existing slope but also by the magnitude of (external) destabilizing forces which may affect the slope to reduce that margin.

Plate 125 Actively unstable slopes, subject to deep-seated earthflows, Poverty Bay, New Zealand. Photo: Ministry of Works, New Zealand

Slopes can therefore be viewed as existing at various points along a stability spectrum ranging from high margins of stability with low probabilities of failure at one end to actively failing slopes, with no margin of stability, at the other. It is useful to define three theoretical stability states along this spectrum based on the ability of dynamic external forces to produce failure. First is the 'stable state', defined as slopes with a margin of stability which is sufficiently high to withstand the action of all dynamic destabilizing forces likely to be imposed under the current environmental/geomorphic regime. Second is the 'marginally stable state', represented by static slopes, not currently undergoing failure, but susceptible to failure at any time that dynamic external forces exceed a certain threshold. Third is the 'actively unstable state', represented by slopes with a margin of stability close to zero and which undergo continuous or intermittent movement. The margin of stability possessed by a slope is a measure of its landslide susceptibility and, together with the frequency and magnitude of dynamic destabilizing factors, provides a measure of probability of failure. In turn, the probability of failure together with its magnitude provides a measure of landslide hazard.

The concept of three stability states offers a useful framework for understanding the causes and development of instability. In this context four groups of destabilizing factors can be identified on the basis of function (Figure 155).

1 *Preconditions* (predisposing factors) are static, inherent factors which not only influence the margin of stability but more importantly in this context act as catalysts to allow other dynamic destabilizing factors to operate more

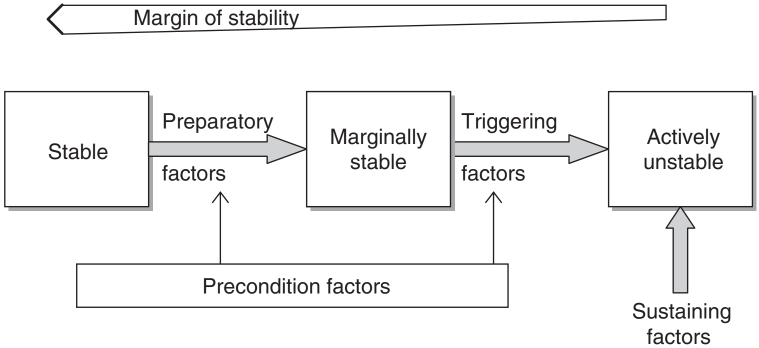

Figure 155 Stability states and destabilizing factors

effectively. For example, slope materials that lose strength more readily than others in the presence of water predispose the slope to failure during a rainstorm; or a particular orientation of rock structure may enhance the destabilizing effects of undercutting.

2 *Preparatory factors* are dynamic factors that by definition decrease the margin of stability in a slope over time without actually initiating movement. Hence, facilitated by preconditions, they are responsible for shifting a slope from a 'stable' to a 'marginally stable' state. Some factors, such as reduction in strength by weathering, climate change and tectonic uplift, operate over long periods of geomorphic time whereas others may be effective in shorter time periods e.g. slope oversteepening by erosional activity, deforestation, or slope disturbance by human activity.

3 *Triggering factors* are those factors which initiate movement, i.e. shift the slope from a 'marginally stable' to an 'actively unstable' state. The most common triggering factors are intense rainstorms, seismic shaking and slope undercutting.

4 *Sustaining* factors are those that dictate the behaviour of an 'actively unstable' slope e.g. duration, rate, and form, of movement.

As most slopes are stable for most of the time, the onset of slope instability from a geomorphic perspective represents an effective hillslope response to a destabilizing change in the boundary conditions, enabling a rapid adjustment and eventual return (or tendency toward) more persistent landscape forms. Thus, given sufficient time, the process of landsliding tends to stabilize a slope by reducing slope angle, height or weight, or by removing susceptible material.

The concept of slope instability may usefully be broadened to include any significant adjustment in processes or forms that tend to change the equilibrium conditions on a slope. For example, a slope where mature soils are being depleted by the onset of gullying can be considered to be undergoing a phase of slope instability. Whether that adjustment is viewed as a perturbation of a system in dynamic equilibrium or a change to another equilibrium state depends on the timescale being considered.

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MICHAEL J. CROZIER

SLOPEWASH

The term 'slopewash' is not well defined in geomorphological literature. It has been used interchangeably with terms such as 'surface wash', 'rainwash', 'unconcentrated wash' and 'rillwash', now largely replaced by more precise terms such as 'overland flow', 'sheet wash', 'rainflow', 'rain-impacted flow', and 'rilling' or 'rill erosion'. The process was first described by McGee (1897), and defined by Bryan (1922: 29) as 'the water from rain, after it has fallen on the surface of the ground and before it has concentrated into definite streams'. Surface wash and rainwash were extensively invoked in early geomorphological literature as processes responsible for concave hillslope profiles, for hillslope profiles with a marked concave 'break', and for the formation of rock-cut pediments in southern Africa (King 1949). However, few field studies (e.g. Schumm 1956) attempted to measure the process, and it was not until Emmett's (1970) study of the hydraulics of overland flow on hillslopes that the complexity of the process was recognized. Subsequent studies on agricultural soils have identified the precise interacting processes involved, which include rainflow, sheet wash and rill erosion. The processes combined in slopewash are frequent on disturbed agricultural soils but their significance on natural hillslopes is less clear. The extensive thin sheets of water described by McGee are rare, even in dry regions where conditions are most favourable. In all areas, overland flow occurs most frequently on bare, relatively impermeable rock surfaces. Depending on the rock, such surfaces will usually yield some fine-grained material which can be transported by shallow flows, but the dominant transport in most cases is probably solution. Overland flow generated where rainfall exceeds surface infiltration capacity (Hortonian overland flow) occurs quite frequently on regolith-covered hillslopes, but is usually localized, reflecting patchy vegetation cover, microtopography, variations in rainfall intensity and marked variations in surface infiltration capacity. This produces a complex, varied interaction of processes along the hillslope, rather than the orderly transition from rainsplash to sheet wash to rill erosion envisaged in earlier literature. Instead, on most hillslopes patches of erosion, erosional lag deposits and sedimentation are interspersed in complex patterns. Often these are random and irregular, but on dryland hillslopes, the patterns are often regular, caused by the shrink–swell characteristics of smectite clays, the moisture requirements of shrubby plants, or selective transportation of material by pulsatory flows. These features may also occur where flow is not Hortonian, but generated as saturation excess, as return flow or as seepage. Such flows are more predictable and usually more uniform across the hillslope and therefore may more frequently cause profile concavity.

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RORKE BRYAN

SLUSHFLOW

Slushflows, also called slush avalanches, are water-saturated snow masses flowing principally along a first-order stream channel (Larocque *et al*. 2001). Their formation is associated with an increase in the water content of a snowpack through rainfall, or through rapid snowmelt or through a combination of both. At a critical point instability occurs and snow mass is released. They are widespread in artic, subarctic and alpine environments but may, unlike snow avalanches (see AVALANCHE, SNOW), be initiated on relatively low angle slopes (Nyberg 1989). They are capable of transporting a high debris load over long distances, and also of causing substantial abrasion and erosion. The material they deposit can form such features as boulder tongues. They can also cause exceptional rates of glacial ablation (Smart *et al*. 2000) and pose a hazard to engineering structures.

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A.S. GOUDIE

SOIL CONSERVATION

Society can try to minimize SOIL EROSION by soil conservation practices, which may be 'active' or 'passive'. Active soil conservation includes positive actions to decrease erosion rates, such as terracing or contour farming. 'Passive' conservation is avoiding actions, such as ploughing on steep slopes or overgrazing erodible soil. Passive conservation is just as valid and often much cheaper than active conservation. We also need to distinguish between soil conservation and sediment control. Soil conservation is taking action to prevent erosion and keep soil *in situ*. Sediment control deals with soil which has already been eroded and transported, keeping it within fields or removing it from water courses.

Conservation strategies can be complex and varied. However, Morgan (1995) proposed that soil is conserved by decreasing EROSIVITY, decreasing ERODIBILITY, improving vegetation protection, or any combination of these. Ancient features, such as the terraced fields of South-East Asia, show that soil conservation has long been employed. However, soil conservation technologies were considerably improved by the US Soil Conservation Service.

The US Midwest suffered severe soil erosion in the 1930s or 'dirty thirties'. By this time, many of the organic soils had been cultivated for over eighty years, so soil organic contents were falling, thus increasing soil erodibility. Many erodible soils were brought into cultivation due to increased grain prices during the First World War. Then, in the 1930s, there was a severe drought. This combination of circumstances led to severe erosion, especially wind erosion, in the Great Plains States, which became labelled 'The Dust Bowl'. The social upheaval associated with these events was graphically portrayed in the novel *The Grapes of Wrath* (Steinbeck 1939), which tells the tragic struggle of a family from Oklahoma.

In response to these problems, an agricultural engineer, Hugh Hammond Bennett, led a campaign to promote soil conservation. Bennett was almost evangelical in his campaigning among farmers and politicians. During one presentation to the US Congress, a dust storm blew into Washington, DC. Bennett declared, pointing out the window, 'there, gentlemen, goes the State of Oklahoma!'. Congress then allocated funding for the foundation of the US Soil Conservation Service in 1934.

Various soil conservation techniques have been proposed and conservation projects can adopt one or a combination of techniques. The choice depends on many factors, including environmental (climate, topography and soil type) conditions and social, economic and political circumstances.

Windbreaks can be very effective in decreasing wind velocity and thus erosivity. They are usually aligned perpendicular to the direction of the most frequent erosive winds and are effective for 10–20 times their height downwind. Dense windbreaks greatly decrease wind velocity, but velocities soon increase in their lee. Better effects are achieved with more permeable windbreaks; velocity is not decreased so much, but their effectiveness in the lee is greater. For maximum benefit, windbreaks with a porosity of about 50 per cent and a mix of heights are recommended.

Hedgerows are effective windbreaks and their large-scale removal from the British countryside since the 1940s has contributed to increased wind erosion of arable soils. Hedgerows also protect against water erosion, by dividing slopes into shorter sections. Their removal allowed erosive runoff to operate over effectively longer slopes, thus increasing erosion risk. The restoration of all hedgerows is not feasible, as modern farm machinery cannot operate efficiently on fields as small as those which dominated the British countryside before the Second World War. However, it is important to identify 'key hedgerows', that protect areas exposed to predominant wind directions and convex slope sections susceptible to water erosion. Their retention or establishment should separate large catchment fields from lower convex–concave slopes. The absence of hedgerows integrates fields into geomorphological systems that are vulnerable to water erosion.

Slope management is a key component of soil conservation strategies. Simply leaving steeper erodible soils with a vegetation cover is a cheap, but effective, form of passive conservation. 'Set-aside', which involves taking areas out of crop production and leaving them with a permanent vegetative cover, was originally a means of decreasing grain surpluses. Carefully targeted on steeper erodible slopes, it could be an effective means of achieving both agricultural and environmental objectives.

Terracing is the most spectacular form of soil conservation and involves dividing slopes into a series of steps, cultivating the flatter sections and protecting the 'riser' with vegetation or masonry walls. In South-East Asia, terraces are extensively used to grow rice. To retain water, small earth walls or 'bunds' are built on the lower sides of terraces. However, terracing poses several problems. First, the risers take up about 10 per cent of land, though this is usually compensated by increased crop yield associated with increased soil water storage. Second, construction and maintenance are costly in terms of human resources; most of the world's areas of extensive terraces were constructed over many generations. Third, it is often difficult to operate machines efficiently on terraces.

Controlled colluviation is particularly applicable in semi-arid countries with a distinct rainy season. Lines of stones are laid out along the contour. When the seasonal rains arrive, soil is eroded and sediment deposited on the upslope side of the stones. Over a few rainy seasons, a fine silty moisture-retentive colluvial soil accumulates and, in semi-arid climates where soils tend to become saline, seasonal flushing with water can desalinize the COLLUVIUM, making it relatively fertile and suitable for crops. The technique is simple, cheap and thus affordable in poorer countries. Check dams operate in a similar way. Walls are constructed across gullies, sediment collects behind them and is periodically removed. Additionally, obstacles (such as straw bales and willow fences) are often placed in gullies to impede runoff and thus encourage sedimentation.

Contour farming involves orientating agricultural operations along the contour, rather than up-and-down slope and is particularly applicable on gentle uniform slopes used for mechanized agriculture (e.g. the Prairies and the Steppes). Complex slopes tend to limit the applicability of contour cultivation in northern Europe. However, slopes must not be too steep (>15°), as water can accumulate in furrows and eventually breach the ridges between them, causing even higher soil erosion rates than the more common up-and-down slope cultivation. Moreover, farm machinery cannot operate safely or efficiently along the contours on steep slopes.

In strip cropping systems, alternate strips of land are arranged perpendicular to the relevant erosive agent, wind or water. The crops themselves protect the soil, braking the velocity of the erosive agent and trapping sediments. Temporary grassland (leys) can form part of a strip cropping system and also increase soil organic matter, lowering soil erodibility. Usually, strips are 15–45 m wide, becoming wider as erosion risk increases.

Rotation is a well-established agronomic technique, by which different crops are grown in an established sequence. A temporary grass ley is usually an integral component of rotational systems, allowing natural recovery. Twentieth-century development and mass production of chemical fertilizers allowed continuous arable production, without temporary leys. Fertilizers provided ample supplies of macronutrients needed for crop production but, on many soils, allowed soil organic contents to decrease, so that erodibility increased. The increased incidence of erosion on arable soils in much of North America and Western Europe has been attributed to long-continued arable cultivation.

Addition of organic matter can decrease soil erodibility. The most common is farmyard manure (FYM) and there are many commercial organic manures. 'Green manures' are crops such as clover and mustard which grow quickly, producing much biomass, but rapidly decompose to increase the soil organic matter. In developing countries, human waste is used. This is usually transported and applied at night and is known as 'night soil'.

Mulching is the use of vegetative or other material on the soil surface, to simulate the protective effects of vegetation. Often, residues from the previous crop are applied, such as straw on wheat fields. Many studies have shown mulching is very effective, especially in the tropics and subtropics. However, in temperate environments, mulches can insulate the soil and prevent it warming in spring.

Hydromulching is particularly applicable to engineered slopes, such as road cuttings and construction sites. Mulch consists of a mixture of materials, such as fibre, straw, paper and shredded wood and is sprayed with seed, protecting the soil surface while seeds germinate and establish a protective vegetation cover. Hydromulching is expensive and used only in high value projects.

Geotextiles are cloths used to protect soil surfaces. Usually, they consist of biodegradable material, such as jute, have a coarse mesh and last for a short time, usually less than two years. This is long enough for seed mixtures to establish protective plant communities. There are also non-biodegradable geotextiles, which are used for permanent stabilization of, for example, channel banks.

Compaction affects the hydrological and thus erosional behaviour of soil, decreasing the size and interconnectivity of soil pores and impeding infiltration, so that runoff and erosion are increased. Increasing use of heavy farm machinery is exacerbating compaction problems. Compaction by animals also poses problems, as they produce fairly small compact hoof imprints. On wet soils, this weakens soil structure, and gives soil a 'puddled' appearance, a process known as poaching.

There are many methods of minimizing soil compaction. Trafficking wet soil should be avoided, though a shallow tillage tool mounted behind the tractor wheel can break up compacted soil. Tramlines, along which all passes for agricultural operations are made, limit compaction to narrow zones. Even within tramlines, it is possible to diminish compaction by using larger tyres or low-ground pressure vehicles (LGPV) to spread the load. Ploughing in very wet conditions and/or with blunt plough shares can compact subsoils. This 'plough pan' must then be disrupted by 'subsoiling', using a deep blade with a 'shoe-like' structure at the end, mounted behind a powerful tractor.

Several crop production methods (known variously as crop residue management systems, no-tillage, minimum cultivation, conservation tillage or direct drilling) minimize compaction, runoff and loss of organic matter by avoiding ploughing and preserving a cover of crop residues. The various terms partly reflect the diversity of systems in use. Residues from the previous crop are left on the soil surface, to simulate the protective effects of vegetation. Further crops are then sown into the residue with minimum disturbance and eventually provide a protective vegetative cover. These systems allow crop production on steeper slopes without increasing erosion risk. Increased soil organic matter also improves moisture retention and encourages earthworms, which increase infiltration rates. There are disadvantages, as lack of tillage can allow weed infestation, particularly by perennial grasses. As compaction can increase without tillage, especially on weakly structured silty soils with little organic matter, a crucial factor for success is soil faunal activity, especially earthworm activity. A rich earthworm population can effectively till the soil on behalf of the farmer. This is why conservation tillage systems became popular on organic-rich soils in both North and South America. There is a gradation from full 'no-till' to traditional ploughing. The selected method depends on the various advantages and disadvantages. A minimum cultivation system, with occasional full ploughing, can be an acceptable balance.

Chemical soil conditioners can decrease soil erodibility, binding particles together, improving aggregation and increasing infiltration rates. Conditioners, such as 'Krilium', 'Flotal' and 'Glotal', were developed in the 1950s and 1960s, and were followed by many ionic and non-ionic conditioners. Field and laboratory experiments showed they were effective, but high cost restricted their use to high value crops or specialist engineering applications (e.g. stabilizing road cuttings, engineered slopes, oil-well heads, temporary helipads and airfields).

Successful soil conservation is not simply an engineering problem, but is a complex amalgam of agronomic, social, economic and political considerations. It is essentially a team effort, requiring the participation of national, provincial and municipal government, farmers, scientists, extension workers and agricultural advisers. Also, it is essential that an effective dialogue develops between soil conservationists and local people, as their involvement, support and participatory agreement are crucial.

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MICHAEL A. FULLEN AND JOHN A. CATT

SOIL CREEP

Slow MASS MOVEMENT processes are generally grouped together under the term soil creep (Kirkby 1967). A distinction is made between continuous creep, which is driven directly by downslope shear stresses, and seasonal creep which is the downslope component of movements which are either randomly directed, or primarily perpendicular to the soil surface.

Continuous creep may extend to depths of 5 m or more, and is driven by gravity shear stresses against the frictional and cohesive strength of the soil. The relationship between stress and strain for soils usually shows three zones. At low stress there is no movement; at high stress there is a more or less linear deformation, at a rate proportional to stress above an apparent failure threshold; at stresses below this threshold there is slow and non-linear increase in strain which bridges between these two behaviours (Figure 156). Continuous creep occurs in this non-linear transitional zone. Clearly continuous creep can occur over only a narrow range of stress–strain conditions, and is, in most cases, associated with the larger mass movements which occur when stresses appreciably exceed the threshold. In

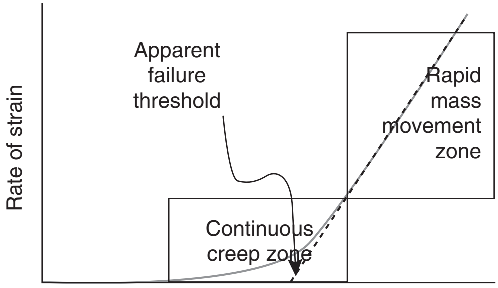

Figure 156 Stress–strain relationship in continuous creep

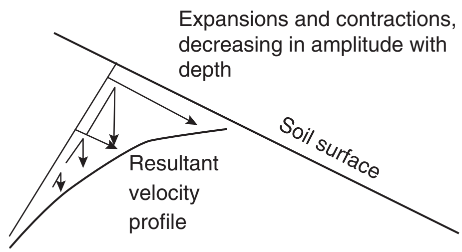

Figure 157 Zigzag seasonal creep movement

many soil materials, the creep itself changes the stress–strain relationship as the soil is reworked by movement.

Seasonal soil creep is driven by a number of processes, some of them biogenic and others driven by the climate. Biogenic movements are more or less random in direction, and include wedging by plant roots, movement of soil by burrowing animals such as gophers, earthworms and termites. Random movements lead to a net migration of material from zones of high soil concentration to zones of lower concentration, and may be considered as a form of diffusion.

The net direction of this random migration is towards the free upper surface of the soil, and it is eventually balanced by resettlement under gravity. The two main climate drivers are WETTING AND DRYING WEATHERING cycles and FREEZE–THAW CYCLES. These cycles lead to an upward expansion or heave during wetting or freezing, followed by a downward settlement during drying or thawing. On a slope these expansions are at right angles to the slope surface, while settlement tends to occur more nearly along a vertical (Figure 157). Thus both random movements and heaves driven by climate produce a net movement which consists of zigzags which move material slowly downslope. To a first approximation the rate of this movement should be proportional to the slope gradient, but this theoretical inference has never been completely validated, because of the large observed variability in creep rates over small areas. All these seasonal soil creep processes are commonly referred to as *diffusive processes*, and have in common a linear dependence of transport rate on slope gradient, and little or no dependence on distance from divide or catchment area because they do not depend on the flow of water. This balance between outward diffusion and settlement under gravity not only moves material downslope but is also one of the main processes responsible for the observed increase in pore space towards the surface of uncultivated soils.

It can be seen that the rate of seasonal soil creep is limited by the depth and amplitude of these random and heave movements, which are always small. Thus there can never be a transition from seasonal soil creep to larger and more rapid mass movement, and it rarely extends to depths where the movement reworks material along potential mass movement failure surfaces. In different areas, the dominant processes reported to drive seasonal soil creep can be almost any of those listed above, according to the faunal and climatic activity in the soil. Seasonal soil creep rarely penetrates to depths of more than 30 cm, and surface translational movements reported generally lie in the range of 2–5 cm per year. If movement is proportional to gradient, then a widely quoted diffusivity for soil creep is $10^{-3}\,m^2\,yr^{-1}$. This means that on a 10 per cent slope, the actual transport will be $10\% \times 10^{-3}\,m^2\,yr^{-1}$, or 1 $cm^2\,yr^{-1}$. These values are low, of the same order of magnitude as those quoted for rainsplash (see RAINDROP IMPACT, SPLASH AND WASH), whereas the highest values reported for SOLIFLUCTION are 10–100 times greater. Both these processes are commonly included with soil creep in the category of diffusive processes, at least to a first approximation. Another important anthropogenic diffusive process is termed *tillage erosion*, in which ploughing and other forms of cultivation produce considerable net downslope movement of material, even if successive passes of the plough are along the contour, and alternately turn the soil up- and down-slope. Rates of tillage erosion are currently 10–1,000 times greater than for soil creep or rainsplash, and have increased greatly with the more widespread use of heavy farm machinery.

G.K. Gilbert (1909) recognized the importance of diffusive processes in creating convex hilltops, which are found on almost all soil-covered slopes. The convex area is usually relatively smooth, and may give way to an area where rills or larger channels begin, or may lead directly into an area where slopes become more uniform in gradient. The width of this convexity varies widely, primarily with climate and lithology, and generally lies in the range 1–1,000 m. Typically narrow convexities are associated with high DRAINAGE DENSITY and broad convexities with low drainage density. The relationship between diffusive processes and hillslope convexity is based on the assumption that transport by diffusive processes increases with gradient, and little or not at all with distance from the divide (or catchment area). If the area around the divide is eroding in the long term, then the volume of material which must be transported to remove the erosion products necessarily also increases continuously away from the divide. The diffusive processes can, by definition, only transport this increase in material through an increase in gradient. If the REGOLITH is deep enough not to constrain movement of material, it necessarily follows that gradient must also increase continuously from the divide, i.e. that the slope profile is convex.

This argument makes very few assumptions, so that the conclusion – that divides are convex – has great generality, and requires only that diffusive processes are dominant close to the divide. It is also fair to invert the argument, since this is a necessary and sufficient condition, and state that divide convexity demonstrates the dominance of diffusive processes close to the divide, and indeed roughly outlines the zone in which they are dominant. In the simplest realistic case, denudation of the area around the divide may be assumed to be constant, at rate T. At distance x from a ridge crest, an amount $T \cdot x$ must be exported per unit length of ridge crest to carry away the eroded material. If the rate of transport by creep processes $= K \cdot s$, where s is slope gradient and K is the rate constant, then to balance the erosion against the transport rate, $T \cdot x = K \cdot s$ or the slope must evolve until the gradient $s = T/K \cdot x$. In other words the gradient must eventually increase linearly with distance from the ridge crest.

It is useful to consider what happens if the assumptions of this argument are not met. First, on a long valley side, there may be deposition at the base of the slope, so the assumption of uniform erosion breaks down, and diffusive processes may be associated with concavity at the slope base. Second, if the regolith is thin over a resistant rock layer, then the K value may drop where the soil is thin, and rise again below the resistant layer, so that there is exaggerated convexity over the rock band, and a short concavity below, as the K value rises to its previous value.

At high gradients, there is some departure from linearity for creep and other diffusive processes. As the gradient approaches a threshold of stability (say 35° or 70 per cent), then material will roll downslope with minimal disturbance, and the rate of sediment transport increases rapidly with increasing gradient. Instead of a linear dependence on gradient, s, sediment transport can be estimated as:

$$\frac{K \cdot s}{(1 - s/s_*)^m}$$

where s_* is the threshold gradient, and the exponent $m = 1 - 2$. Applying the same argument for the slope profile, non-linear diffusion generates a summit convexity which, if there is sufficient relief, straightens out downwards towards a uniform slope at gradient s_*.

In the summit area where diffusive processes are dominant, SLOPE EVOLUTION creates a stable regime in which any minor irregularities in the hillslope surface tend to be obliterated rather than enlarged. Thus the divide area is generally smooth at a scale larger than that of microtopography, and any initial irregularities are progressively removed by the diffusive processes. This broad-scale smooth convex form may sometimes be observed on slopes where there is no soil, particularly on limestones in semi-arid and tropical climates. It may generally be inferred that the convexity was developed under a continuous soil cover, which has since been removed by erosion or washed into joints enlarged by solution.

Seasonal soil creep normally extends to depths of 30–50 cm, and the transport processes act at their full capacity provided that the regolith exceeds this depth. To maintain the rate of soil creep, WEATHERING processes must replace the soil lost by denudation, which is likely to be at about 100 μm yr^{-1} for typical rates and slope forms, increasing with convexity and with the diffusive rate K. Weathering in this context refers to the physical breakdown of rock or *saprolite* (regolith which has been chemically altered *in situ*, retaining the original parent material structure), usually by biogenic or cryogenic processes. Dating of the cosmogenic (see COSMOGENIC DATING) isotope Beryllium-10, from the base of soils in California where bioturbation was by rodents (pocket gophers), Heimsath *et al.* (2001) showed a rate of weathering which decayed exponentially with soil depth, of approximately 280 $\exp(-z/30)$ $\mu m\,yr^{-1}$ at the base of a soil of depth z (cm). To maintain equilibrium between this rate of biogenic weathering and a denudation rate of 100 $\mu m\,yr^{-1}$, the soil should therefore be approximately 31 cm deep. This equilibrium analysis correctly forecasts, for example, that, in comparing the gentle convexities typical of temperate landscapes with the much sharper and narrower convexities of semi-arid badlands, generally thinner regolith is developed on the semi-arid divides.

In summary, soil creep is defined as any slow mass movement. Continuous creep generally occurs in soils which are also subject to rapid mass movements, at shear stresses just below their failure threshold. Seasonal soil creep occurs in a wide range of soils, due to biogenic activity, freeze–thaw and wetting–drying, and the rate of sediment transport is usually assumed to increase linearly with slope gradient. Together with other diffusive processes, it is responsible for developing bulk density profiles and summit convexities. In equilibrium with weathering processes, seasonal creep also determines the depth of the regolith on convexities.

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MIKE KIRKBY

SOIL EROSION

Water and wind can erode, transport and eventually redeposit soils. The initial impact of raindrops can break soil aggregates into primary particles, by the translation of kinetic energy from

Plate 126 A simulated raindrop hitting a moist sandy surface. The kinetic energy of the impacting raindrop causes the formation of an impact crater, around which the water rebounds as a corona. The process occurs very quickly – this photograph was shot at 1:2,000 of a second

the drops to the soil aggregates, a process known as SLAKING (Plate 126). Due to the influence of gravity, more soil particles are splashed downslope than upslope, and detached particles are splashed further downslope. The cumulative effect is a net downslope transfer of soil particles, known as splash erosion.

When rainfall intensity exceeds soil infiltration capacity, runoff occurs. Initially, a sheet of water of fairly uniform thickness flowing over the surface causes quite uniform sheet erosion (see SHEET EROSION, SHEET FLOW, SHEET WASH). However, this state is unstable, as flowing water concentrates in surface depressions and incises into the soil. Where these channels are shallow, the process is RILL erosion. However, if rill erosion continues gullies develop.

The distinction between rill and GULLY erosion is problematic. Early guidelines from the US Soil Conservation Service stated a gully is too wide for a prairie dog to jump across! Later definitions stated a gully would need to be mechanically infilled for agricultural activities to proceed. Others argue that rills are incised into the topsoil (the A horizon), whereas gullies incise into the subsoil or parent material (the B or C horizons).

Eroded soil is eventually redeposited. When this occurs on slopes, the redeposited material is termed COLLUVIUM. It accumulates on concave sections of slopes or against obstructions, such as walls and hedges, as decreasing flow velocities decrease the transport capacity of the runoff water. Sediment can also enter water courses, where it is reworked to form ALLUVIUM. Sediments derived from farmland soils are often rich in agrochemicals, such as phosphate, nitrate and pesticides and can pollute water and damage aquatic ecosystems.

Wind erosion (AEOLIAN PROCESSES) can transport fine sediment considerable distances. For instance, at the Mauna Loa Observatory in Hawaii, the onset of the Chinese spring planting season results in increased dust fallout, even though Hawaii is 5,000 km east of China (Parrington *et al.* 1983). Finer particles, especially silt, are usually transported in suspension above the land surface; sand is usually transported by SALTATION (particles bouncing along the surface).

The severity of soil erosion varies markedly. Langbein and Schumm (1958) proposed a model relating water erosion to rainfall. In very dry climates, there is usually little water erosion, but a slight increase in rainfall often causes a large increase in erosion rates. The occasional rainstorms tend to be very intense convectional storms, which have considerable energy to cause erosion. Also, there is little protection from vegetation, which is sparse in semi-arid environments. In more humid climates, greater vegetation cover protects the soil surface from erosion.

The Langbein and Schumm model has been critically assessed in many parts of the world. Certainly, in the semi-arid tropics, erosion rates are very high. It is estimated that some 6,000 million tonnes of soil per year are washed off the croplands of India and that 1,600 million tonnes per year are transported by the River Ganges to the Bay of Bengal. The Mediterranean environment also has high rates because of the semi-arid climate and long human occupancy of the landscape, which has promoted deforestation (Plate 127).

Plate 127 The effect of a stone protecting soft silty sediments from splash erosion in the Tabernas Badlands of south-east Spain

There are areas where erosion rates should be low, according to the Langbein and Schumm model, but actually are high. This largely reflects human activities, particularly vegetation removal. For instance, destruction of tropical rainforests has greatly increased erosion rates. Rainforest soils contain little organic matter, unlike those of more temperate environments, and removal of vegetation can lead to further losses. Organic matter stabilizes soil structure, so its loss increases erosion risk.

Temperate continental interiors (Prairies of North America, Steppes of Central Asia, Pampas of Argentina) should have a natural vegetation cover of grassland. Under these conditions, the soils are thick, black organic Chernozems. Over the past 100–130 years many of these grasslands have been converted to arable use, particularly for cereal production. The soil organic content has decreased and the soils have become more ERODIBLE. Some 4,000 million tonnes of soil per year have been eroded from the continental USA since the 1930s. This would fill a freight train long enough to encircle the equator twenty-four times!

There is increasing evidence of soil erosion in northern and western Europe. In the Langbein and Schumm model, these areas should experience little erosion. However, the damage to soil structure imposed by increasingly mechanized and intensive agriculture, particularly since the 1940s, is believed to have increased erosion rates. Erosion directly related to cultivation is often termed tillage erosion.

Soil erosion can be described as 'a quiet crisis, one that is not widely perceived. Unlike earthquakes, volcanic eruptions or other natural disasters, this human-made disaster is unfolding gradually' (Brown 1984). Brown estimated that the maximum rate of soil formation is 2 tonnes of soil per hectare of land per year, but that the average soil erosion rate is about 20 $\mathrm{t\,ha^{-1}\,yr^{-1}}$. The calculated global excess of soil erosion over formation is 25,730 million tonnes, roughly equivalent to the amount of topsoil in Australia's wheatlands. These estimates are very tentative, but they indicate the scale of the problem.

It is difficult to define what is 'acceptable' as a rate of soil erosion. 'Natural' or 'geologic' erosion is a natural process by which uplands are denuded over geological time. This is different from 'accelerated' erosion, where human activities are increasing erosion rates much above their background levels. The commonest activity is vegetation removal, exposing soils to erosion.

To define 'acceptable' soil erosion, many soil scientists suggest that the soil system should be in a state of dynamic equilibrium; that is, the rate of loss by erosion should not exceed the rate of soil formation. We know very little about rates of soil formation, except that it is a slow process. Soils form by the WEATHERING of material at the soil–parent material interface, or by deposition of sediment on the surface. The parent material is usually rock, but often includes soft sediments. It normally takes 1,000 years to weather 10 cm of material at the B/C horizon interface. By scraping the bedrock and increasing soil aeration and microbial activity, tillage can accelerate the process up to a maximum of about 10 cm in 100 years (i.e. $1\,\mathrm{mm\,yr^{-1}}$). If erosion does not exceed this value, the soil system is in a state of dynamic equilibrium. This value has been labelled a 'tolerable' or 'T value' and is used in planning SOIL CONSERVATION strategies.

The $2\,\mathrm{t\,ha^{-1}\,yr^{-1}}$ 'T value' may be too high, as it assumes an unlimited supply of weatherable material and ignores the complexity of soil-forming processes. Soils mature through time, usually incorporating organic matter into the topsoil. Erosion preferentially removes topsoil, often with serious implications for soil fertility. The topsoil is the 'seat' of most biological activity and contains most of the soil fauna, organic matter and nutrients, both natural and applied. Therefore, erosion involves more than the loss of physical components of the soil system. Often, the most fertile material is lost and any new soil formed at the base of the profile is much less fertile. The topsoil may be completely stripped away, deposited downslope and then less fertile subsoil or parent material deposited on top. This 'soil profile inversion' diminishes soil fertility. All criticisms suggest the $2\,\mathrm{t\,ha^{-1}\,yr^{-1}}$ 'T' value is too high. However, soil erosion rates can exceed it by orders of magnitude and, in extreme cases, completely remove the soil. In the long term, even low erosion rates can be damaging.

Numerous techniques exist for measuring soil losses. For water erosion a simple technique is to establish runoff plots. These are bounded on three sides (usually by wood, metal or plastic) and runoff and eroded sediment are collected at the downslope end. This approach was established in Germany by Ewald Wollny in the 1880s. Runoff and erosion rates are measured in precisely defined conditions of soil type, slope and vegetation cover. However, it has been criticized, mainly because plot boundaries interfere with erosion processes, for example impeding rill development. Runoff plots were nevertheless adopted by the US Soil Conservation Service, following its establishment in 1934. Standardized 0.01 acre plots were constructed in a range of agricultural environments, principally east of the Rockies and led to the development of the 'UNIVERSAL SOIL LOSS EQUATION' (USLE).

Since the USLE was introduced, many soil erosion equations and models have been developed, including the revised USLE (RUSLE) and the Water Erosion Predictive Equation (WEPP). The 'Erosion Productivity Impact Calculator' (EPIC) attempts to predict the long-term effects of erosion on soil properties and crop productivity, simulating erosion for up to hundreds of years. EUROSEM is an attempt to predict erosion rates in European conditions and LISEM is a model to predict erosion on LOESS soils (Plate 128).

Models are also used extensively in wind erosion research. The most notable example is the Bagnold Equation, first published in 1937 by Major R.A. Bagnold. During military service in

Plate 128 High sediment concentrations in the Yellow River of China. It is estimated that the river transports some 1,800 million tonnes of sediment per year to the Yellow Sea. Much of this sediment is entrained when the river flows through the Loess Plateau, an area of erodible soils derived from wind-blown silt

the British army in Libya, Bagnold studied the dynamics of sand dunes. In the equation, sediment transport is related to wind velocity:

$$Q = 1.5 \times 10^{-9}\ (V - Vt)^3$$

where Q = total sediment load ($t\,m^{-1}h^{-1}$), V = velocity at measuring height ($m\,s^{-1}$), and Vt = fluid threshold velocity for sand movement ($m\,s^{-1}$).

Vt is the wind velocity necessary to initiate effective sand transport. Once that velocity is exceeded, sediment transport increases as the cubic power of wind velocity, so slight increases in wind velocity above Vt cause marked increases in wind erosivity. The Bagnold Equation has formed the basis of other predictive equations, especially in North America. Morgan (1995) comprehensively reviewed erosion models.

There are many techniques to monitor erosion processes in the field and laboratory (De Ploey and Gabriels 1980). A simple technique is to insert erosion pins vertically into the soil, allowing accurate measurements of changing surface levels to show how much erosion (surface lowering) or accumulation (surface raising) has occurred. Simple traps, often referred to as Gerlach 'troughs', can be used in the field to collect sediment. The dynamics of sediment movement can be studied by 'tagging' soil particles with a tracer and using tracer movement to indicate soil movement. Tracers have included painted stones and soil particles, fluorescent dyes, radioactive isotopes and magnetic materials. The fallout of isotopes from nuclear explosions or accidents, such as Chernobyl, has been used to quantify erosion rates. The most widely used isotopes are Cs^{137} and Cs^{134}. These are positively charged and, as they fall to Earth, become attached to negatively charged clays and organic particles. As a site is eroded, it becomes depleted in the isotope so, comparing the radioisotope content of eroded soils with non-eroded soils, such as in a woodland, indicates approximate erosion rates. Another method is to investigate the distribution of fallout isotopes in colluvial deposits. Loughran (1989) summarized field methods to quantify erosion rates. Geomorphologists have also simulated soil erosion processes in the laboratory. However, there are difficulties with this approach, such as how realistically laboratory studies simulate field conditions.

Soil erosion is the product of many complex and interacting factors. For instance, in a laboratory study to assess erosion risk on fifty-five soil samples from the US Cornbelt, Wischmeier and Mannering (1969) found twenty-two soil and surface properties were necessary to explain 95 per cent of soil loss variance. However, Morgan (1995) offered a useful qualitative simplification, stating that soil erosion results from the dynamic interaction of:

1. The energy of the water or wind in causing erosion (EROSIVITY),
2. The inherent resistance of soil to detachment and transport (ERODIBILITY),
3. The protection factor of vegetation.

Many studies have attempted to relate rainfall erosivity to erosion rates. Rainfall erosivity is a function of its intensity and duration and the mass, diameter and velocity of raindrops. The energy of water flowing over the soil surface also affects erosion and is related to its velocity, volume, turbulence and shear stress (Plate 129). Slope angle, length and shape profoundly affect the erosivity of running water. Generally, as slope angle increases, so does erosion risk, because runoff velocity and energy increase. Usually, erodible soils on slopes >10° are particularly susceptible to rill erosion. Rills are hydraulically efficient systems for transporting soil and promote high erosion rates. Increased slope angle also increases the efficacy of splash erosion processes. On longer slopes, runoff has more time to accelerate and thus achieve greater erosivity. Slope

Plate 129 Tu Lin (the 'Soil Forest') in Yunnan Province, China. Highly erosive summer monsoonal rains have incised into soft Tertiary sediments. Human agency has also played a role in promoting erosion. For instance, in the top-centre note the cultivation of melons right at the edge of the gullied area

shape is also important, as runoff tends to accelerate rapidly over convex slope segments, achieving higher velocities and thus greater erosivity, but decelerates over concave slope sections, often leading to sediment deposition. Thus, the typical convex–concave morphology of slopes in humid temperate environments makes them particularly vulnerable to water erosion.

Soil erodibility is influenced by many factors, principally texture and soil organic content. The most erodible materials are silts and sands; which are not cohesive and can be transported by the flow rates characteristic of rills. Soils with low organic matter contents are highly erodible, particularly those with <2 per cent organic matter by weight. As organic matter increases above 2 per cent, soil erodibility decreases to a minimum at about 10 per cent organic matter, a typical value for a deciduous forest topsoil in a humid temperate environment, such as the British Isles. Soils with >20 per cent organic matter tend to be more erodible, as there is less clay to form aggregates with organic matter, and organic material is very light and easily transported. Organic particles have densities typically about $0.8\,\mathrm{g\,cm^{-3}}$ compared with $>2.5\,\mathrm{g\,cm^{-3}}$ for most mineral particles. Low density is the main reason why very organic soils, such as the peaty soils in the Fens of East Anglia, are susceptible to wind erosion.

Vegetation protects the soil from erosive forces by reducing, braking and filtering runoff. Many studies agree that >30 per cent plant cover protects the soil from erosion. Vegetation surfaces dissipate raindrop energy and prevent slaking. This is particularly true for short vegetation. A fall of about 8–9 m is required to achieve terminal velocity, so raindrops falling from tree canopies can regain their erosivity.

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MICHAEL A. FULLEN AND JOHN A. CATT

SOIL GEOMORPHOLOGY

Soil geomorphology is the scientific study of the processes of landscape evolution and the influence that these processes have on soil formation and distribution on the landscape. Soil geomorphology provides a unique framework for an integrated Earth surface assessment. This discipline couples knowledge from soils, surficial deposits, stratigraphy and sedimentation, and parent material with the process-oriented approach of geomorphology in a three-dimensional framework. The soil geomorphology discipline in the United States has provided the primary foundation for interpreting the relation of soils and palaeosols to the landscape.

Soil geomorphology, an interdisciplinary construct, merges two scientific fields: pedology, the study of soils, and geomorphology, the study of landforms and surficial processes. Robert V. Ruhe was one of the first to both quantify landscape form and process in space and time and to integrate these concepts with soil science (e.g. Ruhe 1956, 1960, 1969). Olson (1989, 1997) describes the event chronology leading to the emergence of soil geomorphology as an independent discipline.

Soil geomorphology provides the framework for understanding the geomorphic history of a landscape. A soil-geomorphic approach requires three components: (1) knowledge of the surficial stratigraphy and parent material, (2) geomorphic surfaces defined in time and space and (3) correlation of soil patterns and properties to landscape features. These three components are assessed independently and their results subsequently integrated to produce a soil-geomorphic interpretation or soil-geomorphic model of the landscape and the processes leading to its evolution. A soil-geomorphic model can represent the landscape at defined scales from the hillslope to the continental.

Soil landscape models

A common ground is required to study and understand landscape evolution and the processes that shape the land surface and its soils. The foundation for our current understanding of soils and landscapes lies in the historic concepts of William Morris Davis and Walther Penck. The 'CYCLE OF EROSION' a time-dependent landscape evolution model (Davis 1899), describes landscape progression by downwearing through stages from youth through maturity to old age. In developing this model, Davis was heavily influenced by early theories of evolutionary biology and the concepts of contemporaries including John Wesley Powell and G.K. Gilbert. Slope reduction occurred by uniform downwearing in which geologic differences became insignificant with time as landscapes advanced through the cycle. Process was not a part of this model, a serious flaw. In contrast to the Davisian model, Penck (1924) emphasized backwearing and parallel slope retreat. Penck's model meshes well with our understanding of soil distribution on the landscape and provides a better foundation for soil geomorphology.

Although the Davis and Penck concepts form the basis for general landscape evolution models, they did not emphasize hillslope development in relation to soils. Milne (1936a, b) was one of the first to introduce the CATENA concept to illustrate soil patterns on a hillslope. Here, soil properties depend on topography and are repeated relative to each other from the hillslope summit to the adjacent valley floor. Milne's catena model recognized that the processes of erosion and deposition on the various hillslope positions directly affect the distribution of soil properties. Two variations were recognized: (1) all soils of a catena are formed in a single parent material and (2) soils of a catena are formed in two or more materials. Soils of a catena differ in case (1) because of 'drainage conditions, differential transport and deposition of eroded material and leaching, and translocation and redeposition of mobile chemical constituents' (Milne 1936a). Milne's statement implies that pedogenic processes together with hydrologic properties define the soil landscape and are not restricted to a point on the landscape. In the second case (2), a geologic factor for multiple-parent materials is included. Milne's catena model has evolved into a more limiting model known as a toposequence. Numerous studies have continued to follow Milne's catena model in evaluating soil landscape relations today. The importance of the catena concept to soil science was recognized quickly in the USA. In the USA Soil Survey today, the toposequence is in practice a hydrosequence, i.e. a series of soil profile colours used as indicators of water-table elevation changes along a hillslope.

Wood (1942) and King (1953) created 'the fully developed hillslope' model. Ruhe (1956, 1960, 1975) formulated a soil-geomorphic hillslope model that integrated soil properties with hillslope models and modified the Wood and King models by proposing hillslope elements: summit, shoulder, backslope, footslope and toeslope. These elements are now widely used to describe hillslope positions. Conacher and Dalrymple (1977) extended the two-dimensional hillslope approach to a drainage basin. This nine-unit model, based on form and geomorphic and pedogenic processes, attempted to integrate the components of a hillslope by considering material and water movement.

Landscape morphology

To fully evaluate landscapes and their relations to soils, an understanding of landscape morphology is important. Minimum parameters are gradient, aspect, and vertical and horizontal curvature. Ruhe (1975) described slope curvature using three components: slope gradient, slope length and slope width. A matrix of nine basic forms was used to represent changes in curvature. Huggett (1975) added surface flow lines to the basic slope shapes and Pennock and others (1987) combined hillslope elements and curvature to identify seven hillslope positions.

Soil landscapes and water movement

Water movement is one of the most important mechanisms in landscape evolution and soil development. Water movement is governed by a complex set of interrelated factors both internal (e.g. soil properties) and external (e.g. climate) to the soil-landscape system. Some of the geomorphic models above included water movement, with an emphasis on overland flow or vertical infiltration at a given hillslope position rather than flow through the landscape. This early emphasis reflected the influence of Horton's studies on OVERLAND FLOW in the 1930s. However, in addition to surface runoff and infiltration, throughflow and groundwater recharge are equally important paths for water flow on the landscape. In recent years, the importance of subsurface water movement and transport through the soil landscape has received more attention.

Subsurface-flow landscape models including flownet analysis have become important particularly in wetlands studies.

Soils and geomorphic surfaces

Soils and geomorphic surfaces are closely aligned. A geomorphic surface is that part of the landscape specifically defined in space and time with definite geographic boundaries (Ruhe 1956, 1969; Daniels *et al*. 1971). When a soil-geomorphic study is undertaken, the geomorphic surfaces are delineated based on geomorphic principles. Separately, soils in the study area are mapped. The results are compared and the linkages established. Soil boundaries do not necessarily coincide with geomorphic surface delineations and one or more soils may occur on the same geomorphic surface (Ruhe 1975). The critical point to the interpretations is that the pattern of soils and surfaces should be predictably repeated throughout a drainage basin and represents the landscape as it is at a given time. If three different soils occur on a geomorphic surface on one side slope, this soil complex should recur on similarly situated side slopes throughout the watershed. Geomorphic surfaces represent a time sequence but environment and other factors vary independently of one another through time. The latter phenomena affect the types of soils found on any given geomorphic surface. However, the soils will have a common degree of soil development. This relationship provides a valuable tool for understanding the chronological framework of landscape evolution in a soil geomorphic investigation.

Palaeosols and soil geomorphology

Palaeosols are soils formed on landscapes of the past. Ruhe (1975) defined three types: buried, exhumed and relict soils. Buried palaeosols are soils later covered by younger sediment or rock. Exhumed palaeosols are those buried but later re-exposed on the land surface, and relict palaeosols are those formed on a pre-existing landscape but never buried. As researchers began to demonstrate the close interdependence of soils and landscapes (e.g. Ruhe *et al*. 1967; Ruhe 1969), palaeosols also became indicators for understanding past landscapes. Often, buried soils are known as stratigraphic markers and are especially useful as keys to past environments (e.g. Follmer 1982).

Quantifying soil landscape models

Landscape models have shown that landscapes are predictable and have a large non-random variability component. This non-random component is useful in predicting soils on the landscape if the processes that govern landscape development are quantifiably described. Many new and exciting approaches seeking to capture and quantify soil landscape relations encompass geostatistical techniques, digital terrain models, fuzzy logic, neural networks and expert systems and produce visualizations and interactive, virtual immersive modules previously unavailable (e.g. <http://grunwald.ifas.ufl.edu> and <www.soils.wisc.edu/virtual_museum>).

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SEE ALSO: hillslope, form; hillslope, process; Horton's Laws; hydrological geomorphology; palaeosol

CAROLYN G. OLSON

SOLIFLUCTION

Slow downslope movement of soil mass usually associated with FREEZE–THAW CYCLEs and FROST HEAVE, first termed to refer to the movement of saturated unfrozen soil in the Falkland Islands (Andersson 1906). MASS MOVEMENTs of soils in periglacial regions involve localized rapid soil failures, which result in active-layer detachment slides or flows, and more widely operating slow pre-failure movements (Ballantyne and Harris 1994: 114). The latter is referred to as solifluction as a collective term. In a broad sense, solifluction consists of frost creep (SOIL CREEP associated with freezing) resulting from nearly vertical settlement of soils heaved normal to the slope and gelifluction representing downslope displacement of ice-rich soils during thawing (Washburn 1979: 201). The two components often operate together, displacing soils downslope generally at rates of 0.5–10 cm a^{-1}. Where the ground is underlain by PERMAFROST, solifluction occurs within the ACTIVE LAYER and is distinguished from permafrost creep that originates from plastic deformation of frozen debris. The long-term operation of solifluction results in low-relief, gentle slopes having a number of lobes and sheets.

Processes

Frost creep is subdivided into needle-ice creep, diurnal frost creep and annual frost creep in terms of the time span of operation and the vertical extent of movement. Annual frost creep is often accompanied by gelifluction. In cold permafrost regions, a large part of annual frost creep and/or gelifluction may occur near the base of the active layer, producing a plug-like velocity profile.

Diurnal freeze–thaw cycles lead to either needle-ice creep or diurnal frost creep (Figure 158). The former occurs when nocturnal cooling to just below 0 °C is followed by daytime thawing. Only the uppermost few grains are heaved by ice needles and resettled on thawing by toppling or rotational movement. The resulting downslope movement is superficial but relatively rapid, accelerating approximately with the second power of the slope gradient.

More intensive nocturnal freezing can produce ice lenses at a depth of a few centimetres, which heave the overlying soil layer normal to the slope. On thawing, the heaved soil resettles vertically or, in reality, with some upslope component due to cohesion. Such a freeze–thaw alternation induces diurnal frost creep, the movement of which is proportional to the slope gradient and shallower than 10 cm. The velocity profile of diurnal frost creep is typically concave downslope, in response to the increasing number of freeze–thaw cycles

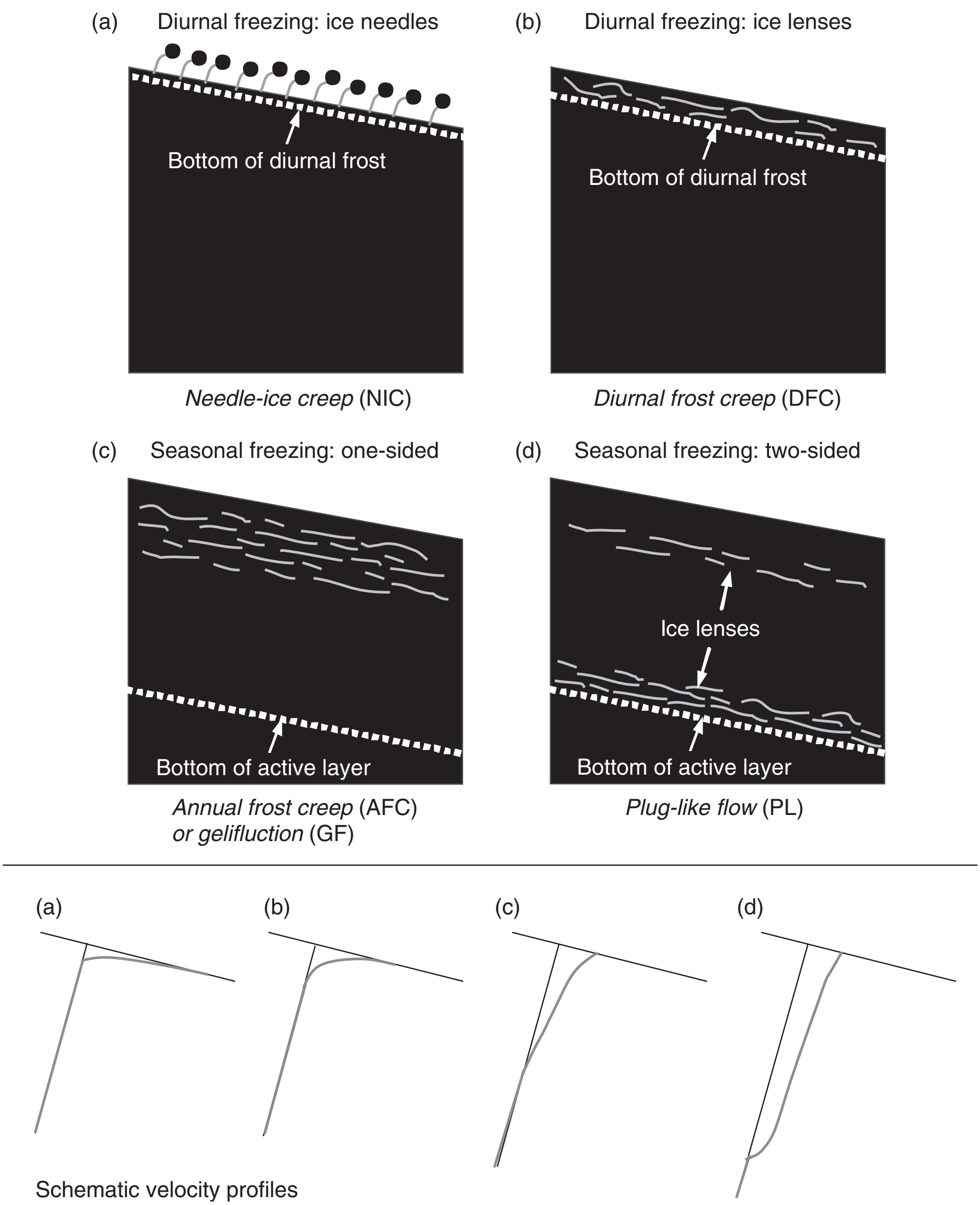

Figure 158 Types of solifluction
Source: Matsuoka (2001)

towards the ground surface. Where needle-ice creep dominates, a gap in velocity occurs between the uppermost grains and the underlying soil (Matsuoka 2001).

Annual freeze–thaw cycles cause frost heave of soils either by one-sided freezing or by two-sided freezing. In regions subjected to seasonal frost or underlain by warm permafrost, one-sided freezing from the ground surface in winter often produces ice lenses within the uppermost few decimetres of soil. During seasonal thawing, the resettlement of the heaved soil results in annual frost creep and/or gelifluction. Reflecting the depth of the ice lenses, the base of movement typically lies at 30–50 cm depth. The relative contribution of gelifluction is estimated by subtracting the potential frost creep, which is the product of the amount of frost heave and the tangent of the slope gradient, from the total downslope displacement. The gelifluction component increases with the silt plus clay content, because excess pore water during thawing reduces frictional strength in the soil (Harris 1996).

Where the permafrost temperature is low, winter freeze-back of the active layer may progress both downward from the top and upward from the base. Such two-sided freezing may produce ice lenses both near the surface and near the permafrost table. The high moisture availability at the latter location favours the formation of numerous thick ice lenses. Thawing of the basal ice lenses induces plug-like flow that shows a velocity profile convex downslope near the base of the active layer (Mackay 1981).

Rates

Field measurements in a wide range of cold regions indicate that rates of solifluction, expressed by either the surface velocity or the volumetric velocity, significantly vary with climate (Matsuoka 2001). In the polar, cold permafrost regions, where the mean annual air temperature (MAAT) is −6 °C or lower, the paucity of diurnal frost creep restrains the surface velocity to below 5 cm a^{-1}, while plug-like flow allows movement of the soil mass deeper than 50 cm. In the alpine, shallow seasonal frost regions, the predominance of diurnal frost creep (including needle-ice creep) raises the surface velocity up to 100 cm a^{-1}, whereas the soil movement is confined largely within the uppermost decimetre; as a result, the volumetric velocity is very low despite the high surface velocity. The volumetric velocity reaches a maximum in regions with warm permafrost or deep seasonal frost (MAAT between −6 and 0 °C), which are located in both subpolar and alpine settings, because diurnal and annual processes combine to dislocate a 30–50-cm thick soil mass with moderate surface velocities.

The slope gradient is another significant control on solifluction rates. Increasing rates with inclination have been reported from a number of polar slopes. In lower latitude alpine areas, however, other factors like soil frost susceptibility, moisture distribution and freeze–thaw frequency obscure the overall dependence of solifluction rates on inclination (Harris 1981: 123–125). Gelifluction can occur on gradients as low as 1°.

Radiocarbon ages of organic materials buried by solifluction deposits show that long-term variation in solifluction rates generally corresponds to climate change during the Holocene, although the responsible climatic factors are not unequivocal (Matthews *et al.* 1993).

Landforms

Solifluction produces characteristic surface features involving lobes, terraces and sheets. Where vegetation is sparse or absent, sorted stripes (a kind of PATTERNED GROUND) often develop on the tread of these features. The most widespread features are lobes 2–50 m in both width and length. Lobes typically occur in alpine regions where heterogeneous surface conditions localize soil movement, whereas sheets, which lack lateral margins, dominate polar slopes that experience more uniform movement (Plate 130). The downslope edge of these features terminates in a steep riser 0.2–2 m in height, where vegetation, coarse debris or downslope declination acts as a brake on soil movement. As a result, the riser is commonly turf- or stone-banked (Benedict 1970). Terraces may develop under the interaction between wind, vegetation and soil movement (Ballantyne and Harris 1994: 261–267). Isolated boulders on slopes subject to solifluction often move as PLOUGHING BLOCKS AND BOULDERS.

The height of lobes reflects the maximum depth of soil movement. The predominance of shallow diurnal frost creep results in small lobes about 0.2 m in height. Such lobes occur mainly

Plate 130 Turf-banked solifluction lobes on a limestone slope, Swiss National Park (2,400 m ASL)

on alpine near-crest locations where thin REGOLITH can only respond to diurnal freeze–thaw action (Matsuoka 2001) or on tropical high mountain slopes where the lack of seasonal variation in temperature highlights the effect of diurnal freeze–thaw action (Bertran *et al.* 1995). Larger and higher lobes develop where solifluction originates mainly from annual freeze–thaw action. The length of lobes depends partly on the time of process operation. Well-developed lobes require several thousand years of activity (Matthews *et al.* 1993).

Numerical simulations suggest that long-term predominance of solifluction, accompanied by subsurface debris production and comminution due mainly to frost weathering, eventually leads to gentle convex-upward slopes, often referred to as CRYOPLANATION surfaces, near mountain crests with possible occurrence of summit TORS (Anderson 2002).

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SEE ALSO: freeze–thaw cycle; frost heave; periglacial geomorphology; mass movement

NORIKAZU MATSUOKA

SOLUBILITY

Minerals, and the chemical elements which constitute them, vary in the degree to which they dissolve in water. During the dissolution process, the constituents of the chemical compounds in minerals split up, or dissociate, into water. It follows that the solubility of such compounds can be assessed by measuring the concentrations of the constituent elements in the water after a period of time. Minerals generally display increased solubility with higher temperatures and with increasing acidity, notable exceptions to the latter being silica (as quartz) which increases in solubility above pH 10 and aluminium (as gibbsite or kaolinite) which is least soluble at around pH 6 but increases both as pH rises and falls from this value. The solubility of gypsum (as measured by Ca in solution) does not vary with pH. For gypsum, the rate of water flow is a governing factor.

Maximum concentrations are reached asymptotically over time but nevertheless some practical measures of solubility can be found. The Nernst equation describes the dissolution trend over time:

$$C = C_{max}\,(1 - e^{-kt})$$

where C is concentration at time t, C_{max} = maximum concentration and k is a rate constant for each solute.

Highly soluble compounds have a high value of k and exhibit steep rise in concentration over time. The process of dissolution involves the redistribution of energy – and for dissolution to occur there must be a decrease in free energy – and if a solute and a solvent are composed of similar molecules in terms of structure and electrical properties, then high solubility is favoured (see further Davidson 1978: 83–85).

Water flow rate is also important in determining the solubility of minerals in the field and there are also many other factors which limit the transfer of *in vitro* observations of solubility to the field (Casey *et al.* 1995). For water flow, the rate of dissolution of the more rapidly dissolving minerals becomes transport-limited as the products of dissolution readily accumulate and further dissolution is then only facilitated by the removal of weathering products and the arrival of further chemically unsaturated water. For slowly dissolving minerals, the rate of dissolution tends to be slow as compared to the flow of water and so the overall solutional losses tend to become rate-limited. Thus water flow rates, and, for bare surfaces, duration of rainfall and thus time of wetting, become important controls on solution rates.

Additionally, laboratory measurements of solubility are usually made under controlled conditions, whereas field conditions tend to be variable and possess different combinations of factors which are in practice difficult to separate, like freezing, thawing, salt and biological weathering as well as dissolution (Trudgill and Viles 1998).

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STEVE TRUDGILL

SOLUTE LOAD AND RATING CURVE

All natural waters contain organic and inorganic material in solution; these are called solutes. Spatial and temporal variation in solute content has been studied as a means of investigating CHEMICAL WEATHERING processes, rates of chemical denudation, evaluating nutrient cycling and elucidating the variable pathways followed by water through the drainage basin. The major solutes of interest are calcium, magnesium, sodium, potassium, chloride, bicarbonates, sulphate, nitrate and silica.

According to the classification of Gibbs (1970), the major natural mechanisms controlling world surface water chemistry are: (a) atmospheric precipitation, both composition and amount; (b) rock weathering; and (c) evaporation and fractional crystallization. When he plotted total dissolved solids (t.d.s.) in milligrams per litre against the weight ratio of sodium to sodium plus calcium for the major rivers of the world, he showed three domains: (1) high t.d.s. and high Na to Na + Ca ratio, where the evaporation/crystallization processes are dominant; (2) average t.d.s. and low Na to Na + Ca ratio; and (3) low t.d.s. and high Na to Na + Ca ratio. From this classification, climate and hydrology are the dominant controls of solute concentration at high and low t.d.s. values and lithology becomes dominant at average t.d.s. values. Meybeck (1988) has provided a more detailed breakdown of river water solutes according to climate, hydrology and relief, on the assumption that the residence time of water in a basin will depend strongly on relief.

Meybeck (1987) estimated the global average chemical denudation rate for crystalline igneous and metamorphic rocks and sandstones and shales at 18–19 tonnes/km^2/year and volcanic rocks at 1.5 times higher. Denudation rate for carbonate rocks is 100 and for evaporites is 423 tonnes/km^2/year. This is in the context of a world average of 42.

The effect of relief on chemical denudation rates varies with geology as carbonate rocks dissolve easily regardless of local topography. No effect of hillslope steepness or extent of recent glaciation has been detected on chemical weathering fluxes in small granitoid watersheds. This implies that physical erosion rates are not critical in influencing chemical weathering of silicates and, by further implication, geology and climate are probably more important than relief on a global basis.

There are further factors that are important. Of these, the anthropogenic and the biotic factors are the most crucial. Vegetation increases chemical weathering by supplying carbon dioxide and organic acids to the soil and it increases water contact time with minerals in the soil by retaining moisture and by locally accelerating water recycling via evapotranspiration-enhanced rainfall. Vegetation also affects weathering by stabilizing soil against erosion and thereby increasing weathering rates in regions of high physical erosion (Berner and Berner 1996). Nutrients, organic carbon and dissolved trace elements will be most affected by anthropogenic and biotic controls, and will, at times, vary quite independently from the topographic, geologic, climatic and hydrologic factors.

The most geomorphically relevant ways of analysing solute data depend on the scale of interest. At individual site and slope scale, diffusive and convective equations have been used to model variations in solute flux (e.g. Carson and Kirkby 1972; Berner 1978), but it is probable that the buried tablet technique (Trudgill 1977) will provide the most reliable relative weathering rate information. At river basin scale, input–output budgets and solute budgets, combined with variable runoff source analysis, are the most popular (e.g. Zeman and Slaymaker 1978; Laudon and Slaymaker 1997). At global scale, Meybeck (1982, 1988) has made major contributions through the use of solute budgets.

Precipitation inputs to the land surface contain solutes in dry and wet fallout. The magnitude and composition of this solute content varies with distance from the ocean. As water moves through the vegetation canopy, the soil and the rock of the drainage basin, the solute concentration and composition will change. Solute concentrations within the soil are influenced by precipitation, interactions with the soil matrix, release of solutes through chemical weathering and biotic uptake and release of nutrients.

The solute content of stream flow will therefore reflect the characteristics of the upstream basin, including its geology, topography, vegetation cover and the variable pathways and residence time associated with water movement through the basin. Concentrations will vary with time in response to hydrologic conditions and will frequently exhibit a dilution effect during storm runoff events.

The precise hydrologic pathway has important implications for residence time and hence on solute enrichment. Identification of the pathway gives an understanding of the magnitude and timing of solute fluxes from different hydrologic reservoirs in the landscape and is therefore essential for the understanding of variations of the stream water chemistry. Hydrograph separation of rain-driven storm flow into its storm and pre-storm components can be carried out by collecting stream water samples for stable isotope analysis before, during and after a storm runoff event. Four assumptions are made: (1) rain isotopic content can be characterized by a single isotopic value; (2) pre-storm component, ground water and vadose water can be characterized by the base flow with a single isotopic value; (3) isotopic content of precipitation will be significantly different from pre-storm runoff values; and (4) contribution of stored surface water to the stream is negligible. Under certain hydrologic conditions, alternative hydrologic tracers (such as silica and electrical conductivity) can be used. The specific advantage of electrical conductivity is that it can be continuously monitored and stored in dataloggers.

Measurements of the solute input into a drainage basin and the output in streamflow provide a means of establishing a solute budget for the basin. The net solute yield (output–input) reflects the production of solutes within the basin. This production may be related to chemical weathering, the uptake of carbon dioxide by weathering reactions and the mineralization of organic material. On a global basis, approximately 50 per cent of the solutes found in river water are the products of chemical weathering, but this value is highly variable depending on lithology.

Rating curves are used to describe relations between solute transport and water discharge. If the rating curve is stable, the water discharge can then be used to predict solute concentration and load. The characteristics of these plots, including slope, degree of scatter and intercept are frequently used to characterize the solute response of a drainage basin.

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SEE ALSO: chemical denudation; denudation

OLAV SLAYMAKER

SPALLING

Spalling is the peeling off of platy fragments from the surface of rock. The resulting 'spalls' vary from a few centimetres to several metres in scale but their thickness is usually from 1–5 cm. The under surface of spalled fragments may be very irregular. Spalling can be transitional with EXFOLIATION or SHEETING though the latter usually occurs on a much larger scale. The term 'flaking' is also sometimes used.

Spalling can be attributed to several processes. Fundamentally, differential stresses in the outer layer of rock cause separation. The source of this differential stress may be from the growth of salt or ice crystals. Chemical change may be a source of differential stress as secondary minerals are precipitated. These occupy a greater volume and therefore may exert an outwards force on the rock surface. Expansion and contraction due to thermal change (e.g. forest fires or insolation) may also produce sufficient differential stress (Gray 1965). Change of internal stress equilibrium (e.g. due to erosion) may cause separation of plates from an intact rock mass. Spalling may be accompanied by surface alteration or by CASE HARDENING.

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DAWN T. NICHOLSON

SPELEOTHEM

'Speleothem' (Greek: *spele* – cave, *them* – make) – a general term for minerals precipitated in CAVEs. More than 250 different minerals are known (Hill and Forti 1997). Calcite deposited in limestone caves is overwhelmingly predominant, accumulating in both air-filled (vadose) and water-filled (phreatic) conditions. 'Travertine' and 'sinter' are alternative terms, or 'tufa' (see TUFA AND TRAVERTINE) when precipitated on organic frameworks. Aragonite is second in abundance. Gypsum is third, found in gypsum caves and also in limestone caves with gypsum interbeds or where H_2SO_4 can react with the rock. Hydrated carbonates and sulphates (e.g. hydromagnesite, epsomite) are quite common, usually as pastes or powders in small amounts. Other minerals are more localized, associated with particular source conditions in bedrocks or clastic fillings; they include native sulphur, many oxides and hydroxides, halides, nitrates, phosphates, silicates, vanadates and a few organic minerals. There is perennial ice in many cold caves.

$CaCO_3$ (calcite and aragonite) speleothems

Calcite occurs mostly as coarse crystals with c axes oriented to growth ('length fast') or in microcrystalline form with c axes oriented across the direction of growth ('length slow'; see Railsback 2000). Many vadose speleothems display alternations of coarser and finer crystals, often with temporary cessations due to drying or dissolution, zones with dust, mud or organic grains: existing crystals may grow through these or new crystals form upon them. Subaqueous deposits are more homogeneous. Pure calcite is translucent, or opaque white due to fluid inclusions. Colour banding (yellow, brown, red-brown) is common, due chiefly to incorporation of fulvic acid chromophores from soil waters (van Beynen *et al*. 2001). Metals such as iron (red), copper (blue) and nickel (bright green) also provide colour, but rarely in visible concentrations (Cabrol and Mangin 2000).

Aragonite generally occurs as needle-like clusters or massive stalagmitic aggregates. Its deposition instead of calcite is attributed to enrichment of Mg^{2+} ions in the feedwater, due to presence of dolomite ($CaMg{\cdot}2CO_3$) or to evaporative effects. Aragonite speleothems are more common in warmer climates, e.g. wet–dry seasonal alternations of calcite and aragonite are reported in Botswana.

Vadose speleothem shapes are created by gravity, or by growth and capillary forces. Principal gravity types are dripstones (stalactites, stalagmites), and flowstone sheets on floors and/or walls. A 'column' is a stalactite–stalagmite pair grown together.

The fundamental form is the 'straw' stalactite, a monolayer crystal sheath enclosing a feedwater canal and growing downwards only. Leakage from the canal may overplate the sheath, creating tapered (carrot-like) stalactites up to one metre in diameter and several in length. Accelerated deposition on protruberances can add a myriad of subsidiary forms such as crenulations, corbels, drapes and lesser stalactites. 'Curtains' grow downwards where feedwater trickles down a sloping ceiling.

The most simple stalagmite is a 'candlestick' adding all new growth at the top under a nearly constant drip. Varying drip or greater fall height causes terraced or corbelled thickening; at the extreme the form is like a pile of soup plates. More common are conical or tapered forms, broadening into domes with flowstone sheets around them. Some stalagmites are >30 m high and domes may be 50 m or more in diameter.

Flowstones are deposited from film flow and accrete roughly parallel to the host surface. They may extend tens to hundreds of metres downstream of their sources and accumulate to thicknesses of several metres. 'Gours' or 'rimstones' are dams building upwards from irregularities in stream channels or on flowstone surfaces. The greatest impound water to depths of several metres. Rims are often strikingly crenulated.

'Helictites' or 'excentrics' grow where crystal or capillary forces predominate, skewing c-axes to create narrow, curvilinear tubes extending out, up and down from rock or parent stalactites, etc. Most are short, <10–20 cm in length. Dense clustering can form tangled masses like the Medusa's hair. 'Anemolites' grow upwind into prevailing drafts. Clusters of needles fanning outwards ('frostwork') are the principal aragonite excentrics.

'Cave pearls' are spheroidal accretions about a nucleus such as grit agitated by water dripping into a pool. 'Popcorn' ('cave coral') describes semi-spherical accretions on flowstone or other surfaces, often in dense, multilayered clusters.

Subaqueous calcite may precipitate from thermal or meteoric waters. The principal forms are spar linings, e.g. most of the 150+ km of passages in Jewel Cave, South Dakota. Deposition extends from the water table to a limiting depth determined by pressure and Ca^{2+} saturation state. Aggregate thicknesses of one metre or more are known. More complex crystal structures and rounded microcrystalline 'clouds' form in static pools. Water surfaces are marked by shelfstone around the edges and floating rafts of calcite accreted to dust particles.

Distribution and abundance

Speleothems can occur as isolated individuals, in clusters, aligned along fractures, or broadcast. Density can increase until all surfaces are covered. Vadose speleothems grow most readily at shallow depths beneath soils rich in CO_2 in tropical and temperate conditions that permit year-round deposition; the largest individuals and greatest densities are found in these settings.

Many caves have several levels. Speleothems are often fewer and smaller in the lower levels, which are usually younger. In any setting speleothem deposition is largely prohibited where there are impermeable beds, e.g. shales, above a cave.

Growth rates, age and environmental studies

Under optimal conditions (large excess of Ca^{2+} ions, high drip rate and evaporation) straws may extend several centimetres in one year. Normal accretion rates in other speleothems probably range between ~1.0 mm/10^3 yr in cold climate flowstones to >1.0 m/10^3 yr in warm cave entrances. Some speleothems grow at constant rates while others vary by factors of ten or more. Many contain hiatuses caused by drought, cold or change of groundwater routing.

Many speleothems can be dated accurately (+/−1 per cent error) by the $^{230}Th/^{234}U$ method if they are less than ~550 kyr in age. Variations of $^{18}O/^{16}O$ isotope ratios during growth may indicate palaeotemperature changes and $^{13}C/^{12}C$ ratios suggest changes of vegetation amount or type. Where present, annual or event banding revealed by u/v fluorescence and other techniques now permits very high resolution reconstructions of past conditions above caves (Hill and Forti 1997: 271–284).

Gypsum speleothems

Gypsum is deposited in three principal modes: (1) as evaporitic growths within bedrocks or cave sediments, which they rupture – 'evapoturbation'. (2) As scattered encrustations or excentric extrusions on rock, sediments or calcite speleothems. Most frequent are 'flowers', extruded, twisting fibrous bundles up to 50 cm in length. Needles grow from sediments and 'hair' from roofs. Larger, bifurcating stalactites are known in a few caves. (3) As regularly bedded floor or wall encrustations in evaporating pools: thicknesses of several metres occur in Carlsbad Caverns, New Mexico, where much is reprecipitated from alteration crusts formed by H_2SO_4 reacting with the limestone walls.

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DEREK C. FORD

SPHEROIDAL WEATHERING

Spheroidal weathering is the phenomenon whereby joint blocks within the regolith become rounded as a result of the separation of concentric layers of block surfaces. Spheroidal weathering is common in basalt and granite but is also found in dolerites, andesite and some sandstones (Heald *et al*. 1979).

There are two main schools of thought concerning the cause of spheroidal weathering. The first envisages that the separation of shells occurs due to residual stress from cooling and contraction. However, this would not explain the presence of corestones in sedimentary rocks such as sandstones. Ollier (1971) also makes the case that spheroidal weathering is a constant volume process, i.e. there is no accompanying expansion or contraction as is the case with EXFOLIATION and SPALLING.

The second school of thought envisages chemical activity, primarily the process of HYDROLYSIS, leading to migration of mineral elements and their concentration in separate layers. Thus distinct bands of accumulation and depletion become established (Augustithus and Ottemann 1966). Activity is most prevalent at corners and edges and so the tendency is for angular joint-bounded blocks to become rounded, or spheroidal in shape.

Ollier (1984) argues that use of the term spheroidal weathering should be restricted to situations where the weathering front attacks the block from all sides, i.e. the block must be beneath the ground surface. Clearly boulders originally rounded by spheroidal weathering may subsequently become exposed at the surface due to erosion. At this point, the style and rate of weathering are likely to change significantly. If the REGOLITH is subsequently completely removed by erosive agents, landforms known as boulder fields and boulder tors may remain on the surface.

The shape and size of corestones and boulders are determined by the spacing of joints in the intact mass. Spheroidal weathering is likely to be more effective in more closely jointed rocks with wide apertures (gaps between blocks). In this way, the opportunity for permeation of the rock with chemical solutions is optimized.

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SEE ALSO: weathering

DAWN T. NICHOLSON

SPIT

An easily recognized deposition coastal landform, which belies the potential complexity of the formation processes (Gilbert 1890; Davis 1896; Zenkovich 1967; Carter 1988). Spit structure and formation are best analysed through a plan-view perspective, as the coastal configuration in which the feature is developed is crucial to the spit's formation. Spits are found on an irregular coastline where sediment availability and wave power allow a constructional smoothing of the coastline by maintaining open coast longshore beach direction (in the form of a spit) into coastal re-entrants/bays.

Spits are essentially narrow depositional embankment-type features that show a dominance of longshore sediment deposition (growth) over cross-shore sediment movement. A spit's elongation relative to width is an indication of both the coastal sediment availability and net longshore-directed wave-generated transport potential. Such sediment can be from a broad size range, though sand-dominated spits are the most common. Gravel-dominated spits are more likely in mid-upper latitudes where gravel is a major component of coastal sediment availability. As spits are essentially a product of breaking wave activity, mud-dominated spits are unlikely to be observed. A spit's presence generates a back-spit energy lee with low-energy currents (tidal and small wave) and fine sediment stores (tidal banks and marshes).

A spit's plan-view shape is linked to the plan view of breaking-wave crests (and hence longshore transport vectors) determined by near-shore bathymetry. Spits tend to develop where wave refraction cannot accommodate to sudden changes of coastal trend and rapid reduction in breaker approach angle reduces the longshore drift rate to zero at this point. This allows beach deposition to overshoot the directional shift in coastline. The spit builds from this depositional nucleus, its orientation a function of wave refraction accommodating to the changing near-shore bathymetry induced by the presence of the spit.

Spits show a sequence of planform changes that are related to variation in both sediment supply and longshore transport potential, and are best developed when near-shore wave approach is angled along the spit. Spits can occur within a re-entrant when wave crests approach parallel to the re-entrant mouth. The regularity of the spit's plan-view form is a function of wave direction and refraction consistency. A spit is connected to the coast and its proximal sediment source by the neck, while a spit extension occurs at the spit's distal end or terminus. A spit per se, is usually only the subaerial (superstructure) expression of a larger submarine feature (spit platform), the distal position of which is the spit ramp. Ramp deposition controls spit growth and usually has a high fine-sediment proportion, even in gravel-dominated spits, related to wave-generated currents. As most of the sediment for the spit platform is supplied by longshore transport, it mimics sediment availability to the superstructure, though tending towards finer sediment. The spit platform requires an increasing sediment volume as the spit progressively builds into deeper water and as the volume of the superstructure generally remains the same, spit elongation rates will decline over time if the longshore sediment supply rate does not increase. Thus sediment supply rate is a major control on spit development.

There is rapid wave shoaling and landward curvature of the breaking wave crest at the spit terminus given its steep bathymetric gradients. This leads to curvature of the distal structure against the general trend of the spit. Curvature, correlating with decreasing breaker height and sediment fining, is enhanced at times of diminished longshore sediment supply. When supply is reinstated, then the spit can extend in line with its original plan and the recurve is isolated. High volume, but episodic, sediment supply can lead to drift-aligned spits where the spit plan outline is essentially rectilinear despite overlapping recurves (Carter and Orford 1991). This scenario is often associated with the initial formation of spits in a disjointed coastline where sediment supply is formed from isolated finite sediment sources (e.g. a drumlin coastline: Orford *et al.* 1996). Once the spit's sediment source is exhausted then continuing wave power starts to rework the existing spit sediment (cannibalization) leading to a thinning of the spit neck, the increased potential of the superstructure to rollover (as bigger waves break closer to the shore) and the landward movement of the spit into swash-alignment. Cannibalization leads to extension or stabilization of the distal position which then tends to act as a down drift hinge position for control of spit form's swash alignment. Spits generally retreat under a rising relative sea level through overwashing and hence rollover. Retreat evidence is provided by truncated back-spit recurves and back-barrier organics emerging on the seaward face.

Tidal currents can influence the spit terminus, especially when spit growth squeezes coastal inlet width. As inlet tidal hydraulic efficiency increases, any ebb-tidal asymmetry leads to protection of the spit terminus by sand shoals related to delta formation. These in turn influence wave refraction at the terminus and allow swash bars to drive onshore and build up the terminus beach. This can be a source for distal aeolian dunes, even given sediment depletion elsewhere on the spit. Changes to delta deposition due to back-barrier reclamation can affect terminus growth: a decrease in sediment supply can lead to a withering of the spit terminus, thinning of the spit and spit beheading by overwash. The control by ebb deltas on wave refraction patterns can lead to apparent opposing spit growth across tidal inlets despite a dominant single direction of sediment supply. Spits that seal off re-entrants form barriers (see BARRIER AND BARRIER ISLAND).

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JULIAN ORFORD

SPRING, SPRINGHEAD

Springs are points where ground water, recharged at higher elevations, emerges at the surface. Depending on the nature of the recharge and of the storage/transmission characteristics of the aquifer through which the water has flowed, they may be permanent (perennial), seasonal or intermittent. In karst areas reversing springs called estavelles are found, particularly in association with poljes. Another common feature of karst areas is the presence of permanent 'underflow' springs and higher, intermittent, 'overflow' springs. Spring flow may show little variation over time or may respond rapidly to recharge, varying over several orders of magnitude. Some springs where the outflow is controlled by a siphoning reservoir system exhibit regular ebbing and flowing with a typical period of minutes to hours. Geysers are periodic hydrothermal springs in which a pressurized body of water is warmed to boiling point and explosive spontaneous boiling occurs as pressure is released.

Springs are found at many elevations from high in mountains to beneath sea level, the vrulja of the Mediterranean being an example of the latter. Spring discharges range over seven orders of magnitude, from seeps to large springs with average flows exceeding $20\,m^3\,s^{-1}$ and instantaneous flows of several hundred $m^3\,s^{-1}$. Most of the largest springs are karstic and only those from fractured volcanic rocks rival their output. The largest is thought to be the Tobio Spring in Papua New Guinea with a mean annual discharge of $85–115\,m^3\,s^{-1}$.

Springs which discharge only percolation water are called exsurgences while those that discharge a mixture of percolation water and water from sinking streams are called resurgences. The term 'rising' is commonly used by speleologists as a synonym for a spring. Those springs where water rises from depth are pressure springs, sometimes termed 'Vauclusian' springs after the Fontaine de Vaucluse in France which has been explored to a depth of 315 m. Where the internal hydraulic head in an aquifer greatly exceeds that required to drive the flow of water, springs exhibit a marked upwelling and are commonly termed 'artesian'.

Karst springs are the output points from a dendritic network of conduits, some of which may be large enough for human exploration (caves). They therefore tend to be both larger and more variable in quantity and quality than springs that emerge from coarse granular or fractured media. The latter may result from the convergence of flow lines in a depression or from the concentration of flow along open fractures such as faults, joints or bedding planes.

Where a spring with a moderate to large discharge has emerged at the same point for a long period a marked springhead or steephead may form. Valleys that begin abruptly at the springhead are termed pocket valleys or reculées. Most are short but some are many tens of metres in height. They may form by headward recession, as water from the spring undermines the rock above it, or by cavern collapse.

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JOHN GUNN

STACK

Stacks are isolated pillars of rock that form when part of a retreating coast is separated from the mainland, usually along joints (see JOINTING) or faults (see FAULT AND FAULT SCARP). Stacks also develop because of folding, tropical KARST submergence, solution pipes, induration and variable rock types. They form in fairly strong rocks with well-defined planes of weakness and are uncommon in weak or thinly bedded rocks with dense joint systems. Stacks can develop from the collapse of arch (see ARCH, NATURAL) roofs, but many form directly from erosion of the cliff face.

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ALAN TRENHAILE

STEP-POOL SYSTEM

Step pools are characteristic bedforms that dominate the channel morphology of steep mountain streams (Chin 1989). Steps are generally composed of cobbles and boulders; they are separated by finer materials forming the pools. Steps and pools alternate to produce a repetitive sequence of bedforms, with a longitudinal profile resembling a staircase. The step-pool morphology similarly develops in bedrock channels and in vegetated basins where channels incorporate woody debris to produce log steps. Step pools are part of a continuum of coarse-grained bedforms that includes POOLS AND RIFFLES (Montgomery and Buffington 1997). Despite external influences, step pools commonly occur with sufficient regularity to produce a rhythmic streambed. They represent a type of meandering in the vertical dimension (Chin 2002).

Step pools are functionally important because they provide hydraulic resistance. Steps induce water to plunge into pools below, promoting tumbling flow where much of the flow's kinetic energy is dissipated by roller eddies. By causing a vertical drop in the water surface elevation as water flows from step to pool, steps also decrease potential energy that otherwise would be available for conversion to a longitudinal component of kinetic energy used for erosion and sediment transport. Steps provide the ability to counteract steep slopes, thereby preventing excessive erosion and channel degradation. The role of step pools is especially important in confined mountain streams where lateral adjustments and energy dissipation by meandering and braiding are prohibited.

Step pools form an integral part of the hydraulic geometry of mountain streams. Consistent relations exist between step length, step height and channel gradient. Step length increases and height decreases with a decrease in slope. Such relations are found regardless of substrate type and the presence of woody debris, suggesting that step characteristics are controlled, at least in part, by flow energy expenditure (Wohl *et al.* 1997). Step pools represent a means of adjusting boundary roughness. They evolve toward a condition of maximum resistance, which is apparently achieved when the ratio of mean step height to mean step length to channel slope is between 1 and 2 (Abrahams *et al.* 1995).

Step pools are mobilized by high-magnitude, low frequency floods on the order of fifty years or more (Grant *et al.* 1990). The specific generating mechanism is incompletely understood. Laboratory experiments suggest that steps may originate as antidunes under high flows (Whittaker and Jaeggi 1982). However, although limited field data support the antidune model, the theory cannot explain step-pool formation in all cases. For example, step pools develop in some channels where flows are unlikely to completely submerge clasts and form antidunes. Alternative explanations focus on flow instabilities and the random movement of large particles.

The step-pool bed configuration controls hydraulics and sediment transport in distinct ways.

For example, velocity and flow resistance fluctuate between step and pool and along with increasing discharge (Lee and Ferguson 2002). Sediment transport is episodic, characterized by alternating transport steps and intervals of non-movement (Schmidt and Ergenzinger 1992). Thus, prediction of sediment transport in step-pool streams using standard equations is problematic. New data from instrumented watersheds (e.g. Lenzi 2001) have the potential to yield considerable insights for bedload transport and associated step-pool processes.

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SEE ALSO: gravel-bed river

ANNE CHIN

STERIC EFFECT

An effect in which the molecular dimensions of the material controls the rate or path of a physical or chemical reaction. A steric effect on a rate process may lead to a rate increase (steric acceleration) or a decrease (steric retardation). The most significant steric effect in geomorphology is SEA LEVEL change. In sea water, steric effects are driven by changes in temperature and differences in salinity (i.e. density). As heat is able to exchange freely with the atmosphere, temperature is the dominant steric parameter, particularly over longer timescales (thousands of years) where salinity remains fairly constant in oceans. The change in seawater level as a result of steric effects is referred to as the steric height. This is defined as the height of the sea as the integral of the specific volume from a specified pressure level to the ocean surface. However, steric heights are hard to calculate and trends difficult to attain, as changes occur over varied spatial and temporal scales.

Steric changes occur on seasonal and intra-annual time periods, mostly as a result of steric effects in the upper 500 m of the oceans where heat exchange is much more rapid. It is estimated that an increase in temperature of 1 °C throughout the uppermost 500 m will result in a sea-level rise of approximately 100 mm. In comparison, colder deeper waters show slower heat exchange, and it has been argued that they should be ignored in steric height calculations. However, a parcel of water at 4 °C at 2,000 m depth has a thermal expansion coefficient (which increases with temperature and pressure) 60 per cent as large as that of a parcel at the ocean surface at 20 °C (Roemmich 1990). A warming of the entire seawater column of 1 °C would raise the sea level by about 0.2 m, whereas a warming of 10 °C would lead to an increase in sea level of about 8 m (as the coefficient of thermal expansion increases with temperature) (Knutti and Stocker 2000).

Steric height variations have been shown to influence global sea-level fluctuations on short-term timescales in particular (seasonal, intra-annual and decadal periods on the order of 10 cm), yet steric effects are not great enough to account for sea-level variations over greater timescales (for instance, the 150 m rise in sea level since the last glacial period) (Roemmich 1990). However, thermal contraction of sea water could account for a sea-level rise between the climatically warm Cretaceous period (144–66 Ma) and the onset of the ice-dominated system in the Cenozoic.

Steric effects hold great contemporary importance with regards to global warming, and several modelling studies have investigated past changes and the likely response of the oceans to future changes (e.g. Knutti and Stocker 2000). A rise in sea level of between 10–50 cm is projected due to steric changes over the next century.

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STEVE WARD

STONE-LINE

Stone-lines are synonymous with 'stone-layers', *nappes de gravats* (French) and *Steinlagen* (German) and are common stratigraphic features within many tropical soils and weathering profiles. They form striking, mainly undulating and approximately downslope-oriented discontinuities in soils, consisting of stringers of resistant, largely unweathered coarse clasts at different depths below the groundsurface. In thickness they may range from several centimetres up to a metre or even more. These layers of coarse material separate the overlying loamy to sandy topsoil, or hillwash, from the strongly chemically weathered subsoil or SAPROLITE. In most cases the substrate is not well sorted and may be angular to subangular or possibly well rounded in shape, comparable to gravels and pebbles. Although usually dominated by quartz, pisoliths, iron nodules and larger fragments of lateritic crusts, as well as human artefacts can occur.

A three-stepped stratigraphic subdivision of the stone-line complex was developed for Central Africa, distinguishing the cover (hillwash) or α-layer, the stone-line or β-layer, and the weathered rock (saprolite) or γ-layer (Stoops 1967). The α-layer is typically a few centimetres to some metres thick. It consists of loose but structured material, of sandy to clayey texture, practically devoid of elements coarser than 4 mm (with the exception of some loose iron concretions, mainly at the surface). It shows no stratification. The hillwash follows the general slope of the hill, albeit with a locally more sinuous path with introversions. The sinuosity amplitude varies between 2 and 4 m, and the height-differences seldom reach 50 cm. At the transition from the hillwash (α-layer) to the stone-line (β-layer), slightly coarser material up to 1 cm in diameter occurs. In striking contrast to the covering hillwash, the stone-line often shows a vertical zonation dominated by weathered vein-quartz and rolled pebbles with iron coatings, which are reminiscent of alluvial gravel. Prehistoric implements have been identified in some locations (β_1-layer). Along some sections only, a gradual transition to the lower, so-called β_2-layer takes place. This part of the stone-line consists essentially of fragments of *in situ* weathered quartz veins and chert bands showing a subangular to angular shape. Surface coatings of clasts are rare. Thickness of this layer can reach several centimetres to two metres. Below the stone-line a profound weathering zone characterized by kaolinitic clay with a subsequent transition to bedrock (mottled and pallid zone) is recognizable. A further subdivison of the γ-layer relating to soil colour and micromorphological properties, in connection with successive alteration of the soil profile (e.g. γ_1-, γ_2-, γ_3-layer) may be present.

Due to a huge variety of stone-line phenomena in the tropics an extensive literature on these stratigraphic features exists, indicating that several theories and geomorphic processes may be considered in explaining their morphogenetic origin. Early in stone-line research in the late 1940s it was thought that stones, originally dispersed over the whole depth of the soil profile, were able to sink through the matrix of fines, and would finally concentrate on top of the underlying bedrock. However, this conception proved wrong as the bearing capacity of a soil generally remains high enough to support stones. Today, it is generally accepted that stone-line formation is closely linked to a catenary context, especially to the domain of hillslope processes and slope morphology. It is equally related to long-term weathering and morphodynamics of landscapes. Nevertheless the interpretation of different stone-line features, in particular whether they are the result of an *in situ*, autochthonous formation by down-weathering, or whether their origin lies in a combination of laterally active morphodynamic processes due to former palaeoenvironmental modifications of the landscape (allochthonous formation) is still controversial.

Stone-line formation

The following explanations of stone-line formation are most likely:

1 Stone-lines are residual surface accumulations (palaeopavements) which were later covered by finer sediments. This process results from selective erosion by episodic sheet wash (see SHEET EROSION, SHEET FLOW, SHEET WASH), soil creep and the formation of slope pediments with retreating scarps, causing the hillslope-oriented accumulation of colluvial- or hillwash-like fine material. This landscape instability is effected by a climate change to drier, arid conditions (Rohdenburg 1969).
2 Stone-lines occuring near river plains can be understood as parts of palaeochannels (e.g. former anastomosing branches), that were formed by redistribution and concentration of gravel by surface water flows and related colluvial activity.
3 Stone-lines may be considered as the result of bioturbation by termites, ants and worms. Selective zoogenic uptake of fine material from bottom to top in a soil leads to a concentration of coarser material in greater depth.
4 It is most likely that the formation of many stone-lines has to be considered within the context of drier climatic conditions during the Last Glacial Maximum (LGM). However, there is also evidence that stone-lines can be interpreted as stratigraphic markers of a Younger Dryas event with cold and dry climatic conditions at the onset of the Holocene (Runge 2001).

Economic significance of stone-lines

The coarse material is frequently quarried for road construction as paving gravel and for mineral exploration (e.g. cassiterite, columbite, gold, monazite, zircon, rutile, ilmenite, diamonds). Due to their greater specific weight and greater mechanical and chemical weathering resistance these minerals are concentrated in stone-lines (placer deposits). Such sites are often the first to be exploited as no heavy equipment is required. The mineral content of stone-lines is often used as a pathfinder towards major hardrock orebodies (Thorp 1987).

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SEE ALSO: saprolite; sheet erosion, sheet flow, sheet wash

JÜRGEN RUNGE

STONE PAVEMENT

Sometimes called 'desert pavement', a stone pavement is an armoured surface composed of a thin mosaic of rock fragments that is set in or on a matrix of finer material. In Australia the rock fragments may consist of SILCRETE fragments, locally termed *gibbers*. Pavements are important because they are a major control on surface stability. They also provide a record of the activities working on desert surfaces. In addition, if they are disrupted, accelerated erosion of the underlying finer material may occur. Pavements also act as a store for material in transit by the wind and may fundamentally affect infiltration, runoff and sediment erosion rates (Poesen *et al*. 1994). They tend to occur in areas with limited vegetation cover, and the presence of vegetation, as in wet years, can encourage increased levels of bioturbation and surface disturbance (Haff 2001). Stone pavements are not restricted to deserts, however, and occur, *inter alia*, in tundra regions.

Stone pavements may display soil horizons or they may be produced without appreciable soil development, especially in the case of immature examples developed on relatively unstable ground surfaces by superficial processes, such as deflation and runoff. In those examples with soil horizonation, there is often a vesicular A horizon of mainly silt-clay-size particles (McFadden *et al*. 1998).

There has been a great deal of discussion about the processes that lead to pavement formation.

A classic model is that of deflational sorting, whereby a lag of coarse material is left at the surface after finer materials have been removed by wind action. Another possible mode of horizontal removal of fines needs to be considered, however, namely water sorting (Cooke 1970). Upward migration processes may also play a role, for the concentration of coarse particles at the surface and at depth, and the relative scarcity of coarse particles in the upper part of the underlying soil profile, suggests that the coarse particles may have migrated upwards through the soil to the surface. This could be achieved by a range of processes, including freezing and thawing, wetting and drying, changes in salt phases and bioturbation. In addition, pavement characteristics may be much modified by the addition of aeolian materials, especially dust, from above (McFadden *et al*. 1987; Wells *et al*. 1985). Pavement characteristics also change with age, with pavement development being greater on older surfaces (Amit and Gerson 1986). In particular, the nature of older surfaces will be characterized by a greater degree of particle weathering caused by processes like salt weathering. They may also display greater degrees of rock varnish cover.

In recent years there has been great interest in the way that pavements recover from disturbance brought about, for example, by the passage of wheeled transport. In the Western Desert of Egypt, vehicle tracks from the two world wars of the twentieth century can still be clearly seen on pavement surfaces. However, in other localities pavements have been seen to heal relatively rapidly following deliberate local disruption of their surfaces. Haff and Werner (1996), working in California, found that gaps healed in around 5 years and that displacement of surface stones by small animals was a major component of the healing process. Similarly, Wainwright *et al*. (1999), using rainfall simulation experiments at Walnut Gulch in Arizona, found that raindrop erosion processes resulted in rapid surface recovery.

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A.S. GOUDIE

STORM SURGE

Storm surge is a response of the ocean to changing atmospheric pressure and strong winds caused by cyclonic weather systems, that can result in higher water surface elevations than are predicted by normal astronomical tides, lasting between an hour and four days but typically in the order of 6–18 hours. Storm surge results from the combined action of extreme wind shear stress on the ocean which moves and holds water against windward coasts (wind set-up), and the inverse-barometer effect of changing atmospheric pressure that increases the mean water surface level as pressure drops (pressure set-up). Pressure set-up increases the average water surface by 1 cm per 1 hPa drop in atmospheric pressure, but in cases of extreme storm surge, wind set-up is much more significant. Wind speeds, the track and the relative position of the storm centre to the shore, the slope of the continental shelf and the configuration of the shoreline (particularly the extent of embayment) are all influential in determining the size of the storm surge. The largest storm surges result from hurricanes (otherwise known as tropical cyclones or typhoons) and raised water levels of up to 8 m have been reported. Significant storms at higher latitudes tend to produce surges in the order of 1–3 m, although on shallow continental shelves higher surges are possible.

In generally low-lying coastal areas, increases in mean water level by storm surge can result in large areas of land inundation, often coinciding with floods and other storm-related effects. The most significant regularly affected locations for storm surge damage are the Bay of Bengal, the south-east coast of the USA and the east coast of China. Surges in the shallow, funnel shaped Bay of Bengal have resulted in more than 100,000 deaths on four occasions since 1897 (Bao and Healy 2002), with the worst event occurring in 1970 resulting in the loss of approximately 300,000 lives. In the USA, although surges result in inundation of significant areas of land and high economic loss, deaths have been few due to good warning systems, and disaster response infrastructure. Considerable effort has gone into understanding and modelling storm surge (Bode and Hardy 1997).

The effects of storm surge can be exacerbated if the surge coincides with periods of high astronomical tide. Additionally, on open coasts storm surge is normally associated with increased wave set-up and the establishment of long-period wave motions in the surf zone, all of which increase water level on the beach, allowing storm waves to penetrate much further inland which can lead to significant coastal erosion.

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SEE ALSO: continental shelf; overwashing; wave

KEVIN PARNELL

STRANDFLAT

The word ‘strandflat’ is a name used for the low country and shallow sea along the western Norwegian coast, and also along coasts in Arctic and Antarctic areas that have been covered by ice sheets during the Quaternary ice age. Apart from long stretches of the west coast of Norway where the strandflat is an almost continuous feature, the strandflat has also been recognized in areas as far apart as the South Shetland Isles, Alaska and western Scotland. The low areas of strandflat often appear as broad glacially moulded coastal rock platforms sometimes as much as 80 km in width and backed by high cliffs. However, these shore platforms generally exhibit considerable local relief thus making it difficult to assign a precise altitude to any individual area of platform.

The strandflat was first described by Reusch (1894) while its possible origins were first considered in detail by Nansen (1922). The various processes of strandflat formation are well summarized by Larsen and Holtedahl (1985) and include marine abrasion, subaerial weathering, glacial erosion, frost shattering and cold climate shore erosion. Larsen and Holtedahl proposed that the strandflat was primarily the result of sea-ice erosion and frost shattering during the Quaternary and that most surfaces had been later modified by marine erosional and glacial erosional processes. They also noted that the Norwegian strandflat surfaces exhibit glacio-isostatic tilting and are therefore likely to have been produced during periods of Quaternary glaciation rather than during temperate interglacial periods.

It is difficult to determine precise ages for the formation of the various strandflat surfaces around the world although recent developments in cosmogenic isotope dating techniques represent one possible way forward of dating individual rock surfaces. Consideration of the Quaternary marine oxygen isotope record and of the glacio-isostatic changes that have affected land areas buried by Quaternary ice sheets implies that the position of relative sea level in any area is unlikely to have remained stationary for any significant length of time, certainly no less than *c.*10,000 years. Accordingly, this would seem to point to the conclusion that individual strandflat surfaces having been produced by cold climate shore processes, must have been repeatedly overwhelmed by ice sheets and subject to marine processes during numerous intervals of cold climate throughout the Quaternary.

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ALASTAIR G. DAWSON

STREAM ORDERING

Stream ordering is a technique for characterizing the constituent parts of a drainage network. Ordering can start from the outlet and move in the upstream direction or it can start from each source and move downstream. The most successful have been those ordering systems which move in a downstream direction. The upstream moving systems require a series of subjective decisions about which upstream extension is the master stream. Horton (1932, 1945) introduced the following ordering system:

(a) Channels that originate at a source, and have no tributaries are defined to be first-order streams;
(b) When two streams of order x join, a stream of order x + 1 is created;
(c) When two streams of different order join, the channel segment immediately downstream of the junction takes the higher order of the two combining streams;
(d) When the highest order stream segment (n) has been defined, then the upstream extensions of that segment are deemed to have the same order (n) all the way to the source. Similarly, stream segments of order (n−1) are extended back to their source and so on.

This hybrid system (first downstream and then upstream) incorporates the subjectivity mentioned and therefore Strahler (1952, 1957) revised Horton's scheme by removing step (d) above. This so-called Strahler system (or Horton-Strahler ordering system) is the most commonly used in hydro-geomorphology.

A third-ordering system is the link magnitude system proposed by Shreve (1966). Source streams or links have magnitude 1. At a bifurcation, the downstream link takes the magnitude of the sum of the magnitudes of the two incoming links. The magnitude of each link is therefore equivalent to the number of sources in the network draining into that link.

The theoretical basis for Shreve's ordering system is his view of the river basin as a random topological structure. Terminology used includes: a node is the one outlet furthest downstream; n sources are the points furthest upstream and there are n − 1 junctions. Edges of the network are links; exterior links emanate from sources; interior links emanate from junctions. A network with n sources has 2n − 1 links, n of which are exterior and n − 1 are interior links.

The most important measuring device that Horton identified was that involving stream ordering. The idea originated with a German hydraulic engineer Gravelius (1914). Chorley (1995) noted that there were two important corollaries of Horton's stream ordering: (a) it placed the emphasis on analysis based on the identification of individual drainage basins. The latter thus emerged as rational, clearly defined topographic units whose geomorphic status was expressed by their order and which prompted geometrical comparisons from one location to another; and (b) the procedure generated a nested hierarchy of drainage basin forms, each of which could be viewed as an open physical system in terms of inputs of precipitation and outputs of discharge and sediment load. A third corollary could be added to Chorley's list. The application of increasingly refined statistical analysis to problems of watershed geomorphology was facilitated by ordering the channels: issues of sample size and representativeness became central considerations in geomorphology.

Perhaps most importantly, the concept of drainage density, which was to become the most important geometric indicator in the work of the Columbia school of quantitative geomorphology, was encouraged by the ordering of streams. Melton (1957) explored the relation between drainage density and stream frequency, as well as the ratio of the two as a measure of the completeness with which a channel system fills a basin outline and as a possible evolutionary index of drainage basins.

Horton's ordering also generated the laws of drainage composition. These were exponential relations between stream order and (a) number of streams of a given order; (b) average length of streams of each order; (c) total stream lengths of each order; (d) basin areas of each order; and (e) average stream slopes of each order. Each of Horton's laws generated a ratio (e.g. the bifurcation ratio from the first law) and these ratios were shown to lie within quite narrow ranges except where differential geological controls were important. Horton's laws provided a topologically and geometrically logical set of procedures for the analysis of fluvially dissected terrain. Much geomorphic work has subsequently been based on systems of stream ordering.

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SEE ALSO: allometry; drainage density; dynamic geomorphology

OLAV SLAYMAKER

STREAM POWER

Power is the rate of doing work (force × distance) and is expressed in Watts which are Joules per second ($J s^{-1}$). Stream power is the rate at which a stream can do work, especially in the transport of its sediment load, and is usually measured over a specific length of channel. It expresses the rate of energy expenditure in flowing water and, as such, could provide a basic integrating theme within the physical environment (Gregory 1987). In hydraulics and fluvial processes, attempts to analyse the processes involved have used a number of variables whereas expressing energy expenditure as stream power is a more fundamental approach. The potential energy that water possesses at a particular location is proportional to its height above some datum which can be sea level or a lake level; this potential energy is converted into kinetic energy as the water flows downhill under the influence of gravity.

Three important aspects of stream power are: how it is expressed, what controls it, and how has it been utilized and applied. Stream power (ω) was first expressed (Bagnold 1960) as the product of fluid density (ρ), discharge (Q), acceleration due to gravity (g) and slope (s) in the form:

$$\omega = \rho Qgs$$

This expression for power can of course be applied to any fluid, and Bagnold used a similar approach in relation to wind movement over the Earth's surface. Bagnold's (1960) definition has subsequently been applied to the rate of sediment transport (Bagnold 1977) expressed as the amount of energy expended per unit area of the bed. Such unit stream power could be obtained per unit channel width (w) or bed area as:

$$\omega = \frac{\rho Qgs}{w}$$

which, because $Q = wdv$, is simplified to $\omega = \rho gdvs$ thus including depth (d) and velocity (v), and this is often referred to as specific power. Stream power is measured in Joules per second ($J s^{-1}$) (Watts) and unit stream power is expressed per square metre ($J m^{-2} s^{-1}$ or $W m^{-2}$). Unit stream power is expressed per unit length of channel or per unit channel area, and when comparing results from several areas it is important to differentiate between the results achieved by several methods. Values of unit stream power ω range from less than $1\,W m^{-2}$ in flow between rills to $>12{,}000\,W m^{-2}$ in riverine flood flows in India, up to $18{,}582\,W m^{-2}$ for large flash floods, and up to $3 \times 10^5\ W m^{-2}$ for the Missoula flood (Baker and Costa 1987) in the Quaternary, the largest known discharge of water on Earth.

Controls upon stream power can be deduced from its component elements, of which g is constant and ρ, Q and s are variables. Along a single channel, slope may tend to decrease downstream, whereas discharge will increase, and there can be significant variations in water quality and sediment transport which affect the value of ρ. Along a single river unit, stream power tends to peak in the middle parts of some basins; it is lower upstream, where discharges are relatively small, and lower downstream, where river gradients have their lowest values. Stream power can be calculated for the discharge in the channel at any one specific time or it can be calculated for the estimated channel capacity flow or for some flood flow value; the pattern of stream power distribution down valley may vary somewhat in each case.

Applications of stream power have now been made to many aspects of analysis of river systems. Most usefully in relation to *sediment transport*, stream power has been used instead of stream discharge, velocity or bed shear stress to relate to sediment motion and transport, especially that of bedload (e.g. Allen 1977). This approach is arguably more geomorphological than using hydrological parameters. It has also been used more generally as a means of considering the efficiency of sediment transport; by comparing the power needed to transport the sediment along a particular reach with the power actually available, *critical power* was defined as the power just sufficient to transport the sediment through the reach (Bull 1979, 1991). Interest in the significance of large flood events has led to the estimation of *flood power*, including that related to palaeofloods (Baker and Costa 1987), and a threshold for catastrophic modification of the channel or fluvial landscape has been suggested as a unit stream power of $300\,\mathrm{W\,m^{-2}}$ (Magilligan 1992).

Variations in unit stream power along a river channel have been used to explain patterns of the pool riffle sequence; to determine bedform type for specific sediment size; to relate to the channel HYDRAULIC GEOMETRY; and to explain river channel patterns. Such patterns have been classified according to amount and size of bedload and stream power (Schumm 1981) and channel SINUOSITY has been related to stream power (Schumm 1977). Three major types of floodplain (Nanson and Croke 1992) have been differentiated according to stream power values, including High energy ($\omega > 300\,\mathrm{Wm^{-2}}$), Medium energy ($10 < \omega < 300\,\mathrm{Wm^{-2}}$), and Low energy ($\omega < 10\,\mathrm{Wm^{-2}}$), and stream power variations have also been related to the pattern of the river long profile. These are all ways in which stream power can be related to aspects of *channel morphology*, showing how knowledge of spatial variations of stream power can be the basis for useful applications. In the case of British rivers, Ferguson (1981) demonstrated a thousand fold range in the values of specific power with a clear distinction between values of 100 and $1{,}000\,\mathrm{Wm^{-2}}$ in the high runoff, steep slope areas of the west, contrasting with the values between 1 and $10\,\mathrm{Wm^{-2}}$ in the low slope, low runoff areas of the south and east. Relations with channel morphology and their spatial variations can be employed to provide explanations for the pattern of long profiles or channel patterns, for example, by considering downstream variations in power to be with minimum unit stream power expenditure, or between equalizing power expenditure and minimizing power expenditure.

Variations in power can also be useful in *river management* problems, as employed by analysing channel adjustments downstream from river channelization works (Brookes 1987); in this case the relationship between bankfull discharge per unit width and water slope was subdivided according to lines of equal specific stream power, showing that eroded sites had specific powers in the range $25–500\,\mathrm{Wm^{-2}}$ whereas the remaining sites without change had specific powers between 1 and $35\,\mathrm{Wm^{-2}}$. Such examples can be the basis of general guidelines for river managers, with stream power per unit bed area as a criterion for stability in stream restoration projects; a simple classification for guiding river restoration has been developed (Brookes and Sear 1996) by using the ratio between available stream power and erodibility of the substrate. In Denmark, straightened channels tend to recover naturally above a threshold stream power of $35\,\mathrm{Wm^{-2}}$ and it is only channels with very high energies that regain some or all of their original sinuosity, so that thresholds could easily be developed for other river environments. Types of river channel adjustment have therefore been related to thresholds of stream power (Brookes 1990), indicating how temporal variations of stream power can be used to understand *variations over time* that have occurred. In a study of the arroyo systems of the northern part of the Henry Mountains in south central Utah (Graf 1983) it was shown that, whereas stream power decreased in the downstream direction during a deposition period which occurred before 1896 when channels were small and meandering, after 1896 total stream power increased in the downstream direction because channels were in the floors of arroyos that confined discharges and resulted in channel erosion and throughput of sediment. In 1980, however, the rate of downstream change in total power was intermediate between the depositional conditions of the 1890s and the erosional conditions of 1909, with deposition occurring in the smallest and largest channels but not in the mid-basin areas. Stream power can also be used as a unifying theme for analyses of urban fluvial geomorphology (Rhoads 1994) and although values are not always easy to calculate it remains a very important variable in fluvial geomorphology.

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KENNETH GREGORY

STREAM RESTORATION

Stream restoration is the changing of physical, chemical and biological characteristics of a lotic system to match those of a former natural stream condition, or one that has not been disturbed by humans. One common definition of stream restoration is the 'reestablishment of the structure and function of a stream ecosystem' (National Research Council 1992: 17). Ecological restoration, in general, assists the recovery of an ecosystem that has been degraded, damaged or destroyed and restores its historical trajectory (SER 2002).

Other terms in use include stream rehabilitation, reclamation, reconstruction, mitigation, and 'creation' of new functions and values. Reconstructed channels include a new ecological structure so that the desired flora and fauna can return. Rehabilitated or reclaimed streams are partially restored in that only selected functions and values are returned, primarily to serve a human purpose (e.g. flood control, water supply, land stabilization). Most stream restoration projects are actually partial rehabilitations. It is exceedingly difficult to restore all functions and values of the original stream.

Restoring streams that have been routed through pipes is known as 'daylighting'. Many urban (see URBAN GEOMORPHOLOGY) streams have been paved or lined with large rocks, to carry more water faster and without erosion. Efforts to restore these streams are complex, involve many people, lengthy planning times and are very costly (Plate 131).

Plate 131 Concrete-lined channels and restrictive boundary conditions complicate urban stream restorations

Planning: restore to what conditions?

Stream restoration efforts usually cannot achieve pristine or prehistoric conditions because of the massive cultural changes that would be required. Many streams have been dammed, leveed or channelized to convey floodwaters. Full 'restoration' of these streams would require returning processes like flooding, meander migration, channel avulsion, formation and destruction of LARGE WOODY DEBRIS jams, and backwater sedimentation.

Where management approaches alone cannot achieve the desired ecological functions and values, channel reconstructions using large equipment may be needed. With continued monitoring and adaptive management, reconstructed channels can return some lost ecological functions and values.

Streams that function within natural ranges of flow, sediment movement, temperature, channel migration, and other variables, are said to be in dynamic equilibrium. Restoration projects attempt to restore and maintain DYNAMIC EQUILIBRIUM and ecological integrity. Streambank erosion may be part of this dynamic equilibrium if it is balanced and overall dimensions of the channel remain the same. Successful stream restorations integrate geomorphic processes into their designs. Measurement of channel-forming flow conditions is critical to restoring a stable form, pattern and profile (Rosgen 1996).

Restoration design

Design approaches may be based on stream classification and regional curves for hydraulic geometry (Riley 1998; Rosgen 1996), regime and tractive force equations, and reference reaches (Newbury and Gaboury 1993; Rosgen 1996), but the most rigorous approaches feature more sophisticated hydraulic engineering. Risk of failure can be minimized by incorporating the appropriate levels of management (removing disturbance factors) and engineering controls (e.g. weirs, riprap). Planning and design are best accomplished with an interdisciplinary approach (USDA *et al.* 1998), involving knowledgeable fluvial scientists, ecologists and land users.

'Natural channel design' uses reference reaches in dynamic equilibrium with desired ecological functions and values. Stream classification systems have been developed to assist this process and to promote accurate stream morphological descriptions (Rosgen 1996). When used with experience and scientific design approaches, natural channel designs can result in successful restorations. When used as a 'carbon copy' or cookie-cutter approach, however, the results can be less than successful.

Most stream restoration projects involve modifying an existing stream's location, alignment, meander pattern (see MEANDERING) cross-section dimension, longitudinal profile, or aquatic or terrestrial habitat. Streambank erosion, sediment transport, flooding and sediment deposition are processes that support ecological functions. Having these physical processes become self-sustaining is what separates stream restoration from stream stabilization, stream reconstruction or stream rehabilitation projects.

Soil bioengineering

Soil bioengineering can be an important stream restoration component (Figure 159). Live plants, plant materials and man-made materials are used as systems that interface with Earth materials to create stable streams and banks, and achieve the desired ecological functions and values. Practical application of soil bioengineering techniques requires knowledge of overall performance criteria, design flow and sediment conditions, and habitat suitability.

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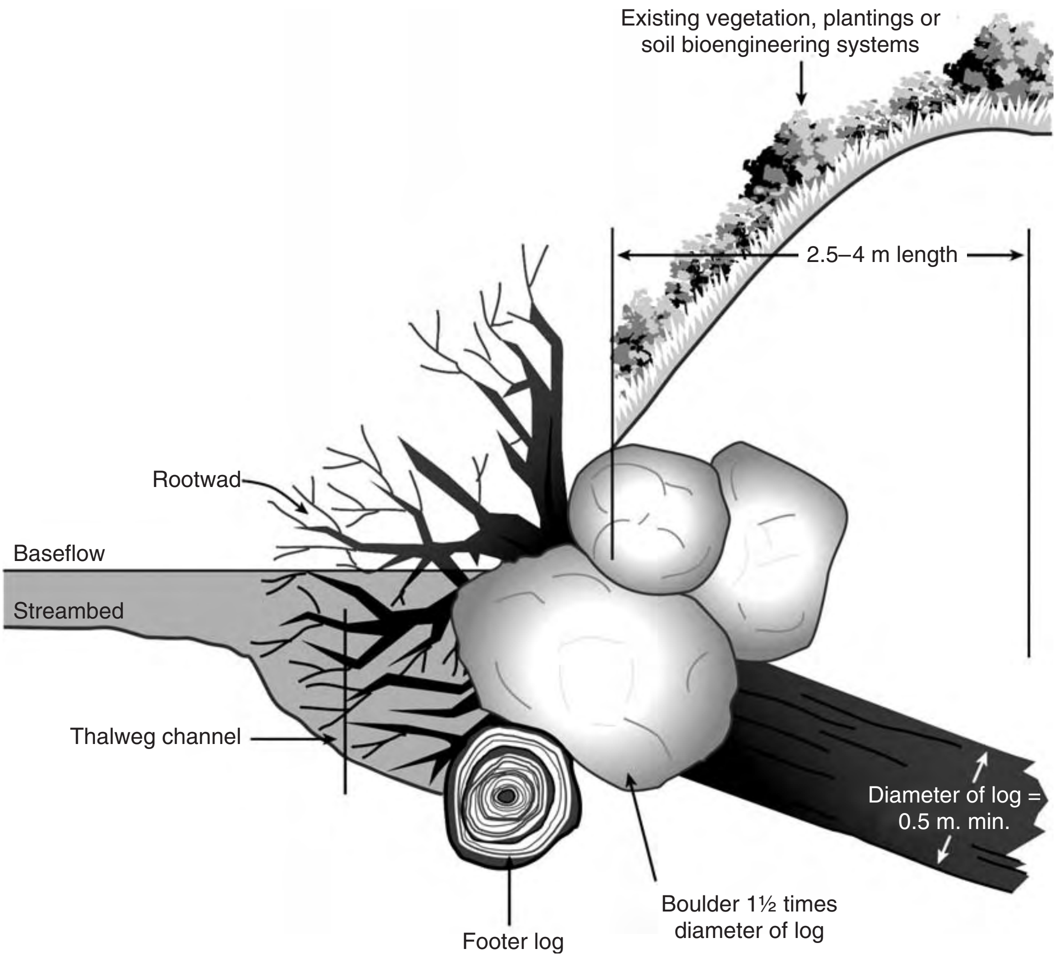

Figure 159 Soil bioengineering example: rootwad

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SEE ALSO: bankfull discharge; biogeomorphology; fluvial geomorphology; mining impacts on rivers; riparian geomorphology; river continuum; river restoration; sediment budget; step-pool system; stream ordering; stream power

JERRY M. BERNARD

STRIATION

Striations are shallow scratches or grooves cut by brittle impact into rock surfaces, boulders or pebbles. Striations may be up to a metre or more in length. They occur widely in areas of former glacial erosion where rock fragments, sand and silt grains transported in the basal ice have impacted surfaces as the ice moved forward jerkily by basal sliding. Some striations have nail head or wedge shapes, with the broad section being at the down-ice end. Striations occur frequently in glacierized areas, especially on fine-grained physically hard rocks such as quartzites and massive limestones. Glacial polish often occurs with striae. Crossing striations may reflect local ice flow variations, changes in ice flows from different glacial centres during one glaciation, or multiple stages of glaciation. Striations may also form where sea-ice, lake-ice, snow banks, screes, large rock masses and debris creep, flow, avalanche, slide, fall or shear over or onto rock surfaces. ICEBERGS may striate the upper surfaces of clasts in current-winnowed boulder lag horizons of glacimarine deposits. Detritus carried by high velocity water flow in jökulhlaups can also form striations in channels eroded in rock.

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ERIC A. COLHOUN

STROMATOLITE (STROMATOLITH)

A term first used by Kalkowsky (1908) to describe some sedimentary structures in the Bunter of North Germany. A more modern definition (Walter 1976: 1) is that they are 'organosedimentary structures produced by sediment trapping, binding and/or precipitation as a result of the growth and metabolic activity of micro-organisms, principally cyanophytes'. They can develop in marine, marsh and lacustrine environments and, though they form today where conditions permit, they reached the acme of their development in the Proterozoic (Hofman 1973). The largest known forms are mounds several hundreds of metres across and several tens of metres high. Gross morphologies vary in the extreme and range from stratiform crustose forms, through nodular and bulbous mounds and spherical oncoids, to long slender columns, erect to inclined, and with various styles of branching. Classic examples are known from coastal regions like those at Shark Bay in Western Australia and from pluvial lake shorelines in areas like the Altiplano of Bolivia where they form massive calcareous encrustrations and bioherms (Rouchy *et al*. 1996).

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A.S. GOUDIE

STRUCTURAL LANDFORM

Structural landforms are those which in their appearance reflect, and are adjusted to, geological structure of underlying bedrock. This effect is achieved through direct or indirect control which structural elements exert on the course and intensity of exogenic processes shaping the landforms. 'Structure' here is usually understood *sensu lato*, i.e. it encompasses such diverse phenomena as facies differentiation, lithological contrasts, fracture patterns, faults and folds, tectonic disposition of strata, geometries of intrusive and extrusive bodies, etc. In general, structural landforms develop through differential weathering and erosion which exploit the structures and emphasize unequal resistance of adjacent rock complexes, whereas relief features created by direct action of endogenic forces fall into the category of tectonic landform (see TECTONIC GEOMORPHOLOGY).

Following the above definition, structural landforms may be subdivided into several categories. There are landform assemblages specific for certain rock types, for example granite (see GRANITE GEOMORPHOLOGY) or carbonate rocks (see KARST). In many cases, their development and appearance are controlled by jointing patterns, in the way that joints of regional extent (master joints) and zones of dense fracturing are exploited towards topographic depressions, whereas more massive compartments are left standing as residual hills, uplands or big boulders. Thus, joint-aligned valleys, rows of sinkholes, basins at joint intersections, and many TORS and domed INSELBERGS may be regarded as structural landforms. A number of small-scale features, such as KARREN on carbonate rock outcrops or deep clefts, may develop along joints and are therefore also structural. In igneous rocks in particular, many structural features originating at the consolidation stage subsequently become avenues for weathering and become decisive for the shape of minor and medium-scale landforms (Twidale and Vidal-Romani 1994).

Another group includes landforms reflecting the variable dip of sedimentary strata and, at the same time, the unequal resistance of consecutive layers against exogenic agents. Undeformed, horizontal layers give rise to plains, if at low altitude, or plateaux, if elevated and bounded by marginal escarpments. A plain or a plateau surface is usually underlain by a resistant rock layer, such as quartz sandstone or massive limestone. Dissection of a plateau may expose underlying strata of variable resistance, the stronger of which will support structural benches on valley sides, such as in the Grand Canyon of the Colorado River. Tilting of strata induces differential denudation, in the course of which rock complexes of lower strength are eroded into valleys or rolling plains, whereas more resistant rocks give rise to parallel ridges, escarpments or mid-slope ledges. If the dip is less than 10°, highly asymmetric ridges called CUESTAs develop. With the dip in the range 10°–30°, less asymmetric monoclinal (or homoclinal) ridges form, whereas in the case of even steeper tilt a symmetric ridge named a HOGBACK will originate. In areas built of dipping sedimentary rocks, drainage patterns usually show much adjustment to structure too. Dendritic patterns are typical for negligible dip, whereas with increasing differential erosion they will evolve into trellis patterns. The spatial arrangement of ridges depends on the regional pattern of deformation. If a simple tilt is involved, ridge axes will generally follow straight lines, perpendicular to the dip. If a central part of a former sedimentary basin is downwarped, concentric ridges facing outwards will develop (the Paris Basin is an example), whereas inward facing ridges will typify domed structures with breached central parts.

In mountain areas built of folded sedimentary rocks typical structural landforms are anticlinal ridges and synclinal valleys or, in the case of INVERTED RELIEF, anticlinal valleys and synclinal ridges. Asymmetric homoclinal ridges and hogbacks occur frequently too, but the degree of structural deformation in mountains is usually so high that the spatial extent of these landforms is limited.

A further category includes landforms built of igneous rocks, the present-day appearance of which may reflect the way of magma emplacement. Large granite intrusions in post-orogenic settings may assume the form of large-radius domes (laccoliths) which, after unroofing, are reflected in the topography as upland terrains, sloping in all directions. Dartmoor upland in south-west Britain is one example. Rising igneous domes induce updoming of overlying strata which then may be differentially eroded towards triangular faces called FLAT IRONs (Ollier and Pain 1981). Smaller linear intrusions, i.e. dykes and sills, are typically composed of material more resistant than the host rock, hence differential denudation leaves them appearing as laterally extensive, vertical, jagged ridges (for dykes) or topographic steps (for sills). Denudation of a former volcano may expose deeper parts of the vent filled with solidified, resistant magma, which then becomes a steep-sided conical hill, called a neck. Necks often display impressive columnar jointing of rock, the most famous example being perhaps Devil's Tower in Wyoming, USA.

Many landforms built of extrusive rocks may also be regarded as structural, including rhyolitic domes and plugs, sloping flanks of shield volcanoes, or plateaux underlain by horizontal or gently inclined lava flows (traps). Subsequent tilting and erosion of a multiple lava flow area may produce an assemblage of landforms similar to those developed on tilted sedimentary rocks.

Structural landforms are of various sizes, from features of regional extent to local manifestations of small-scale structures. Mega-scale examples are extensive plains developed upon flat-lying strata or dome mountains. Medium-scale landforms include plateaux, cuesta ridges, domed hills and intermontane basins. Even smaller are tors and mid-slope benches, whereas joint-aligned pools in a bedrock river bed would indicate the most localized structural control.

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PIOTR MIGOŃ

STURZSTROM

A sturzstrom is a high volume of mostly dry rock material caused by the collapse of a slope or cliff created by large falls and slides moving at high velocities and for long distances, even on a gentle slope (Hsü 1975). Sturzstroms can reach velocities of over $50\,m\,s^{-1}$ and can travel over distances of kilometres. The accumulation volume may exceed 1 million m^3, covering a total surface of over $0.1\,km^2$. In relation to its velocity and dimensions, this kind of landslide can be extremely costly in terms of human lives and damage.

Alternative terms for sturzstrom are rock avalanche, rockfall avalanche or rock-slide avalanche (Angeli *et al.* 1996). Examples of historic events are the Elm sturzstrom of 1881 in Switzerland (Heim 1932), the Valpola rock avalanche of 1987 in the Italian Alps and the Frank landslide of 1903 in Canada. In Europe the highest concentration of these phenomena is found in the Northern and Southern Calcareous Alps (Abele 1974).

A sturzstrom can develop (1) by the fall or slide of a rock body which during movement progressively looses its cohesion by turning into dry debris and, thus, continues its advancement as a debris avalanche, (2) by the sudden mobilization of a debris deposit by a debris avalanche or DEBRIS FLOW, either because of the fall of an overhanging rock mass or by seismic shocks.

Although several investigations have been carried out, no universally accepted explanation has yet been proposed. The mechanical analysis of sturzstroms includes two stages: the initial failure and subsequent streaming. Explanations include turbulent grain flow with dispersive stresses arising from momentum transfer between colliding grains (Cruden and Varnes 1996), fluidization of particles caused by incorporating air, the existence of a cushion of trapped air, acoustic fluidization or rock melting by frictional heat (Erismann and Abele 2001).

In alpine regions the ice retreat of the last alpine glaciation caused significant stress relief on mountain slopes. This unloading is one of the causes of Holocene sturzstroms. However, there is evidence that the process not only took place shortly after the ice retreat (as found between 12,000 and 10,000 BP) but that the loss of shear strength needed much more time to cause the process. The Eibsee sturzstrom (Zugspitze, German Alps, volume: $400 \times 10^6\,m^3$) has been dated at 3,700 years BP which may fit this hypothesis. The exact triggering causes which may enable realistic predictions of sturzstroms requires further attention.

Sturzstroms are highly destructive. Once an event has occurred it is important to discover whether the mass has dammed the valley. If a lake has formed, maximum effort must be given to take control of the breaching dam, because the resulting flow may cause a second disaster.

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RICHARD DIKAU

SUBAERIAL

Refers to all conditions or processes occurring in the open air or on the land surface (e.g. subaerial weathering) as opposed to those occurring in submarine (underwater) or subterranean (underground) environments. The term also applies to materials and features that are created and/or located on the Earth's surface (e.g. subaerial aeolian dunes, subaerial volcano, etc.), and sometimes is inclusive of fluvial forms (of rivers). The notion that all things in the landscape are created by subaerial processes and conditions is termed subaerialism.

STEVE WARD

SUBCUTANEOUS FLOW

The lateral transfer of water in the subcutaneous (or epikarstic) zone. The subcutaneous zone is a highly weathered region in well-developed KARST environments lying in the upper part of the percolation zone, between the soil and the relatively unweathered and permeably saturated phreatic zone below. When water stored in the subcutaneous zone is full (e.g following precipitation) a potentiometric surface (epikarstic water table) is produced. Any further input is subsequently transferred laterally, with movements of water occurring along preferred pathways (neighbouring joints and shafts) down the hydraulic gradient in the epikarstic water table (Williams 1983). Corrosion is intensified where flow routes converge above the more efficient paths, resulting in differential surface lowering, accentuating over time and by irregular distribution of solution. Uniform percolation through the subcutaneous zone will result in uniform surface lowering. Rates of subcutaneous flow vary considerably, commonly between 2–10 weeks, but are hard to gauge (especially in old, well-developed karsts). The term is also used to refer to the process of piping in soils.

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SEE ALSO: epikarst

STEVE WARD

SUBGLACIAL GEOMORPHOLOGY

Subglacial geomorphology has the most profound and indelible imprint on any glaciated landscape. Subglacial geomorphic processes are among the most complex, yet least understood set of glacial processes. Our restricted knowledge stems from the inaccessible nature of what occurs beneath an ice mass and the limited extent of modern analogues. No other aspect of glacial geomorphology impacts on the daily lives of millions of people to such a degree as does subglacial geomorphology, for example, in terms of foundations, roads, railroads, waste disposal sites, agriculture, aquifers and construction materials.

The subglacial environment is that glacial subsystem directly beneath an ice mass that includes cavities and channels that are not influenced by subaerial processes. Subglacial geomorphology considers all aspects of topographic change beneath ice masses as a result of erosional and depositional processes (Figure 160). The subglacial environment is a boundary interface where complex sets of processes interact altering the morphological, thermal and rheological states of the interface between the ice mass and its bed. The key to understanding subglacial geomorphology lies in the mechanics of this interface. This boundary zone migrates across the landscape with every ice advance and retreat. All terrains covered by glaciers have been affected and altered by the passage of this interface. This interface is a function of the prevailing basal ice and bed conditions; a complex relationship between basal ice dynamics, sediments and bedrock, subglacial hydraulics and the glacier bed ambient temperature. Changes in basal ice and/or bed conditions may be widespread or local, and develop rapidly or slowly. These fluctuating conditions may be of enormous magnitude or simply be minor variations at the interface, the former being detectable whilst the latter may leave little or no imprint on subglacial geomorphology.

Subglacial geomorphological processes are constrained by the temperature at the ice–bed interface. Basal thermal regimes can be either temperate or polar but in all likelihood most ice masses are polythermal. In temperate, wet-based glaciers, typical of almost all ice masses today and most ice masses during the Pleistocene and earlier, basal temperatures are found at -1 to $-3\,°C$. In polar, cold, dry-based glaciers, typical of a few isolated central areas of East Antarctica today and possibly the central areas of the vast Pleistocene ice sheets, the basal parts of an ice mass are frozen to the bed with temperatures of -13 to $-18\,°C$ (Van der Veen 1999). Thermal conditions at the ice–bed interface are more

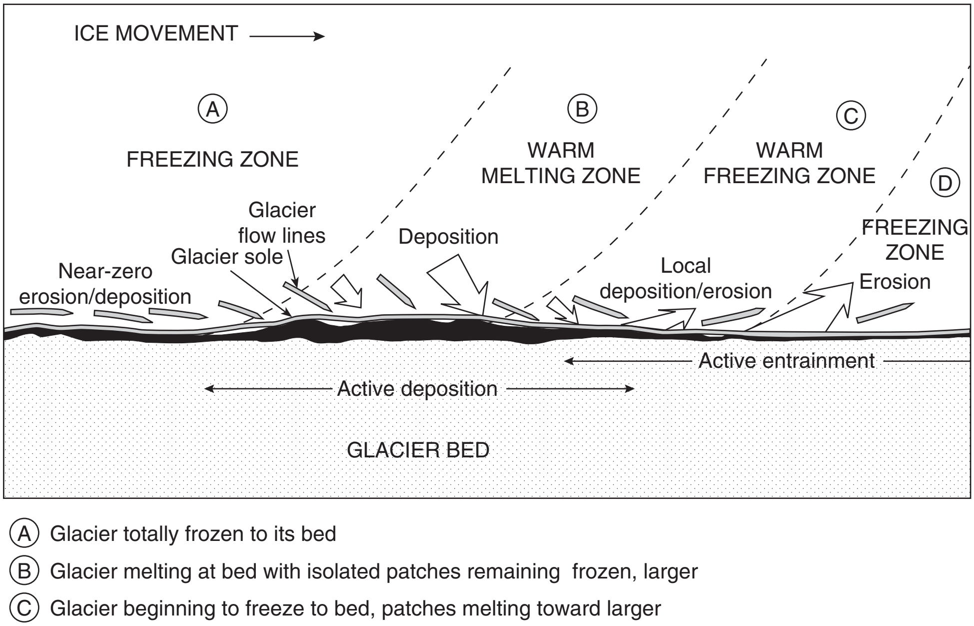

Figure 160 Models of subglacial thermal regimes, their spatial relationships and processes of subglacial erosion, transportation and deposition

complex than implied by these two thermal states, in fact basal ice temperatures vary temporally and spatially producing polythermal conditions (Menzies and Shilts 2002).

In considering ice–bed interface temperature fluctuations, the following parameters converge to establish the specific thermal conditions: (a) rate of snow accumulation and snow temperature; (b) geothermal heat flux; (c) mean annual surface air temperature; (d) ice surface velocity; (e) basal ice velocity; (f) subglacial meltwater flux and temperature; and (g) the imprinted 'memory' of previous thermal conditions (Figure 161a). The relevance of these two extreme thermal states is that under temperate conditions, meltwater occurs at the ice–bed interface, basal ice containing debris can be released, and meltwater processes and/or saturated debris moving as a deforming layer can be accomplished thereby permitting various landforms to develop. Under polar conditions, with the ice mass frozen to its bed, no meltwater is present and only ice movement by plastic deformation occurs thereby geomorphological processes are constrained but not altogether suspended.

It is apparent that polythermal bed conditions prevail in the long term under any ice mass, the switch from wet to dry and back again repeatedly results in localized erosion and deposition of much glacial sediment. Although long distance transport does occur beneath and within ice masses the dominant form of transport is short distance of <10 km resulting in subglacial sediments being typically locally derived, transported and deposited. Figure 160 illustrates the probable relationship between areas of dominantly erosive glacial action and depositional processes under ice sheet conditions. Beneath valley glaciers, due to much shorter distances and thinner ice cover (lower basal ice pressures), a similar, but less complete, set of conditions may prevail (Benn and Evans 1998).

Subglacial erosional landforms

Subglacial erosion processes are pervasive at the active ice–bed interface. In some terrains under certain subglacial conditions, erosive processes become dominant. Almost any description of the impact on the land surface of past glaciers focuses on the grandeur and size of FJORDs and the sculpted bedrock features of once glaciated terrains. However, our understanding of how these distinctive features were fashioned remains limited. Erosional forms exist at an immense range of

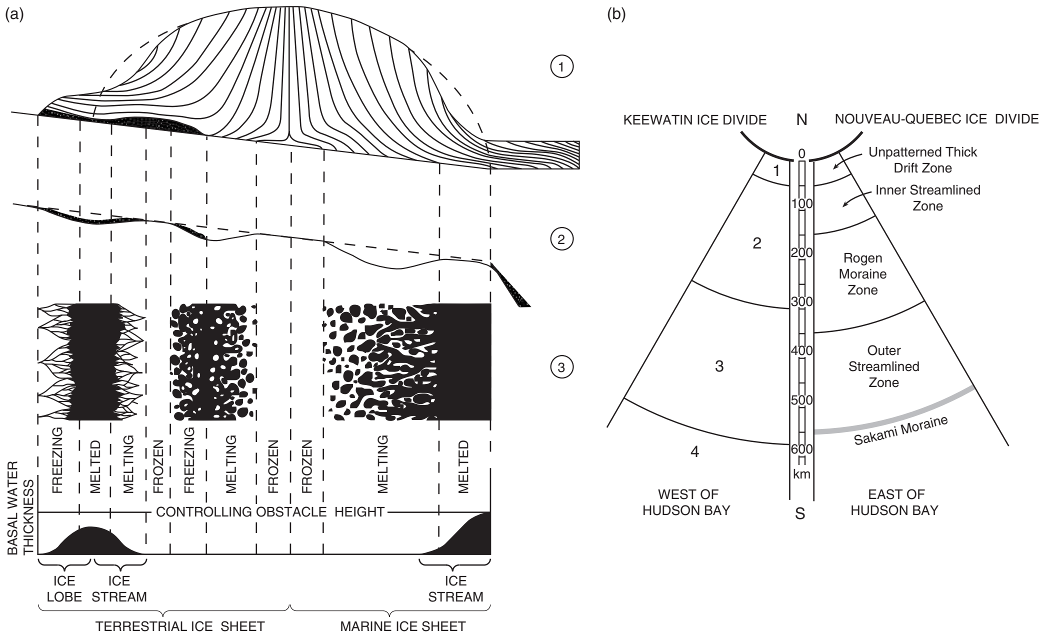

Figure 161 Subglacial thermal zonation in relation to subglacial processes beneath a steady-state ice sheet with terrestrial and marine margins. 1, internal ice flow trajectories; 2, subglacial topography created by subglacial erosion; 3, areal distribution of subglacial meltwater (reprinted with modification after Hughes 1995)

scales from surface microscopic features on particles to the megascale landforms. Regional erosional features can be subdivided into regional and local forms. However, some forms, such as ROCHES MOUTONÉES, occur at all scales. Regional erosional features are either areal (spatially pervasive) or linear (spatially discrete) landscape types. In the former case, scoured terrain develops where limited debris existed at the ice–bed interface and dominantly polar bed conditions prevailed. In contrast, linear erosion processes are differential in their impact upon terrain being confined within specific areas. Linear erosional forms are indicative of subglacial bed states in which rapid but spatially restricted basal ice movement and/or meltwater channelling has occurred.

Terrains of regional areal erosion typically exhibit low relief amplitude, limited sediment deposition and have a moulded and scoured appearance. Geological structure has often been partially exhumed in these terrains and irregular depressions and small roches moutonnées occur. These terrains, known as 'knock and lochan' in north-west Scotland, are typical of this form of GLACIAL EROSION. Similar landscapes exist in shield terrains in Canada (Plate 132a) and Fennoscandia, along the edges of the Greenland and Antarctic Ice Sheets, in Patagonia, and South Island, New Zealand. The form of this regional erosional landscape appears related to relatively slow-moving ice masses under polar-bed conditions with limited debris present. A range of wear processes operate across bedrock surfaces where occasional protuberances lead to ice pressure melting and meltwater discharges often under high pressure. Within this landscape, at a lower scale, crag and tails, and roches moutonnées are prevalent. Typically, P-forms and associated forms related to rapid subglacial meltwater flow are common.

Beneath specific zones within ice sheets, regional linear erosion appears to occur probably as a consequence of ice streaming and fast-moving, but spatially restricted, basal ice. Under these conditions, major linear forms of glacial erosion develop, such as incised bedrock troughs, fjords and tunnel valleys. At meso and microscale fluted bedrock, striae and grooves are witness to this style of selective erosion.

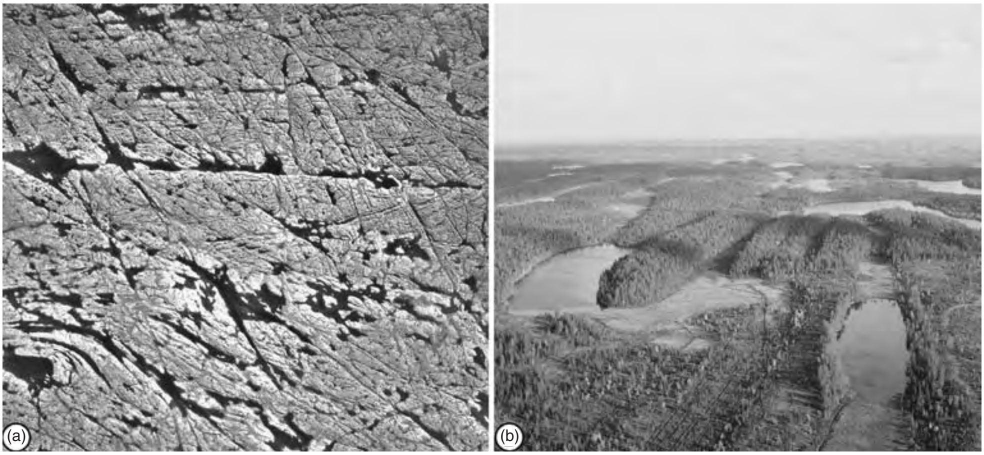

Plate 132 (a) Areally scoured glaciated landscape of the Canadian Shield near La Troie, Quebec (photo courtesy Government of Canada); (b) subglacial depositional landscape of drumlins of the Kuusamo drumlin field, Finland (photo courtesy of R. Aario)

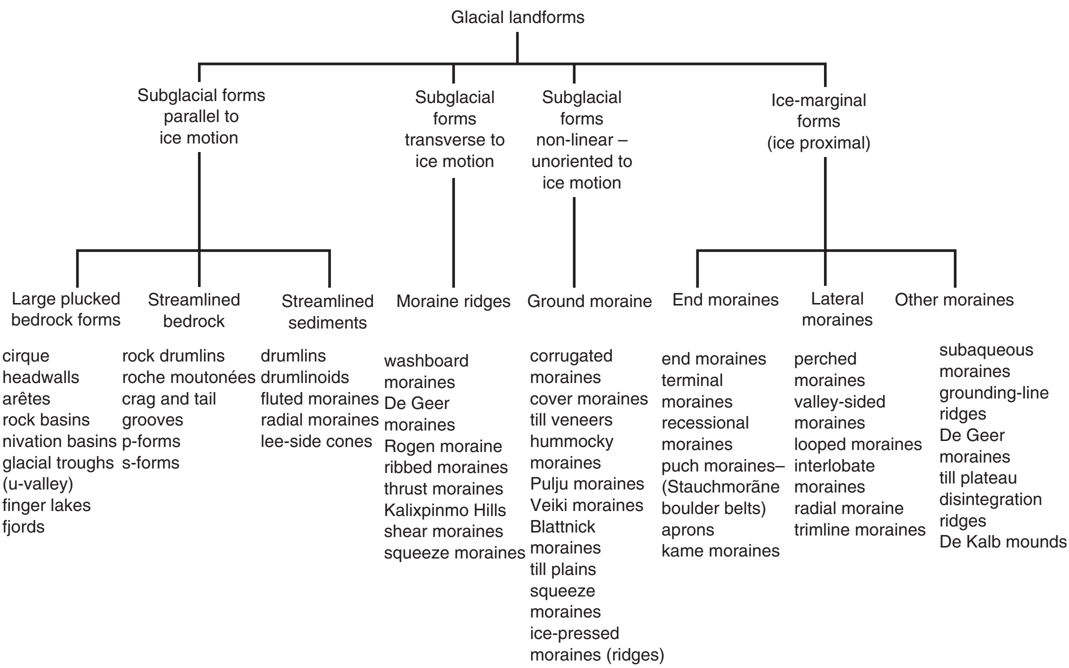

Figure 162 Subglacial landform types

Subglacial depositional landforms

Subglacial landforms can be subdivided into those developed under (1) active ice flow (advance and active retreat phases); and (2) passive or 'dying' ice flow (retreat phase). It is apparent that at least one group of landforms, previously considered as separate entities, may be related to each other as variant bedforms in a continuum of landforms at the subglacial interface. Thus drumlins, fluted moraine and Rogen moraine may be a spectrum of forms subglacially developed that evolve as a function of thermal, topgraphic, glaciological and rheological conditions (Plate 132b).

Suites of landforms can be attributed to active ice and to indirect or passive ice action (Figure 162). Most of the forms are constructional under active and/or indirect passive ice action while others may be partially erosional following ice retreat and the influence of subaerial processes. The location and type of subglacial landforms reflect the complex patterns of processes beneath the ice in differing locations under varying glacidynamic conditions (Figure 161b).

Subglacial geomorphology is a reflection of a myriad of erosional, transportational and depositional processes that produce a complex and varied glaciated landscape within which the influence of underlying topography may be muted due to the effects of deposition or sharpened due to erosional processes. The uniqueness, for example, of fjord landscapes or the gently rolling plains of the Prairies in North America stand as remarkable testaments to the impact on the Earth's surface of subglacial geomorphology.

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JOHN MENZIES

SUBMARINE LANDSLIDE GEOMORPHOLOGY

The term submarine landslide is often used as a generic descriptor for an erosive and/or depositional feature, of sediment or rock, preserved on the seafloor, or below it (if ancient) after a MASS MOVEMENT. Less generic definitions of seafloor failures such as DEBRIS FLOWS, avalanches, spreads, and the like suggest a rheology of the material as it fails. This is often hard to discern when only a scar is left as evidence of erosion, and as evidence of deposition only the rare sediment or debris pile that can be discerned by remote sensing techniques. An excellent review of submarine landslides (Plate 133) can be found in Hampton *et al.* (1996).

Detection

Most submarine landslides are old and may take place unnoticed if contemporary. Therefore, evidence of an event is obtained by remote sensing techniques such as seismic reflection, side-scan sonar and multibeam bathymetry. Data are then analysed using a seismic interpretation package that can keep track of individual horizons, or assembled in a Geographic Information System (GIS) database.

Remote sensing techniques used in the ocean employ the reflection of acoustic waves to measure the depth and physical properties of material on the surface and in the subsurface. In general, higher frequency acoustic energy yields higher resolution returns, but the energy will not propagate as deep. A 12 kHz (kilohertz) signal is used to measure water depth, and has very little subseafloor penetration, whereas a 3.5 kHz signal can penetrate over 100 m in soft sediment, and a high power, low frequency airgun source can penetrate many kilometres below the seafloor. Once a potential feature is identified, groundtruthing and geotechnical analysis of the material can be done using a variety of coring devices.

Types

Submarine mass movements have many characteristics similar to subaerial landslides, and much of the terminology is the same. Submarine failures tend to be much larger, and often better preserved over the timescale of tens of thousands of years due to slow diffusion processes in deepwater environments. Despite their size and preservation, submarine failures are hard to detect, and there is seldom a witness to describe the dynamics of the slide as it occurs. Furthermore, there is a tendency for submarine sediment failures to disintegrate, leaving a deposit that is difficult to discern on the surface. We are left to make inferences based on the resulting morphology of the erosive and depositional features, and physical properties of the surrounding material.

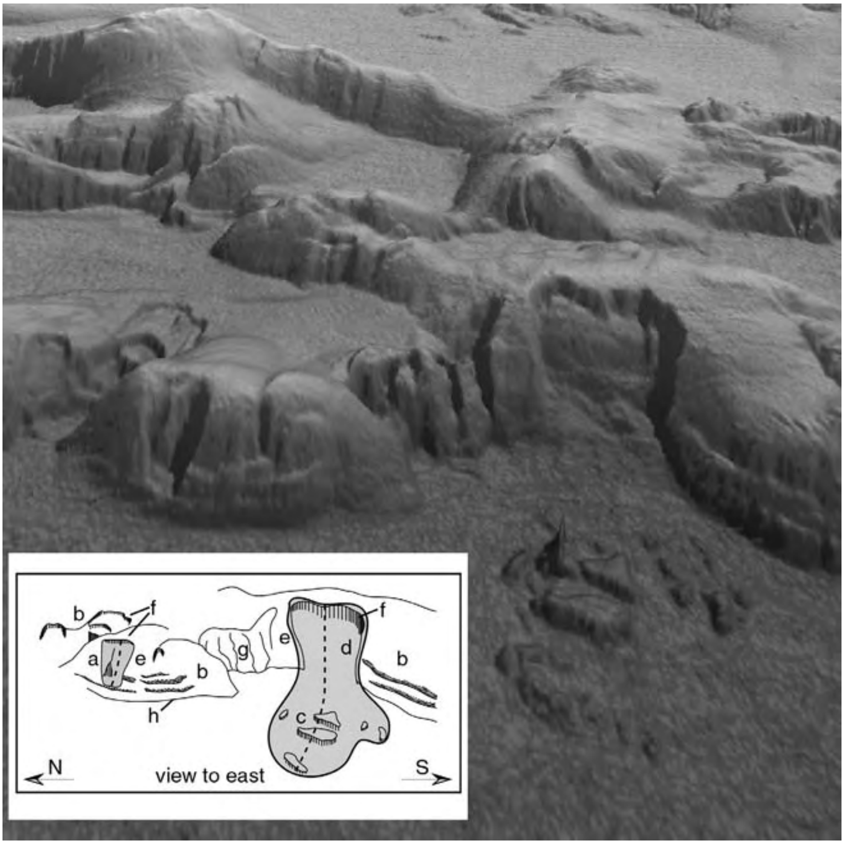

Plate 133 Submarine landslides imaged using multibeam bathymetric data from the base of the convergent continental margin, offshore Oregon, USA (modified from McAdoo *et al*. 2000). Approximately 3× vertical exaggeration, view is looking east with north to the right. The field of view at the base of the slope is approximately 30 km. The letters in the inset line drawing are referred to below. Disintegrative slides (a) have characteristic scar morphology, but lack bathymetric evidence of failed material at the base of the slide. Blocky slides have cohesive blocks of material (c) that can be related to a particular slide scar (d). The unfailed slope adjacent to the failure scar (e) is used as a proxy for the pre-failure slope conditions. Regions of the slope where material has clearly been eroded (g) are difficult to classify as a landslide as they lack characteristic evidence of discrete slide events such as a distinct headscarp (f). A series of curious terraces (h) rim the very base of the slope, and often coincide with the lowermost terminus of the slides, resulting in a 'Hanging Slide' (a). Notice the very steep seafloor with little visible evidence of erosion between slides on the westernmost ridge (closest) and on the other background ridges (b)

Failed material can either remain cohesive (a 'blocky' slide) or disintegrate into a flow. *Mass movements* (translational and rotational slides and slumps) move downslope under the force due to gravity and other body forces. The physics of *mass flows* (TURBIDITY CURRENTs, avalanches, debris flows, etc.) is governed by tractive stresses associated with a fluid.

A displaced mass of sediment or rock originates from and moves along a *rupture surface* (failure or slide surface), and portions of that failed mass may remain in contact with the rupture surface. If the rupture surface is curved (concave upward), the displaced mass rotates, and the failure is often termed a *rotational* slide or slump. If the failure plane is flat, say a bedding plane, then the slide is described as *translational*. As a block of failed material moves downslope it may disintegrate and become either a debris flow (large clasts supported by a fine-grained matrix) or *turbidity* *current* (turbulent flow of fluid supported sediment; Morgenstern 1967). LIQUEFACTION occurs when repacking of sediment grains causes a buildup of fluid pressure to the point where individual grains lose contact with each other, reducing the material's shear strength to nil. This often occurs during earthquakes, and mass flows can occur on very low inclines.

Mechanics

Failure occurs when the downslope-oriented shear stress exceeds the shear strength of the slope material. As stresses for any given environment are likely to be similar (with the possible exception of earthquakes), there is a tendency to focus on the strength side of the equation, which can range from unconsolidated sediment to well-lithified igneous rock.

The strength can be defined by the effective cohesive strength (C_0), the stress acting normal to the failure surface (σ_n), the PORE-WATER PRESSURE (p_f) and the angle of internal friction (ϕ)

$$\tau_f = C_0 + (\sigma_n - p_f) \tan \phi$$

Many submarine landslides occur on slopes well below the angle of repose (see REPOSE, ANGLE OF), therefore require a mechanism to reduce the strength of the sediment at the locus of failure. For a material with a given C_0 and ϕ, the reduction in strength can come from either an increase in p_f or a reduction in σ_n. Sediment strength is not affected by water depth, as hydrostatic conditions prevail to the depths of most slides. Therefore, rock or sediment on the seafloor will not fail on slopes less than ϕ without a transient of some sort.

There are several processes that increase p_f which in turn reduces the EFFECTIVE STRESS ($\sigma_n - p_f$), hence reducing the strength of the material. Gas hydrates (a methane-ice compound found in the sub-seafloor of many continental margins) dissociate with a fall in sea level (reduction of pressure) or an increase in temperature, and release bubble-phase methane which can get trapped, leading to an increase in p_f. Large failures that occurred during sea-level lowstand and may be related to hydrate dissociation include the Storegga slide offshore Norway, where 5,000 km^3 of material failed 8,000 years ago (Bouriak *et al*. 2000), and numerous slides in the Beaufort Sea (Kayen and Lee 1991). Earthquake generated seismic waves propagating through sediment can also increase pore-water pressures as the waves pass rapidly enough to prevent pressure dissipation (Lee and Edwards 1986). Accelerations from earthquakes can cyclically reduce and increase σ_n, therefore the resulting increase in p_f (which is frequency dependent) may be as important as a 'critical acceleration' required for failure. In shallow water, storm waves may also generate increased pore pressures due to cyclic loading, and failure can be initiated in zones of increased shear stress halfway between the crests and troughs of the waves (Seed and Rahman 1978). These failures will occur in shallow water (continental shelf and upper slope), and should not be deep-seated.

Post-failure evolution

Depending on the stresses present, sediment properties, and slope morphology, a failed mass can either stop a short distance from the source, or travel very long distances. One method of determining the likelihood of a sediment failure to disintegrate or remain as cohesive packages under cyclic loading is to consider the pre-failure water content and steady-state effective stress (see Hampton *et al*. 1996). For material with a high water content and effective stress, pore pressure is likely to increase during a transient loading event, and the effective stress will decrease, resulting in a tendency towards disintegrative failures ('contractive' sediment, as the fabric tends to collapse). In contrast, sediment that already has a high effective stress and lower water content, cohesive failures are more likely to occur as the sediment dilates, reducing the effective stress.

The post-failure dynamics are a complex combination of viscous and plastic elements. Norem *et al*. (1990) proposed an approach to modelling submarine landslides with a combination of these plastic and viscous terms:

$$\tau = \tau_c + \sigma_n\ [1 - (p_f/\rho_s H)] \tan \phi + m\,\rho\ (dv/dy)^r$$

where τ is the total shear strength mobilized during the flow, τ_c is the yield strength (similar to C_0 in the static case), $p_f/\rho_s H$ is the ratio of pore pressure to sediment density (ρ_s) and failure thickness (H), with an additional viscous term where m is the Bingham viscosity, ρ is the fluid density and a vertical velocity gradient term $(dv/dy)^r$ where r ranges between 1 for macroviscous and 2 for inertial flows. This formulation may be able to predict the final shape of a failure deposit, but does not explain the incredible mobility of slides such as the Nuuanu slide off Hawaii, where individual blocks travelled hundreds of kilometres. A hydroplaning model, where the failed material traps a layer of water above the rupture surface, helps explain the mobility of some failures (Mohrig *et al.* 1998)

Location

Submarine landslides are ubiquitous throughout the oceans, occurring on ACTIVE MARGINs and PASSIVE MARGINs, continental shelves and slopes, in the heads of FJORDs and near active deltas, on the seaward side of barrier reefs (see BARRIER AND BARRIER ISLAND), and on the flanks of mid-ocean ridges, seamounts, GUYOTs, or volcanic islands. Triggers of offshore landslides include earthquakes, large storms, SEA LEVEL change, rapid sedimentation, gas hydrate dissociation, or channel erosion in SUBMARINE VALLEYs and CANYONs among other things. Listed below are environments particularly susceptible to failure.

RIVER DELTAS

As regions of rapid sedimentation and erosion in canyons and channels, RIVER DELTAs offer zones of steep slopes and high fluid pressure that facilitate failure. When deposited on the seafloor, fine-grained sediment often has high water content yet low permeabilities, which can lead to fluid overpressure and underconsolidation. When deposited, organic-rich sediment decays producing bubble phase hydrocarbons, which subsequently aids overpressure.

The steady accumulation of fluid overpressures, and the increasing slopes are not sufficient to generate a noticeable failure – a transient such as a storm or earthquake is a necessary trigger. In 1969, large swells from Hurricane Camille in the Gulf of Mexico cyclically loaded the seafloor sediment yielding a build-up of fluid pressure that caused a failure in 100 m of water. In 1980, a magnitude 7.0 earthquake caused liquefaction of overpressured sediment on a 0.25-degree slope in the Klamath River delta, offshore northern California.

FJORDS

The offshore portion of deltas at the heads of FJORDs is a region highly susceptible to landslides. The high, often organic-rich, sediment loads of streams that drain glaciers can create elevated fluid pressures in the delta due to rapid sedimentation and decomposition that creates gas. The combination of fine-grained rock flour and coarse sediment, fluid overpressuring, and the steep nature of fjords make for a highly susceptible environment for failure during earthquakes, which are common in isostatically rebounding or tectonically uplifting regions where glaciers are likely.

The very nature of a fjord increases the landslide hazard. Fjords are an attractive place for human settlement as they offer good deepwater ports, flat land, and fresh water supply in mountainous regions close to the sea. Despite the fact that the Port of Valdez, Alaska (located at the head of a fjord) was destroyed following a submarine landslide that retrogressed onto land following an earthquake, the Trans-Alaska pipeline (which is responsible for bringing all of Alaska's North Slope crude oil to market) ends there. Were a similar event to have happened twenty years later, an unprecedented ecological disaster may well have occurred. To add to the hazard, the funnel-shaped geometry of fjords helps to focus tsunami waves. In an extreme example, a subaerial landslide at the head of a fjord in Lituya Bay, Alaska in 1958 triggered a 525 m high wave witnessed only by a few lucky boaters.

CANYONS

Submarine canyons are the primary transport avenues for sediment moving from land to the deep water. As such, these dynamic environments tend to have significant landslides. Canyon heads sequester the sediment that moves in littoral currents on the shelf. Following the 1989 Loma Prieta earthquake in central California, a failure occurred in the head of Monterey Canyon. These canyon-head failures likely fluidize (see FLUIDIZATION) into erosive turbidity currents that eroded the canyon floor, and are deposited on flat seafloor as turbidites. Successive incision events can eventually lead to large, slope-clearing landslides on the canyon walls (Densmore *et al.* 1997).

Regularly spaced canyons that occur on continental slopes but do not breach the shelf break, and are isolated from downslope erosive flows, are termed 'headless' (Orange and Breen 1992; see also GROUND WATER). Elevated seepage forces generated by the topography focusing groundwater flow to the seafloor can assist in failure, especially during transient events such as earthquakes (McAdoo *et al.* 1997). These failures are most likely small, but are a key factor in shaping the seafloor landscape.

CONTINENTAL SLOPE

Submarine landslides occur on continental slopes regardless of tectonic activity, latitude or sedimentation histories, however the temporal frequency of occurrence may vary substantially. The continental slopes are the regions in the ocean where the steepest sedimented slopes tend to exist, therefore it is a likely locus for landslide activity. The slope gradients, however, rarely approach the *angle of repose*, except in *canyons*, where *overconsolidated* (well-*lithified)* material is exposed at the seafloor, therefore either an increase in pore pressure or a significant transient is required for failure.

Many consider earthquakes the de facto trigger for slides on continental margins, even on passive margins where seismic events are less common. Curiously, one of the better concrete examples of an earthquake-triggered landslide occurred after the magnitude 7.2 Grand Banks earthquake in 1929 offshore from Canada's passive North Atlantic margin, and there have been many large earthquakes in regions with little evidence of substantial landsliding.

Landslides on open continental slopes may be associated with changes in sea level. Gas hydrate dissociation occurs during times of falling sea level, and may provide enough free gas to reduce the strength (increasing p_f) of the overburden, causing failure. Sedimentation rates are likely to be higher on the continental slope when sea level is low. Rivers bypass the sediment sinks in estuaries (see ESTUARY)and on the CONTINENTAL SHELF, and deposit material directly on the continental slope and in deep water by way of canyons. Rapid sedimentation not only increases pore pressures within the sediment, but provides a source of material to fail. Layers of finer grained mechanically weaker material deposited during highstands may provide a sliding surface following low stand sedimentation events. Continental slopes may well enter an equilibrium (see EQUILIBRIUM SLOPE) condition not long after a lowstand has provided material on the slope for which to fail.

VOLCANIC ISLANDS

Some of the largest and potentially most devastating submarine landslides occur on the flanks of volcanic islands and the mid-ocean ridges. Steep slopes are made of often highly fractured, hydrothermally altered and thermally stressed material. Earthquakes, rapid surface changes and rapidly expanding gases come with movement of subsurface magma. A block from prehistoric Nuuanu slide off Oahu, Hawaii is so large (30 km long, 17 km wide and 1.8 km thick) and far from its source (almost 100 km) it was initially identified as a seamount. A slide such as this may have been responsible for a tsunami that deposited material hundreds of metres high on the Hawaiian island of Lanai. Another slide on the Canary Islands in the Atlantic Ocean covered an area of 2,600 km^2 and 150 km^3 of material. It has been speculated that the tsunami from this slide was responsible for depositing a car-sized boulder some 20 m above sea level on Eleuthera Island, Bahamas on the other side of the ocean.

Hazard

Most submarine landslides are small, and are not likely to be noticed. The *hazard* associated with offshore slides is of primary concern to structures such as oil drilling rigs and production facilities, seafloor pipelines and cables. In 1969, Hurricane Camille in the Gulf of Mexico caused a drilling platform offshore from Louisiana to be buried in mud and moved 30 m downslope. Some slides begin in the offshore and can retrogress headward towards land. Following the 1964 Alaska 'Good Friday' earthquake, the Port of Valdez waterfront slumped into the ocean, killing thirty people.

TSUNAMI

Another significant *hazard* that may result from submarine landslides is TSUNAMI. There are two features that are unique to submarine landslide-generated tsunami. First, if the landslide was triggered by an earthquake, the tsunami generated tends to be larger than one might expect for the earthquake alone. Second, whereas earthquake-generated tsunami tend to affect the very large areas, and even cause significant damage throughout the ocean basin, landslide-generated tsunami tend to have more localized affects. On 17 July 1998 in Papua New Guinea, a tsunami with waves up to 15 m high followed a magnitude 7.0 earthquake, killing over 3,000 people. This was an unusually large tsunami given the size of the earthquake, and the region of devastation was limited to a small (~40 km) stretch of coast. Detailed multibeam and seismic surveys of the offshore revealed a large *amphitheatre* where a landslide likely triggered by the earthquake was initiated. This landslide probably contributed significantly to the amplitude of the tsunami.

The magnitude 7.2 earthquake offshore from Canada in 1929, triggered a large submarine landslide. This landslide quickly transformed into turbidity currents that travelled over 700 km into the abyssal plains, severing trans-Atlantic cables. Based on the timing of the cable breaks, the turbidity current velocity was calculated to be 55 $\mathrm{km\,hr^{-1}}$. The seafloor displacement from the landslide caused a tsunami that hit Newfoundland's Burin Peninsula, killing twenty-eight people. Canada's worst earthquake-related disaster occurred on a passive margin because of a tsunami set off by a submarine landslide.

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BRIAN G. McADOO

SUBMARINE VALLEY

Submarine valleys (canyons), sometimes tightly linked to fluvial systems, represent one of the main components that contribute to modelling the sea-bottom. They are one of the main ways for sediments to move from coastal zones to the deep oceanic basins, strongly cutting long bands of submarine scarp. Submarine valleys, localized on the main continental margins, show different features and evolutionary patterns according to whether they have a subaerial or submarine origin (De Pippo *et al.* 1999).

The characters and forms of some valleys confirm the hypothesis of their prevalently erosive origin as continental valleys. This hypothesis, even if it is acceptable for important palaeovalleys now located in shallow waters and filled with recent sediments, can hardly explain, except for particular cases, the characters and the evolution of active canyons. In fact, the bottom of most submarine valleys is a thousand metres deep, below the lowest depth (−200 m) reached by fluvial erosion during the maximum lowstand of sea level in the last glacial.

It is possible to distinguish the origin of submarine valleys on the base of certain main factors, such as eustatism and tectonics, that control their genesis and evolution. In particular the presence of valleys with a subaerial origin is more frequent in the less steep and deep areas that have experienced important sea-level changes, with the partial emersion of the continental shelf. Deep valleys, instead, are cut on steep continental margins where a big sedimentary supply activates big turbiditic flows that cut the trace of the primordial canyon.

In elevated depth areas, where the contrast between the sea waters and the waters filled with solutes and suspended sediments ceases, the mixture of water and sediment begins to slide offshore, also favoured by the shelf slope. The high sediment supply allows the erosion of the sediments previously deposited in the channels during the periods of insufficient solid supply by moving the sediments in the delta, from which turbiditic flows can originate following tectonic and/or seismic events. These flows are able to strongly erode the substratum, producing a deep incision on the shelf, where a valley could be already present. In shallow waters, instead, where the continental shelf temporarily emerges as a consequence of eustatic sea-level changes, the river directly affects the substratum, originating a submarine valley.

The tectonic characteristics of the area in which a canyon is present play a fundamental role in its genesis and evolution through time. At the same time the ampleness and the inclination of the shelf and continental slope, with the sedimentary mechanisms that regulate the transfer of the sediments from the shelf towards deeper basins, influence the evolutionary pattern of submarine valleys. In fact in basins fed by important fluvial systems, the sedimentary circulation is much more active than in areas where the only supply source is represented by gravitational processes. These processes are influenced, as well, by the eustatic sea-level changes, becoming more frequent in the lowstand periods than in the highstand ones.

The different morphologies observed in the canyons are tightly linked to different genesis and sedimentary processes. The cross section is one of the main morphological factors to be considered in the analysis of a canyon: the shape gives decisive information to define canyon evolution. A submarine valley with a V-shaped cross section generally points to dissection because of very speedy flows, while a U-shaped section points to the presence of slow and sporadic flows or to a relatively recent formation of the valley. The prevalence of depositional over erosional processes is another factor that contributes to canyon evolution by favouring the formation of a U-shaped, or at least, of a flared cross section. This causes the filling of the submarine valley bottom, with the consequent evolution of the canyon from a V towards a U-shape section. On the other hand, it is more frequent that the canyon presents steep walls and therefore a V-shaped section when erosion prevails over deposition. Slow phenomena, such as subsidence, can contribute to the evolution of the submarine valley cross section. In fact when canyons lie on slightly subsident basins they show a flared shape, while in more tectonically active areas, valleys frequently show more V-shaped sections.

The erosive processes in a canyon are strongly influenced by lithology. Canyons cut on very resistant bottom rocks, will tend in time to fill rather than become deeper. That happens in submarine valleys located along forearc systems, where the valley bottom shows a high resistance to erosion with a consequent widening of the valley section.

Another important aspect to be taken into account for the analysis of a canyon is its longitudinal profile. It can generally be pseudorectilinear to meandering with a high sinuosity index. In the areas that are not excessively steep that have been affected by tectonic events, but which are now stable, rectilinear canyons prevalently develop. The meandering ones, instead, generally develop in tectonically active areas and with more elevated gradients. The meandering of a canyon is function of its own form and longitudinal development. In fact the bending ray of a canyon and the meander wavelength increase both with width and inclination of the longitudinal profile. The presence of deep meandering channels points out the existence of transport processes characterized by frequent, continuous and slow turbiditic flows tightly connected to the inclination of the valley. In particular the local increment of canyon slope gradient is compensated by the increment of the profile sinuosity to maintain constant the solid supply.

It is possible to establish a relationship between the increment of slope gradient and the increment of the canyon sinuosity by the analysis of morphological parameters. Therefore the rectilinear or meandering profile of a submarine valley is mainly correlated both to the slope gradient value and to transport processes, and also to the nature of the material that flows inside.

The very rectilinear profile of a canyon can indicate that it has been cut on an important tectonic feature. Nevertheless valleys with a good rectilinear layout can change after tectonic events, because of continental shelf gradient variations, or also after eustatic sea-level changes.

In particular slope gradient variation owing to tectonic, tilting and subsidence processes, can cause the valley to evolve from a rectilinear towards a meandering profile. The evolution of a single submarine valley can also be strongly influenced both by the convergence and the feeding of the numerous canyons existing on the continental slope, causing consequent valley deepening and widening.

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SEE ALSO: canyon; rejuvenation; sea level

TOMMASO DE PIPPO

SUBMERGED FOREST

Submerged forests are former land surfaces on which *in situ* rooted tree stumps and associated organic deposits are found in the intertidal zone and on the continental shelves. They have been described by Geikie (1885), Reid (1913), Godwin (1943), Heyworth (1978) and Tooley (1979) in north-west Europe and by Krishnan (1982) and Mascarenhas (1997) in India.

Submerged forests owe their existence to a variety of causes: sea-level rise, coastal erosion, SUBSIDENCE and land uplift. The submerged forest at Rossall Beach on the Lancashire coast, United Kingdom, is the basal organic deposit of a kettle hole, the ramparts of which have been removed by erosion. Whereas at Hartlepool in north-east England (Tooley 1979) the submerged forest, which contains a Neolithic age human skeleton, is associated with estuarine clays. At Formby on the Lancashire coast the submerged forest formed in palaeo-dune slacks (Tooley 1979) and associated with them are human and animal footprints of Neolithic to Bronze Age (Huddart *et al.* 1999).

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MICHAEL TOOLEY

SUBSIDENCE

Ground subsidence occurs only in specific environments, where any of three distinctive processes can occur. Compaction of porous and deformable clay, peat or silt causes surface subsidence as the soils restructure with declining pore space, normally accompanied by abstraction or expulsion of interstitial ground water. Collapse or deformation of the ground into natural caves occurs mainly in limestone, gypsum and basalt, and comparable dissolution occurs on salt. Large-scale processes include crustal sag, delta compaction, earthquake movements and volcanic deflation. Mining subsidence and the collapse of old mines are entirely artificial processes that are not a part of geomorphology.

Clays are deformable because their water is loosely bonded to the chemically complex clay mineral particles. They may therefore compact (and cause subsidence) when the water is squeezed out under load or in response to water abstraction. Clay subsidence may be regional or localized, but this ranks as the most widespread and most destructive subsidence process. Entire cities have subsided, with Venice (Plate 134), Mexico City, Tokyo, Shanghai and Bangkok among the better known examples.

The natural process of clay compaction causes water to be squeezed out during slow consolidation. Artificial removal of the water, by abstraction pumping, induces more rapid compaction. The subsidence hazard lies in alluvial sequences with alternating beds of poorly consolidated sand and clay beneath large cities. Convenient water supplies are pumped from the incompressible

Plate 134 Boardwalks across the Piazza San Marco in Venice, which has now subsided below the level of most winter high tides

sand aquifers, and the overall decline in pore-water pressures causes the adjacent clays to compact. The amount of subsidence is proportional to the groundwater head decline (Poland 1984), but greater subsidence occurs on younger clays that are less consolidated by self-weight, and also on those with higher contents of unstable smectite. This clay mineral forms primarily by weathering of volcanic rocks in wet tropical environments, and is a factor behind the major subsidence of Mexico City.

The role of pressure decline within over-pumped aquifers is now recognized worldwide, and is clearly seen in the subsidence record of Venice. Long-term subsidence of the entire delta region combines with rising sea levels to create a continuing problem at Venice, but subsidence was greatly accelerated when groundwater abstraction caused clay compaction in response to head decline (Figure 163). Pumping controls have allowed head recovery, but 90 per cent of the induced compaction is irreversible, and the unabated natural subsidence continues to cause increasingly frequent flooding of the city.

Groundwater withdrawals can be matched by extraction of petroleum, natural gas and steam as causes of major ground subsidence, where the loss of hydraulic support may also affect rocks other than clay. Oilfield subsidence includes the classic case at Wilmington, USA, which caused the Long Beach harbour area of Los Angeles to subside by nearly 9 m.

Shrinkage due to water loss causes subsidence in any clay at shallow depths. Climatic changes

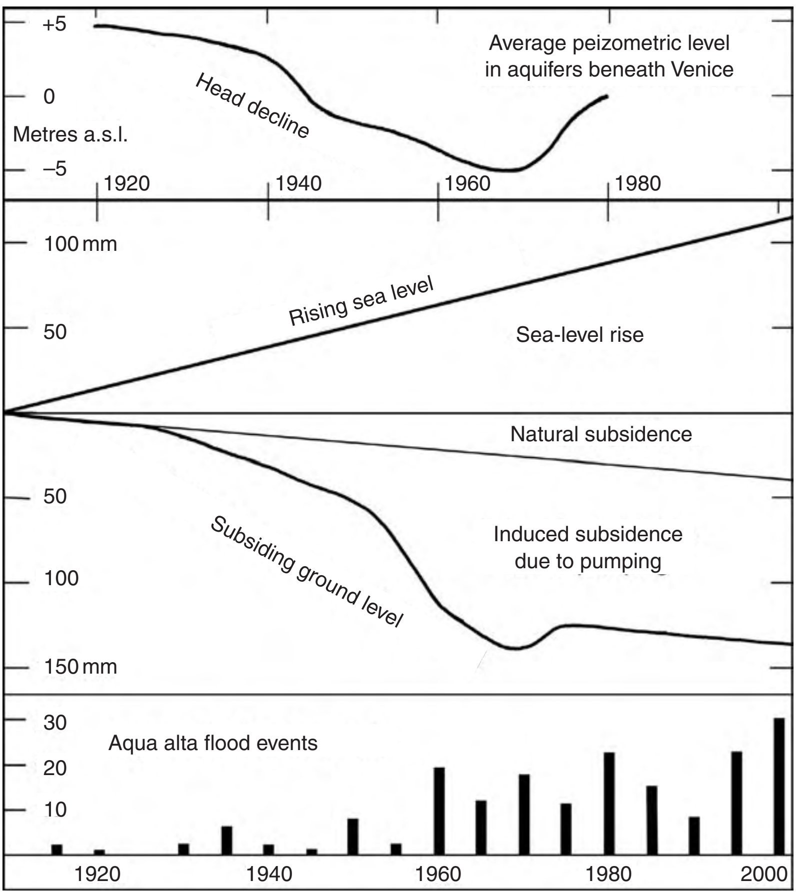

Figure 163 Causes and effects of the subsidence of Venice over the last 100 years. The bar graph shows, for each 5-year period, the numbers of flood events when high tides reach more than 600 mm above the level that initiates flooding in the lowest part of Piazza San Marco

affect large areas. Britain's dry summers since 1976 have produced a wave of subsidence damage to older houses on foundations so shallow that clay beneath them suffered first-time shrinkage in the new regime of drier climates. Subsidence is greater when trees extract excessive amounts of soil water during times of drought.

Subsidence on organic peat soils is induced by loading and/or drainage, and is similar to clay except that the compressibility of peat is far greater. Drainage causes immediate subsidence of peat, by about half the head decline, as recorded by the Holme Post in the English Fens (Hutchinson 1980). Through multiple phases of land drainage, the Post has emerged from the subsiding peat because it is founded on the stable clay beneath. After the first major movements, subsidence was about a quarter of head decline in each subsequent phase of drainage. Following the initial drainage and compaction, subsidence on peat continues due to wastage. This is loss to the atmosphere of the peat left above the water table and thereby exposed to microbial oxidation, causing annual surface lowering of 5–100 mm. In the drained English Fens, rivers have to be maintained between high banks across the subsidence bowls, land drainage water has to be pumped up into the elevated rivers, and buildings on piles progressively rise from the peat.

After clay, limestone KARST provides the world's most widespread subsidence environment, wherever it is at outcrop or beneath a soil cover. Caves constitute a subsidence hazard by threatening total ground collapse over small areas. The roof of a cave develops a natural arch in compression, which is stable when it is thinned down to about a quarter of its span (by surface lowering and roof stoping). Cave roofs need to be about double this thickness to take loads imposed by engineering, and variable degrees of fracturing and fissuring make cave roof stability difficult to assess. Also, locations of unseen caves are generally unpredictable, so cave collapse is a random and potentially destructive subsidence hazard. Most of the world's natural caves are in limestone, while gypsum and basalt are the other significantly cavernous rocks. However, collapses of rock over caves are very rare events; almost none of the world's limestone gorges originated as a collapsed cave.

A more widespread hazard is the development of subsidence dolines in soil covers that are washed into caves or fissures in underlying limestone. There are various types of DOLINEs (alternatively known as sinkholes), each distinguished by its processes of erosion and collapse. Suffosion and dropout dolines (known collectively as subsidence dolines) form entirely within the soil profile (Plate 135); infiltrating rainfall washes soil down into pre-existing rockhead fissures at rates which can be significant to engineered structures. Slow slumping of non-cohesive sandy soils produces suffosion dolines that may damage structures but are not life-threatening. In a cohesive clay soil, cavitation initiates over a rockhead fissure, and grows slowly beneath an arched soil roof. It propagates upward until the surface collapses instantly and without warning; such a 'dropout' can be a major subsidence hazard in soil-covered karst, and individual failures can be up to 100 m across.

Subsidence sinkholes are the most frequent subsidence geohazard in karst terrains, especially the rapidly formed dropouts. Sites of new failures are impossible to predict, except that buried limestone boundaries with allogenic water input are recognizable subsidence zones, notably where soil cover is 1–15 m thick (though ground collapses have occurred where the cavernous limestone lies beneath 100 m of soil cover). Dropouts are caused by water flow, and therefore occur most frequently during rainstorms, and/or where drainage paths have been modified or water tables have declined due to over-abstraction. Most doline collapses are induced by engineering activity, and are therefore avoidable (Newton 1987).

Plate 135 A dropout doline at Ripon, Yorkshire. Only the soil profile is exposed after weak cover rocks and soil failed into cavernous gypsum beneath

Rock salt is rapidly soluble in natural waters. Total removal of salt beds can develop within a few years, orders of magnitude faster than mineral losses in limestone, and ground subsidence occurs at significant rates where salt is being removed by circulating ground water. In lowland sites, including the Cheshire saltfield of England, dissolution at the rockhead leaves insoluble material in unstable residual breccias beneath the drift cover. Rates of dissolution and subsidence are a function of groundwater flow patterns, and overall natural subsidence is generally $>0.1\,\mathrm{mm\,yr^{-1}}$; during the Pleistocene, this created subsidence hollows that are now occupied by the mere lakes. All salt subsidence is vastly accelerated by any brine-pumping operations that draw in new supplies of chemically aggressive fresh water to replace the abstracted brine. Traditional wild brining targets the 'brine streams' that are zones of enhanced groundwater flow just above rockhead, and thereby causes the linear subsidences above them to deepen by $100\,\mathrm{mm\,yr^{-1}}$ or more, ultimately creating new ribbon lakes known as 'flashes'.

It is significant that all the widespread subsidence processes on clay, limestone, gypsum and salt are induced or accelerated by human activities. Subsidence over active and abandoned mines is entirely due to humans. The implication therefore stands that subsidence is a process and a geohazard that can largely be controlled. The same applies to other, less common types of subsidence.

Collapsing soils are sediments prone to internal structural collapse when water is added to them. Weak clay bonds between the grains of loosely packed, silt sediments are broken by the introduction of water; the soil then densifies by repacking under self-load or imposed load, in the process known as hydrocompaction. Potentially collapsible soils are wind-deposited LOESS and some alluvial silts that were rapidly deposited and then desiccated in large basinal fans. Hydrocompaction can promote ground subsidence of up to 5 m over wide areas in semi-arid regions where the soils have not been previously wetted.

Clays and silts are generally weak when wet and saturated, but become solid when frozen in PERMAFROST. Subsidence is then inevitable where ice-rich soils are thawed by stripping vegetation or placing warm buildings directly on the ground, though sands and gravels are largely thaw-stable.

The only subsidence that is totally natural, and therefore not controllable, is that originating with deep-seated processes. True tectonic subsidence occurs over boundary zones of subduction and also where plates are necked and thinned in zones of tension; the London area is subsiding by more than $2\,\mathrm{mm\,yr^{-1}}$ due to thinning and sinking of the North Sea Basin. Such rates of subsidence may be critical at coastal sites, especially when combined with the current rise in world sea levels (at 1–2 $\mathrm{mm\,yr^{-1}}$), largely due to global warming that has been continuous for about the last 500 years.

Major deltaic basins are sites of very slow subsidence due to a combination of factors. Within the sediment piles, soft clays are consolidated into mudstones with reduced porosities, and this compaction causes subsidence. At the same time crustal sag of the overburdened basin floor provides a second component of subsidence. Venice lies towards the margin zone of the Po Valley deltaic basin, where long-term mean subsidence is $0.4\,\mathrm{mm\,yr^{-1}}$ due to both compaction and sag; this is the natural, uncontrollable component of Venice's continued subsidence (Figure 163).

Accelerated tectonic subsidence is accompanied by earthquakes when crustal deformation is transferred to fault displacement. Large areas in Alaska subsided by up to 2.5 m during the 1964 earthquake, causing permanent drowning of coastal forests. Subsidence is more widespread as a secondary effect of earthquakes, due to liquefaction of unconsolidated sand soils. During the period of earthquake vibration, sands may temporarily behave as a liquid, so that structures subside into them. The vibrations also cause densification of loose sands by improved grain packing, and this causes permanent subsidence of the ground surface.

Entire volcanoes, and their immediate surrounds, can deflate due to migration or retreat of magma from beneath them. Parts of the Naples region subsided by 10 m due to 1,200 years of deflation of the Campi Flegri volcanic centre. This is one case where subsidence is welcome, as the reverse uplift is due to volcanic inflation that normally precedes an even more destructive eruption.

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TONY WALTHAM

SUFFOSION

An erosional process, occurring in areas where well developed KARST is overlain by unconsolidated superficial materials (usually loess or till). Suffosion is associated with PIPES AND PIPING, and is also known as ravelling, describing both catastrophic and gradual collapse of the superficial material into the bedrock cavity. The sediments slump down into widened joints and cavities in the bedrock surface, producing an irregular land surface, often exhibiting dimpling and multiple suffusion dolines (also termed shakeholes). The unconsolidated sediments can be susceptible to collapse compression upon saturation, with infiltrating water beneath the regolith creating subsoil KARREN and widened joints connected with deeper cavities. Fine materials are often eroded internally by a combination of solution and downwashing (mechanical suffosion), suggesting that these processes are related to karst underwatering. Layers of more compacted material may arch over the collapsing soil, and these are often attributed to lowering of the water table. Suffosion dolines can form within seconds and are the most widespread problem encountered in karst terrains in the construction industry, save pollution of aquifers. The depressions thereby formed typically range from 1 m diameter and 0.5 m deep, up to 100–200 m diameter and 10–50 m deep (though these are less frequent) (Ford and Williams 1989: 525).

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SEE ALSO: doline

STEVE WARD

SULA

Rocky zones along stream courses, especially in the tropics, where channels divide into anastomosing channels associated with cataracts (Zonneveld 1972). They have been explained in terms of rivers crossing an irregular weathering front, having limited amounts of abrasive material to enable incision, the induration of outcrops with ferromanganese compounds, and possible inheritance of braided conditions from Quaternary dry periods. The view that tropical rivers lack abrasive tools because of the speed and thoroughness of weathering reducing the amount and size of sediment load has been advanced by many climatic geomorphologists (e.g. Büdel 1982), but reliable data on this are sparse.

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A.S. GOUDIE

SULPHATION

A reaction between sulphur dioxide from the atmosphere and building materials (including stone) to form gypsum (calcium sulphate). It is often regarded as a feature of polluted urban atmospheres. Sulphur dioxide becomes oxidized (sometimes in the presence of catalysts such as soot or metal-rich particles) on moist surfaces to form sulphuric acid. The sulphuric acid then reacts with the stone in the following way:

$$CaCO_3 + H_2SO_4 + 2H_2O \rightarrow CaSO_4{\cdot}2H_2O + CO_2 + H_2O$$

According to Cooke and Gibbs (1993), where little moisture is present two alternative series of reactions may occur. Hydrated calcium sulphate may form first from the reaction of calcium carbonate and sulphur dioxide, as represented in the following reaction:

$$CaCO_3 + SO_2 + 2H_2O \rightarrow CaSO_4{\cdot}2H_2O + CO_2$$

Subsequently, this hydrated calcium sulphate may become oxidized to form gypsum in the presence of catalysts. Alternatively, sulphur dioxide may react with bicarbonate solutions formed by reaction with rainfall containing dissolved carbon dioxide as follows:

$$Ca(HCO_3)_2 + SO_2 \rightarrow CaSO_3 + 2CO_2 + 2H$$

The deposition of sulphates on material surfaces, often in the form of a normally dark gypsum crust, can be seen as both a consequence and a cause of weathering. This is especially true of limestones, where such a crust may be harder than the material on which it has developed. Other properties may also be different and may cause an acceleration in the speed of stone degradation. Amoroso and Fassina (1983: 264) identify three mechanisms that may account for this. First, a variation in volume occurs because gypsum has a greater volume than the quantity of calcite it replaces. One volume of calcium carbonate forms over two volumes of hydrated calcium sulphate. This causes expansive stresses in pores and cracks. Second, calcite and gypsum have different thermal expansion characteristics. The linear coefficient of the thermal expansion of gypsum is about five times that of calcite. This difference may be further increased when blackened crusts develop because they tend to absorb a larger amount of radiation than white surfaces. Third, the development of a crust will reduce the permeability of the material, which will in turn increase the water retention beneath.

Sulphation can lead to blister development and lamination, possibly through a combination of the factors just described.

Gypsum crusts in urban environments can also develop on non-carbonate rocks such as sandstone, either as a result of sulphation of calcium derived from an external source (e.g. the mortar surrounding the stone), or because of the accumulative role of lichens, algae, bacteria, etc., or because of chemical reactions with certain minerals within the rock (McKinley *et al*. 2001).

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A.S. GOUDIE

SUPRAGLACIAL

Supraglacial refers to the surface zone of glaciers and ice sheets where there is distinctive drainage, sediment sources, transport and deposition and a set of landforms characteristic of the supraglacial association and landsystem (Paul 1983). Many of the landforms produced are ephemeral and do not survive deglaciation but some do, particularly where associated with stagnant ice.

Supraglacial debris enters glacial transport from a number of sources such as mass movements from nearby mountains, tephras and from subglacial sediment carried to the surface. Minor inputs include meteorites, pollutants from anthropogenic sources and sea spray salts. Usually in high mountain glaciers where the surface ice is adjacent to valley slopes, or where isolated peaks (nunataks) protrude through the ice sheet, mass movements from the slopes are the dominant surface sediment supplier. These processes include rockfall, slides, snow and ice avalanching, debris flows, slushflows and streamflows. The supplied supraglacial debris is incorporated into the glacier by snow and ice burial in the accumulation area and by falling into crevasses and MOULINS (cylindrical vertical shafts). This is then transported by high-level, passive debris transport and suffers little modification, retaining its primary, parent material characteristics. These are angular to very angular particle shape, with common elongate to slabby particles, and coarse-grain size, with little fines.

Surface snowmelt allows downward water percolation through the snowpack and if the melting exceeds refreezing, water accumulates as slush swamps. Usually water drains down a gradient forming rills and eventually a dendritic drainage network. If discharge is high enough surface channels form, being well developed on the ablation area ice because of its low permeability. These channels are smooth-sided, offering minimal water flow resistance, hence high velocities and may follow structural weaknesses, like foliation. On warm ice the surface channels are usually short and water is diverted into the glacier via crevasses and moulins. On cold glaciers supraglacial channels commonly flow towards and alongside the margins. Supraglacial lakes can form during the early ablation season but in temperate glaciers they drain as the englacial drainage opens. In cold ice these lakes can persist and where there are large amounts of supraglacial debris, differential melting causes widespread formation of water-filled hollows. Supraglacial lakes, or ice-cauldrons, also form in depressions created by geothermal melting and subsidence.

Supraglacial debris is spatially variable and it has an important control on ablation dynamics. It acts as insulation and retards surface melting. Differential ablation causes dirt cone development (Drewry 1972) and glacier tables can form by boulders protecting a pedestal of ice from melting. Supraglacial lateral MORAINEs are ice-cored, debris accumulations at valley glacier margins, often formed by scree cone coalescence. Ablation reduction by thick debris cover means that these landforms stand above the adjacent relatively clean ice. Medial moraines are the supraglacial expression of a vertical medial debris septum which can extend to the glacier base but the surface debris is usually more concentrated and laterally extensive than that within the glacier because glacier ablation redistributes the debris. They have been classified into several types by Eyles and Rogerson (1978), including ice stream interaction medial moraines and ablation-dominated medial moraines. The former are created by lateral moraine confluence at the junction of two glaciers where the medial moraine marks the suture line between the glaciers. The second type form where ridges of englacial debris are revealed downglacier by ablation. They can form in two ways. A rockwall on a nunatak can supply debris to a small glacier area forming a linear debris plume downglacier from that point. This debris descends in the accumulation zone in the flow direction as it becomes buried. In the ablation zone it is revealed as a supraglacial medial moraine. Alternatively, folding of debris-rich, ice stratification may occur, particularly where ice flows into a restricted channel (Hambrey *et al.* 1999). The axes of these folds are usually parallel to the ice flow direction. When the debris-rich ice reaches the ablation zone, surface melting reveals the anticlinal crests of the longitudinal, debris-rich folds. If the folding is relatively open then a series of small medial moraines may define the axis of a fold.

Some glaciers have much supraglacial debris mantling the lower part of their ablation zone, particularly where mass wasting delivers large debris volumes in both the accumulation and ablation areas, as in high relief mountains, or where topography allows the movement of basal debris to the surface along shear planes. Uneven reworking and debris deposition in multiple events during ablation is responsible for a distinctive landform assemblage which can lead to topographic inversion on the glacier surface. The result is a complex sediment assemblage of faulted and folded fluvial, lacustrine and mass movement sediment (Benn and Evans 1998; Huddart 1999) which is finally deposited on the substrate. This can include supraglacial ESKERs which can be let down from supraglacial and englacial channels, ice-walled lake plains (Huddart 1983) and several KAME types. Supraglacial debris can also contribute to the GLACIMARINE environment as part of morainal banks and from iceberg rafting.

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SEE ALSO: esker; glacial deposition; ice stagnation topography; kame; moraine; moulin

DAVID HUDDART

SURGING GLACIER

Surging glaciers are often instantly recognizable from their 'looped medial moraines' or teardrop-shaped loops of debris on the glacier surface. These loops record cyclic changes in velocity between a trunk GLACIER and its tributaries. Additionally, folding, thrust faulting and severe crevassing of ice at the snout is induced by compressive flow. A glacier is said to surge when it exhibits major fluctuations in velocity over timescales that range from a few years to several centuries, swinging between phases of rapid and slow flow. The phase of rapid motion is called the 'surge' or 'active' phase during which ice is transferred from the upper part of the glacier (the reservoir area) to the snout. This results in a dramatic advance of the snout and a concomitant thinning of the reservoir area. The period of slow flow between surges is termed the 'quiescent phase' and is characterized by the build-up of ice in the reservoir area and the mass stagnation of the snout. This gradually increases the surface gradient of the glacier until the next surge is initiated. Glacier flow velocities in the surge phase can be ten times faster than those of the quiescent phase. The surge and quiescent phases are usually of constant length for each individual glacier, resulting in a periodic cycle of surges. However, cycle length varies greatly between glaciers and glaciated regions. Glacier surge velocities also vary considerably, ranging from 50 m per day on the Variegated Glacier in Alaska to maximum speeds of only 16 m per day on Svalbard. The length of the quiescent phase also varies, maximum periods of time being 50–500 years for Svalbard glaciers compared to 20–40 years for most other regions.

Geographically, surging glaciers tend to cluster in Alaska, Yukon and British Columbia in North America, Svalbard and Iceland in the North Atlantic and the Pamirs in western Asia. Further examples have, however, been identified in the Canadian high arctic, Greenland, the Caucasus, Tien Shan and Karakoram mountains in Asia and in the Andes. Several glacier characteristics in these regions have been correlated with surging behaviour, such as glacier length and overall gradient. Bedrock types also appear to be linked to surging.

Glacier surges are not triggered by climatic fluctuations but instead result from changes in the internal dynamics of the glacier system. Variations in basal sliding appear to drive the changes in glacier flow velocity in a surging glacier, and this is driven by reorganizations of the subglacial drainage system. Based upon the few intense studies undertaken on surging glaciers, rapid sliding appears to be initiated by rising water pressures at the glacier bed. This is a function of the trapping of subglacial meltwater by conduit closure due to ice creep. Although this mechanism of fast flow initiation applies well to temperate glaciers it does not explain fully surges by subpolar glaciers. A study on Trapridge Glacier in the Yukon, which is frozen to its bed at the snout, suggests that deformable sediment beneath the warmer ice further up glacier plays a significant role in surging (Clarke *et al.* 1984). A wave-like bulge develops on the glacier surface at the boundary between cold and warm-based ice. This bulge moves down the glacier mostly by subsole deformation. This is initiated by changes in the subglacial water flow through the deformable substrate.

A linkage between surging behaviour and regional climate has been suggested by Budd (1975), who indicates that the concentration of surging glaciers in specific regions, and the variability of surge velocities and phases between regions, must reflect some climatic control. Specifically, in a continuously fast-flowing glacier the annual mass balance can be discharged by normal flow velocities. In contrast, in a surging glacier the mass throughput is too great to be discharged by slow flow alone but too small to sustain fast flow over long periods. Consequently, surging glaciers build up mass slowly until fast flow is triggered. This fast flow drains and exhausts the supply from the reservoir, thereby reinitiating slow flow. This means that the velocity of a surging glacier is constantly out of equilibrium with climate.

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DAVID J.A. EVANS

SUSPENDED LOAD

The suspended load of a river comprises mineral and organic matter that is dispersed through the flow by turbulence. Typically, the mineral load is dominant and consists of grains ranging in size from clay ($<4\,\mu$m) up to sand grade (<2 mm). Often, suspended particles exist as 'flocs' or clumps of finer grade particles linked by weak chemical forces. The suspended load is quantified in terms of its concentration (mass of sediment per unit volume of water), discharge (sediment mass flux per unit time – also referred to as the 'load'), and particle-size distribution (proportions of the load in given size fractions).

The clay-silt fractions (often termed 'washload') are largely sourced from erosion processes outside the river channel; being more easily suspended, they are well mixed through the flow and travel long distances in suspension. Sand requires more intense turbulence to remain suspended, thus it tends to be concentrated near the stream bed and tends to be sourced from the stream-bed material. In cobble/boulder-bed rivers, the suspended load is also augmented by fine particles generated by abrasion of cobbles/boulders as they roll, hop and slide along the bed as part of the BEDLOAD. Downstream through a drainage basin, the suspended load increasingly dominates over the bedload.

Since the fine washload is easily suspended, its concentration generally depends only on the relative rates of supply of sediment and water to the channel (i.e. rather than being limited by the physical capacity of the stream to suspend it). Indeed, when the concentration of mud-rich washload becomes sufficiently large, the fluid properties change from those of a water flow to those of a hyperconcentrated or debris flow. While turbulence intensity does limit the suspended sand concentration in sand-bed streams, in gravel or rock-bed channels the primary limitation on suspended sand concentration may be sand availability. Thus in most rivers, the suspended load tends to be 'under-supplied' with respect to the river's potential capacity to transport it. Because of this, it cannot be determined by physics-based formula but must be measured.

Accurate suspended load measurement at a river section requires measurements of both sediment concentration and water velocity. One approach is to collect point samples of water and to make point velocity measurements at intervals over the flow depth, then plot profiles of concentration and velocity. Special 'point-samplers' are used for this purpose. These cable-suspended samplers, comprising a brass bomb and an internal sample bottle, have a solenoid-operated valve to control the water-sample capture and they are designed so that they sample isokinetically (i.e. they accept a sample at the ambient water velocity). A second, simpler approach is to use a 'depth-integrating' sampler. With this, the inlet nozzle is kept open while the sampler is traversed from the water surface to the bed and back again, so performing a mechanical integration of the concentration and velocity profile. Details about samplers and methods developed in North America can be found in Edwards and Glysson (1999).

A single suspended sediment measurement is time consuming, and where a continuous record is required it is usually obtained by collecting 'index' samples at one location and establishing a relationship between the index sample concentration and the cross-section mean concentration. Index samples may be collected manually, but in remote locations or in 'flashy' small basins they are more often collected by an automatic pumping-sampler or by measuring a surrogate property such as water turbidity. Turbidity sensors have been widely used to date (e.g. Gippel 1995). They may be either transmissivity (attenuance) or back-scattering (nephelometric) types. Since they do not sense sediment concentration directly, a further calibration relation must be established between the turbidity sensor and suspended sediment concentration. The turbidity signal depends both on sediment concentration and particle characteristics, notably particle size and shape. Light scattering is greatest from clay particles and least from sand grains, thus turbidity sensors are more sensitive to the washload. Suspended sediment concentrations at-a-site tend to increase with water discharge, and so a relatively simple way to estimate the suspended discharge over a period of time is with a SEDIMENT RATING CURVE that empirically predicts sediment concentration at a given water discharge.

Globally, average annual river suspended sediment discharges vary greatly. The primary factors causing this variation include basin area, topography, precipitation, lithology, tectonics, vegetation cover, land use, floodplain sequestration and sediment entrapment in reservoirs. On a unit area basis, the largest sediment discharges occur in tectonically active, high-rainfall, steeplands around the western Pacific basin (e.g. Taiwan and New Zealand, where sediment discharges can exceed 20,000 $\mathrm{t\,km^{-2}\,yr^{-1}}$). According to Milliman and Syvitski (1992), the largest sediment discharges from single river basins to the oceans are from the Amazon River ($1.2 \times 10^9\,\mathrm{t\,yr^{-1}}$, due to its vast area) and China's Huanghe River ($1.1 \times 10^9\,\mathrm{t\,yr^{-1}}$, due to its large areas of landuse-affected erodible loess terrane). Milliman and Meade (1983) estimated the total worldwide delivery of suspended sediment to the oceans at $13.5 \times 10^9\,\mathrm{t\,yr^{-1}}$. Milliman and Syvitski (1992) revised this figure and considered that it might have been at least $20 \times 10^9\,\mathrm{t\,yr^{-1}}$ before the proliferation of dams during the twentieth century, which have intercepted a significant fraction of the sediment discharge. They considered that global suspended sediment discharges have more than doubled as a result of widespread deforestation and farming over recent millennia.

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D. MURRAY HICKS

SYNGENETIC KARST

This refers to karst that has developed concurrently with the diagenesis and consolidation of the host karst rock. It is most well known in calcareous dune limestones, rocks which are largely of biogenic origin, being comminuted shell fragments consolidated into calcareous AEOLIANITE, although there may be significant admixtures of quartz and other non-calcareous minerals. Such rocks are usually found within 40° north and south of the Equator and karst development in them is especially well known in the coastal regions of South and West Australia (White 2000), but is also found in other regions such as the eastern and southern Mediterranean and on Caribbean islands.

The dunes developed during glacio-eustatic oscillations during the Quaternary that exposed large amounts of shell material in coastal regions. Dunes were blown inland and accumulated in a series of ridges roughly parallel to the coast. The first stage in diagenesis was fixing by vegetation, the growth of which promoted soil development and production of biogenic carbon dioxide in the root zone. This encouraged dissolution of the dominantly aragonitic sands by infiltrating rain water. Saturation with respect to calcium carbonate is achieved close to the surface, under conditions that are open system with respect to carbon dioxide. Further percolation of this saturated solution down into the sands away from the root zone results in degassing of carbon dioxide and evaporation, with the result that the water becomes supersaturated and calcite is precipitated in interstitial pores in the dune, thus cementing it. Cementation produces a surface case-hardened crust over the dune that advances progressively downwards from the surface. Vertical solution pipes with case-hardened rims often perforate the dune crust. Cementation also occurs in the water saturated (phreatic) zone, and eventually the entire dune becomes calcreted.

During this process of dune cementation streams flow from inland areas towards the coast, but are blocked in their progress by dune ridge barriers that lie parallel to the coast. The streams pond on the upstream side and their waters permeate the porous sands to emerge on the seawards side. As the dune sands become indurated, stream flow is restricted to defined routes which eventually develop into blind valleys leading into caves many of which have flat water-table controlled roofs. The cave ceilings are also case hardened. Some cave ceilings collapse and give rise to large chambers that become profusely decorated with speleothems. Further collapse develops collapse dolines.

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PAUL W. WILLIAMS

SYSTEMS IN GEOMORPHOLOGY

Students of the Earth's surface have been referring to 'systems' for nearly three centuries: but when Buffon (1749) did so, he was concerned with the theoretical structures (which we should now term cosmogonies) of the likes of Whiston, Burnet and Woodward; whereas Playfair (1802: 102) in his famous discussion of the 'system of vallies' was dealing with concrete entities. These two aspects were brought together in A.N. Strahler's enormously influential paper on the dynamic basis of geomorphology (1952). Here Strahler considers (pp. 934–935) that the 'fullest development' of geomorphology will occur 'only when the forms and processes are related in terms of dynamic systems and the transformation of mass and energy are considered as functions of time'. He goes on (p. 935): 'Many of the geomorphic processes operate in clearly defined systems that can be isolated for analysis.'

Strahler's view derived from early papers by the biologist, Ludwig von Bertalanffy, notably that published in *Science* in 1950. The move towards the use of systems as a theoretical device, unifying all sciences and extending to fields such as economics, was christened General Systems Theory (GST) by von Bertalanffy (1950). Both the theoretical and the practical aspects of GST were adopted by several of Strahler's students, most notably by the British geomorphologist R.J. Chorley (1960; Chorley and Kennedy 1971) and the American S.A. Schumm (1977). The two versions of 'system' can be seen side by side in the Twenty-third Binghamton Symposium (Phillips and Renwick 1992).

So what *are* systems? and what has been their dual role in geomorphology since 1952? According to Hall and Fagan, 'A system is a set of objects together with relationships between the objects and between their attributes' (1956: 18). They acknowledge that this is an imprecise definition, but say 'This difficulty arises from the concept we are trying to define; it simply is not amenable to complete and sharp description' (*idem*). Further, they acknowledge the existence of abstract or conceptual systems (pp. 19–20) as well as natural systems (pp. 23–24), which latter they consider as either open (exchanging 'materials, energies or information') with their surroundings; or closed (with no such imports and exports). Chorley and Kennedy (1971: 2) modified this distinction by relabelling Hall and Fagan's 'closed' systems as 'isolated' and identifying a new category of 'closed' system which exchanged energy (or, presumably, information) but not mass with its environment. In this sense, there must be very few true isolated systems studied by geomorphologists, but there may be several categories of effectively closed systems, including the planet Earth itself. Chorley and Kennedy went on (pp. 4–10) to identify four classes of system of relevance to physical geography: morphological, cascading, process-response and control. They discuss each in turn, devoting most attention to the process-response examples together with related considerations, including input and output, equilibrium, thresholds (see THRESHOLD, GEOMORPHIC), trends and simulation models. Finally, Chorley and Kennedy consider control systems, a topic taken up and elaborated by Bennett and Chorley (1978) in a book ranging far beyond geomorphology.

So how have these ideas – of both the conceptual and the natural systems – actually been employed by geomorphologists? Chorley (1960) followed Strahler's (1952) lead in emphasizing the distinction between historically focused geomorphology (explicitly, the ideas of the American colossus, W.M. Davis and, in particular, his concept of the geographical cycle, see CYCLE OF EROSION), which he termed 'closed system' thinking: and the 'open system' approach deriving from GST and the study of process. Chorley lists (1960: B8 and 9) seven advantages which will accrue from the adoption of the 'open system' view. They are:

1 Emphasis will be placed on the 'tendency to adjustment' between form and process.
2 Emphasis will be directed towards 'the essentially multivariate character' of geomorphic processes.
3 It will allow consideration to be given both to progressive changes and to stasis or abrupt changes (i.e. both concepts such as DYNAMIC EQUILIBRIUM and thresholds).
4 It will foster 'a less rigid view regarding the aims and methods of geomorphology' than (he contends) is held by those concerned with studying the historical development of landscapes and landforms.
5 Emphasis will be placed upon the whole landscape, rather than on 'the often minute elements … having supposed evolutionary significance' (by this he clearly has in mind the 1950s' obsession with terrace and PENEPLAIN remnants).
6 It will encourage 'rigorous' work in areas where there are few or no traces of erosional history.
7 There will be major implications for geography as a whole, especially by directing attention 'to the increasingly hierarchical differentiation which often takes place with time'.

More than half a century on, it is not evident that these advantages have been realized. One very important drawback to the use of 'systems thinking' was identified by the historical geographer, Langton (1972). If we accept Hall and Fagan's definition, then it is evident that we can identify or create as many systems in thought or reality as we wish. This becomes, then, fundamentally an exercise in classification. It should then be clear that we need to be able to rate the value of different classifications and this, Langton contends, can only be done if we can identify the optimum outcome from the system we define. Whilst this may be feasible, say, in evaluating the water yield from a DRAINAGE BASIN in terms of irrigation, or as a site for a hydro-power operation, it is not at all clear that a slope, a PINGO or even the drainage basin can be legitimately thought to possess a 'goal'. This problem is, it seems, a fundamental one. And very closely linked to this inherent difficulty is that of identifying unique and non-overlapping systems in the real world. This is absolutely crucial for process-response systems if their operation is to be fully understood. Consider a drainage basin. Can its physical watershed be truly and exactly fixed? Can we be certain that the watersheds for mass or energy inputs and outputs coincide with the topographic limits? (Consider the question of subsurface water and solute movement.) If our designation of discrete basins may seem accurate, can we be as content with the identification of 'hillslopes' or even 'channels'? If we do not have this certainty – so different from the identification of (say) a household 'hot water system' – then it must surely cast doubt on the validity of our conclusions about the systems' operations. As a result, by the end of the twentieth century, the use of the term 'system' in geomorphology seemed really loosely, if at all, related to the specific ideas of von Bertalanffy, Strahler and Chorley. An example would be the text by Allen (1997) which opens with a chapter entitled 'Fundamentals of the Earth surface system' (pp. 1–50) with very little in the way to indicate direct descent from – say – Chorley and Kennedy, in terms of vocabulary or illustrations.

Before 1952, it was possible to think of 'systems' at all levels, both conceptual and practical. What might be termed the 'systems boom' of the 1950s–1970s in geomorphology argued for a situation in which a particular theoretical vision would inform the whole choice of research topics as well as the vocabulary in which results would be couched. In general the link between philosophical concept and practice was never really made. As a result, we have a residual terminology which really owes more to the vernacular than to either Strahler or Chorley. And we also see far more use of systems as a research tool in precisely those applied situations where a desired outcome can be specified, thus confirming Langton's assessment.

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SEE ALSO: Cycle of Erosion; drainage basin; flow regulation systems; integrated coastal management; land system; peneplain; step-pool system; threshold, geomorphic

BARBARA A. KENNEDY

T

TAFONI

Tafoni (singular tafone) are cavernous weathering forms which typically are several cubic metres in volume and have arch-shaped entrances, concave inner walls, overhanging margins (visors) and fairly smooth gently sloping, debris-covered floors (Mellor *et al*. 1997). They occur in many parts of the world (see Goudie and Viles 1997: table 6.1), including polar regions, but may be especially well developed in coastal and dryland situations. They occur in a wide range of rock types, but especially in medium and coarse-grained granites, sandstones and limestones. Indeed it is only rocks with relatively closely spaced discontinuities (bedding planes, foliation, joints) such as shales and slates, that seem to be relatively unaffected by these cavernous weathering forms.

The cavernous hollows of tafoni are believed to result largely from flaking and granular disintegration caused by a range of possible weathering processes that include hydration, salt crystallization, and chemical attack by saline solutions. Some workers have found clear evidence of SALT WEATHERING being involved, while others have not. The role of case hardening in their foundation is also the subject of debate, but can help to explain the formation of the visor. It is also possible that tafoni develop through a positive feedback effect, in that once a hollow is initiated it creates an environment in which weathering is favoured (Smith and McAlister 1986). For a cavity to grow there needs to be a mechanism to remove flakes and spalls. Wind may play a part, as may organisms such as pack rats. Although some early workers thought that the actual excavation of a cavity might be achieved by wind abrasion, many tafoni occur in

Plate 136 A large tafoni developed in granite near Calvi in Corsica

Plate 137 A remarkably developed tafoni in volcanic rocks in the Atacama Desert near Arica in northern Chile

environments where sand blasting does not occur or they may have an aspect (i.e. the leeward side of a boulder) or a height up a cliff face that precludes such a mechanism. It is clear that in whatever way they form, tafoni grow significantly over tens of thousands of years (Norwick and Dexter 2002).

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A.S. GOUDIE

TALSAND

Large-scale infillings of ice-marginal valleys occurring in the Pleistocene lowlands of northern Germany and Poland. However, the term is also employed to represent any flat sandy region of glacifluvial and/or periglacial origin, set below the general level of Pleistocene uplands (Geest Plateau) (Schwan 1987). This is referred to as a talsand plain. Talsand is thus a geomorphological term that is largely restricted to north German Quaternary geology, translated as valley sand.

Talsand is composed of glacifluvial or periglacial sediments, though often an overlying bed of aeolian sands is present. The majority of talsand was deposited during the last glacial period (the Weichselian glacial *c.*70–10 ka years ago – the equivalent of the Devensian glacial in the UK, and the Wisconsin glacial in the USA), though formation may have initiated in the preceding late Saalian period. The overlying aeolian deposit typically originates from the late Weichselian and often continues into the Holocene. The transformation from fluvial to aeolian sediments is predominantly due to increased wind intensity and the lowering of ground water due to permafrost degradation (Schwan 1987).

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STEVE WARD

TALUS

A term of French origin, which can have different meanings. For Anglo-Saxon geomorphologists it is used as a synonym for scree. It is an accumulation of weathered rock fragments under a cliff. The mechanisms which control this accumulation are, however, complex. The weathering of the underlying rock face is most often linked to frost (see FROST AND FROST WEATHERING) or earthquake action. The transit of rock debris from the cliff to the deposit can be due to one or several processes. Consequently, the morpho-sedimentological characteristics of this deposit depend on the type of these processes and the lithology.

In rare cases the accumulation is exclusively linked to rockfall. This rockfall talus presents a steep slope, a longitudinal sorting, and does not show any stratification in cross section. In most cases transit is due to a combination of different processes like avalanches or DEBRIS FLOWS. The morpho-sedimentological characteristics of the deposit will depend on the process or the dominant processes.

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VINCENT JOMELLI

TALUVIUM

Taluvium is a slope deposit composed of rock fragments in a fine matrix, thereby bridging the gap between TALUS (composed of rock fragments) and colluvium (fine material only). Given time, talus will eventually change into taluvium, and subsequently into COLLUVIUM. The development of taluvium from talus has been related to weathering of the rock, although formation by the incorporation of finer airfall material in the talus lattice, such as by loess (Pierson 1982) and volcanic tephra, has also been suggested.

Taluvium deposits are stratigraphically, texturally and hydrologically heterogeneous, as talus formation and emplacement of the fine matrix is typically an episodic process. The accumulation of a fine matrix increases the total surface area of particle contact, increasing internal friction and encouraging aggregation. However, further accumulation of fines decreases the void ratio, impeding drainage and resulting in increasing pore-water pressures. Thus, the slope stability of taluvium (typically 25–28°) is predominantly influenced by its hydrological properties, with saturation resulting in a reduction in its stability angle to approximately half its dry value.

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STEVE WARD

TECTONIC ACTIVITY INDICES

Morphometric techniques used as reconnaissance tools to identify areas experiencing rapid tectonic deformation. Among the indices that have been found most useful are the following (Keller and Pinter 2002).

The hypsometric integral

The hypsometric curve describes the distribution of elevations across an area of land. It is created by plotting the proportion of total basin height (h/H = relative height) against the proportion of total basin area (a/A = relative area). The total height (H) is the relief within the basin (the maximum elevation minus the minimum elevation). The total surface area of the basin (A) is the sum of the areas between each pair of adjacent contour lines. The area (a) is the surface area within the basin above a given line of elevation (h). The value of relative area (a/A) always varies from 1.0 at the lowest point in the basin (where $h/H = 0.0$) to 0.0 at the highest point in the basin (where $h/H = 1.0$).

A simple way to characterize the shape of the hypsometric curve for a given drainage basin is to calculate its *hypsometric integral* (H_i). The integral is defined as the area under the hypsometric curve. One way to calculate the integral for a given curve is as follows [8,9]:

$$H_i = \frac{\text{Mean elevation} - \text{minimum elevation}}{\text{Maximum elevation} - \text{minimum elevation}}$$

A high hypsometric integral indicates a youthful topography. Digital Elevation Models (DEMs) make calculations easy (Gardner *et al.* 1990).

Drainage basin asymmetry

The Asymmetry Factor (AF) has been developed to detect tectonic tilting transverse to flow:

$$AF = 100\ (A_r/A_t)$$

where A_r is the area of the basin to the right (facing downstream of the trunk stream), and A_t is the total area of the drainage basin. AF values are sensitive to tilting perpendicular to the trend of the trunk stream and the greater the divergence of the AF values from 50 the greater is the degree of tilt.

Stream Length–Gradient Index (SL Index)

This is represented as:

$$SL = (\Delta H/\Delta L)L$$

where ΔH is the change in elevation of a stream reach and ΔL is the length of the reach. L is the total channel length from the midpoint of the reach upstream to the highest point on the channel. The index is used to identify recent tectonic activity by identifying anomalously high index values on a particular lithology.

Mountain-front sinuosity

This is an index that reflects the balance between erosional forces that tend to cut embayments into a mountain front and the tectonic forces that tend to produce a straight front coincident with an active range-bounding fault (see FAULT AND FAULT SCARP). It is expressed as:

$$S_{mf} = L_{mf}/L_s$$

where S_{mf} is the mountain-front sinuosity, L_{mf} is the length of the mountain front along the foot of the mountain at the pronounced break in slope, and L_s is the straight-line length of the mountain front. Given that mountain fronts associated with active tectonics and uplift are straight, they have low values of S_{mf}.

Ratio of valley floor width to valley height

This may be expressed as:

$$V_f = 2V_{fw} / [E_{ld} - E_{sc}) + (E_{rd} - E_{sc})]$$

Where V_f is the valley floor width-to-height ratio, V_{fw} is the width of the valley floor, E_{ld} and E_{rd} are elevations of the left and right valley divides respectively, and E_{sc} is the elevation of the valley floor. High values of V_f are associated with low uplift rates that have enabled streams to cut broad valley floors. Low values of V_f are associated with uplift.

The use of indices such as these, particularly in combination, enable the production of a relative tectonic activity class designation for an area.

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A.S. GOUDIE

TECTONIC GEOMORPHOLOGY

As Burbank and Anderson (2001: 1) remark, 'The unrelenting competition between tectonic processes that tend to build topography and surface processes that tend to tear them down represents the core of tectonic geomorphology.' Immersed in small-scale process geomorphology, geomorphologists have been remarkably slow to explore both the significance of the conceptual advances provided by the PLATE TECTONICS model and the range of new techniques and data sources relevant to the quantification of long-term denudation rates (Summerfield 2000: 3). Nonetheless, over recent years the enormous growth in topographic data (see, for example, DIGITAL ELEVATION MODEL) and in geochronologic data (see, for example, COSMOGENIC DATING; FISSION TRACK ANALYSIS) have provided new opportunities to address long-standing questions of landscape evolution at the regional and continental scales (Morisawa and Hack 1985).

Plainly there are many Earth features that owe their form in large measure to tectonic activity (see ACTIVE AND CAPABLE FAULT; FAULT AND FAULT SCARP; FOLD; MANTLE PLUME; PULL-APART AND PIGGY-BACK BASIN; RING COMPLEX OR STRUCTURE; TECTONIC ACTIVITY INDICES). Equally there are miscellaneous types of tectonic activity (see CYMATOGENY; DIASTROPHISM; EPEIROGENY; ISOSTASY; SEAFLOOR SPREADING; WILSON CYCLE). At the large scale some major landscape features are associated intimately with various types of plate boundary (see ACTIVE MARGIN; ESCARPMENT; MOUNTAIN GEOMORPHOLOGY; PASSIVE MARGIN; SEISMOTECTONIC GEOMORPHOLOGY; VOLCANO, etc.)

The scope of modern tectonic geomorphology can be appreciated by the context of the recent text by Burbank and Anderson (2001). They start with a consideration of *geomorphic markers* (features or surfaces that provide a reference frame against which to gauge differential or absolute tectonic deformation). Next they consider dating methods, which are vital to determine rates at which faults move or surfaces deform. Then they pass on to stress, faults and folds, before analysing geodetic methods for analysing short-term deformation (e.g. GPS). This is followed by a consideration of the evidence for, and consequences of, palaeoseismology. Then they look at the balance between erosion and uplift, and deformation at various timescales from the Holocene to the Late Cenozoic. They conclude with a discussion of numerical modelling of landscape evolution.

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A.S. GOUDIE

TERMITES AND TERMITARIA

Termites, insects of which there are several thousand species, are members of the Isoptera order, and about four-fifths of the known species belong to the Termitidae family (Harris 1961). They vary in size according to their species, from the large African *Macrotermes*, with a length of around 20 mm and a wingspan of 90 mm, down to the Middle Eastern *Microcerotermes* which is only about 6 mm long with a wingspan of 12 mm.

They occur in great numbers – 2.3 million ha^{-1} in Senegal and 9.1 million ha^{-1} in the Ivory Coast

(UNESCO, UNEP, FAO 1979). The vast majority of termite species are found in the tropics, though their distribution is wider than this; they extend to 45–48°N and to 45°S.

Termite mounds and hills are the most striking manifestations of termite activity, and have a large range of sizes and morphologies (Goudie 1988). The heights of termite constructions vary considerably according to species. There are records in the literature of mounds attaining heights in excess of 9 m, though most are less than this. Among the species that create the tallest mounds are *Bellicositermes bellicosus*.

In general the densities of mounds vary considerably according to both environmental conditions (e.g. soil type) and termite species. So, for example, the density of the very large mounds produced by *Macrotermes bellicosus, Macrotermes subhyalinus, Macrotermes falciger, Bellicositermes bellicosus* and *Nasutitermes triodiae* tends to be less (often around 2–10 ha^{-1}) than those for the smaller types of mound (often around 200–1,000 ha^{-1}).

Undoubtedly soil characteristics are an important control of mound formation. Mounds tend to be rare on sands (where there is insufficient binding material), on deeply cracking self-mulching clays (which are unstable), or on shallow soils (where there is a shortage of building material) (Lee and Wood 1971). Soil drainage may also be important, as are human activities. Although in most cases it is obvious that particular mounds have been produced by particular species of termites there have been some arguments about the origin of some mounds found in the tropics and elsewhere (see MIMA MOUND). For example, Cox and Gakaha (1983) have argued that certain mounds in Kenya are created by the mole-rat, *Tachyoryctes splendens*, whereas Darlington (1985) argues that the same mounds are produced by a type of termite – *Odontotermes*.

Plate 138 A large termite mound developed by *Macrotermes* in the Mopane woodland of northern Botswana

Whereas termites live in hidden subsoil chambers or in the conspicuous mounds just discussed, they have a significant effect on soils, partly because of their mechanical activities and partly because of their feeding habits. They can, for example, cause soils to become rich in calcium, play a major role in nutrient cycling, remove organic litter from soil surfaces and mobilize particular soil fractions (e.g. clay). Whether they contribute to laterite formation or lead to its degradation, however, has been the subject of debate (Runge and Lammers 2001).

Termites can contribute to accelerated rates of soil denudation. Lee and Wood (1971) identify three main ways in which termites can do this:

1 by removing the plant cover;
2 by digesting or removing organic matter which would otherwise be incorporated into the soil, and thus making the soil more susceptible to erosion;
3 by bringing to the surface fine-grained materials for subsequent wash and creep action.

The huge numbers of termites and their large total biomass in favoured localities ensures that these three mechanisms are important. The live weight biomass of termites can be comparable to the live weight biomass of large mammalian herbivores in tropical areas. However, in addition to the potential for erosion and sediment yield caused by mound formation and abandonment it is important to remember the other major consequence of termite-caused soil translocation. This is the construction of covered runways or 'sheetings'. These are constructed of soil particles cemented together with salivary secretions (Bagine 1984).

Through their effects on denudation termites may have more widespread effects on fluvial systems. Drummond (1888: 158) postulated that while

Egypt was the gift of the Nile, that river's sediments resulted from 'the labours of the humble termites in the forest slopes about Victoria Nyanza'.

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A.S. GOUDIE

TERRACE, RIVER

A river terrace is the planar surface that remains after the river, which formed it, incised its former valley floor. River terraces are abandoned river channels and floodplains. Their presence in river valleys throughout the world provides a record of changes in the flow regimes of rivers and the sediment supplied to them over time.

The flat surface of a terrace, called the *tread*, represents the highest elevation of the valley floor before incision occurred. Terrace treads, also called benches or platforms, are composed of alluvium, bedrock, or bedrock covered with a thin deposit of alluvium. They dip downvalley, recording the gradient of the channel that formed them, unless tectonic or isostatic uplift has subsequently altered their slope. The steep slope between treads or between the tread and the active floodplain is called the *riser*. Flights of terraces represent multiple, discrete episodes of downcutting, punctuated by periods of stability or AGGRADATION. Terraces may be continuous along a valley, or discontinuous if portions of the same terrace have become separated by tributary entrenchment or other geomorphic processes. More recent deposits, including those of mass movements, alluvial fans, volcanic ash, or windblown fines, may bury terrace treads. In a tectonically active area, faulting can alter the height relationships between terraces.

Genetically, terraces are considered to be either depositional (fill) or erosional (cut) landforms. Depositional terraces form as a result of the aggradation and later entrenchment of alluvium. They are abandoned floodplains, and stratigraphically show vertical and lateral processes of sediment accretion. Erosional terraces are surfaces formed by the erosional removal of bedrock or alluvial fill from the former valley floor. The term river terrace is used inclusively to describe the landform without specifying the materials (bedrock or fill) or the genetic processes (erosional or depositional) responsible for a specific feature. When the ages or relative ages of a flight of terraces are known, terrace surfaces are usually identified numerically, with the lowest number (1) used for the oldest surface. By recording changes in the flow regime of rivers, terraces are important indicators of tectonic, climatic, and even anthropogenic environmental history.

Terrace-forming processes

Different processes and circumstances, acting alone or in combination, cause rivers to incise, stranding their former valley floors above the active channel. Incision may begin gradually or catastrophically. A river cuts down through its own deposits when greater flow energy, lower BASE LEVEL, and/or less sediment load increase its erosional capacity. Discharge and stream energy can be increased by changes in climate and by processes, including base-level lowering, that steepen the channel gradient. Climatically driven incision occurs when climate becomes wetter, when ice melts (warmer climate), or when upstream climate–vegetation–soil relationships lead to conditions of flashier rainfall runoff. The latter would occur where decreasing precipitation causes a marginally semi-arid area to become more arid such that vegetation becomes sparser, soil erosion accelerates, infiltration rates decrease and the proportion of rainfall flowing to the river as storm runoff thereby increases. Anthropogenic activities, including forest removal and road building, also increase the flashiness of runoff. Factors that cause the channel gradient to steepen increase the energy of the flow, allowing the river to entrain materials previously deposited. Increased gradients result from tectonic and isostatic uplift, headward propagation of knickpoints, faulting, or lowering the erosional base level. Many of the terraces in contemporary landscapes are Pleistocene or Holocene in age and record changes in river regimes due to shifts in climate and in geomorphic process regimes between glacial and interglacial periods. Among the multiple factors leading to Pleistocene river incision and terrace formation was the lowering of sea level during glacial periods, when more water was stored on land as ice. Glacial period sea levels dropped more than 100 m, lowering erosional base levels and steepening continental river channel gradients accordingly (see GRADE, CONCEPT OF).

Changes in sediment supply tip the balance between aggradation and degradation in a river. A river's capacity to transport sediment derives from the volume and calibre of sediment supplied to it and the energy available to move the sediment. In a period of little environmental change, the channel geometry of a river, including its gradient, represents the discharge, energy and sediments characteristic of that environment at that time. When active glaciers, for example, provide an abundant sediment supply, the resulting channel will become steep, swift, shallow and probably braided. Reducing the sediment supply leaves such a river, adapted for heavy sediment loads, with excess energy. Thus, reducing the sediment supply upstream leads to more aggressive erosion and downcutting downstream, a condition that will continue until the flow regime becomes more in balance with the available energy and sediment. The supply of sediment to a river can be diminished by changes in climate or land management practices that increase vegetative cover or decrease mass movements and wind erosion on upstream land surfaces. Dams stop sediment, too. Such changes increase the erosional energy of the river downstream, possibly to the point that it will begin to incise its own deposits. RIVER CAPTURE of a higher gradient, sediment-laden headwater river by a lower gradient piedmont river causes the piedmont river to gain additional sediment input from the high gradient headwaters, and the headwater river to adjust to a lower local base level. Steepening gradients in the capturing stream can cause flow energy to increase, and the discharge of the capturing stream may also increase due to the increased size of its drainage basin. Finally, catastrophic events, e.g. outburst flooding from a glacial lake or a landslide-dammed lake, can trigger catastrophic incision and initiate terrace formation.

Depositional terraces

A depositional (fill, aggradational) terrace is a former floodplain that has been incised by the river. It differs from an active floodplain by being too far above the river channel to convey the overbank flows of the mean annual flood. Massive alluvial deposits in large terraces represent conditions of abundant sediment supply to the river (Figure 164a). The sediments in depositional terraces were built up by the accretion of alluvial sediment, either vertically or laterally, during a period in which that surface was the active floodplain of the river. For floodplains built by vertical accretion, a cross section of the terrace would reveal horizontal stratification, with flood deposits on top of flood deposits sorted by size and fining upward. Gravels in the deposits would be rounded and probably imbricated and aligned to point downstream. If the floodplain had built up by lateral accretion, the deposits would have the stratigraphy of POINT BARS. Depositional terraces may be metres to kilometres wide, and follow a present or former course of the river for thousands of kilometres, often on both sides of the channel. Flights of depositional terraces occur where rivers have developed new floodplains between separate episodes of downcutting.

In many examples around the world, repeated episodes of valley filling followed by major evacuation of sediment have created younger (inner) depositional terraces developed on fill material which was emplaced after the valley was scoured out between higher, older terraces (Figure 164b). Depositional terraces can be distinguished from erosional terraces in that (1) the tread surface of a depositional terrace represents the uneroded surface of a valley fill and (2) the underlying rock surface (if its topography can be determined seismically or viewed in a transverse cut) may be very irregular.

Erosional terraces

The surface of an erosional river terrace was levelled by lateral fluvial erosion before the river

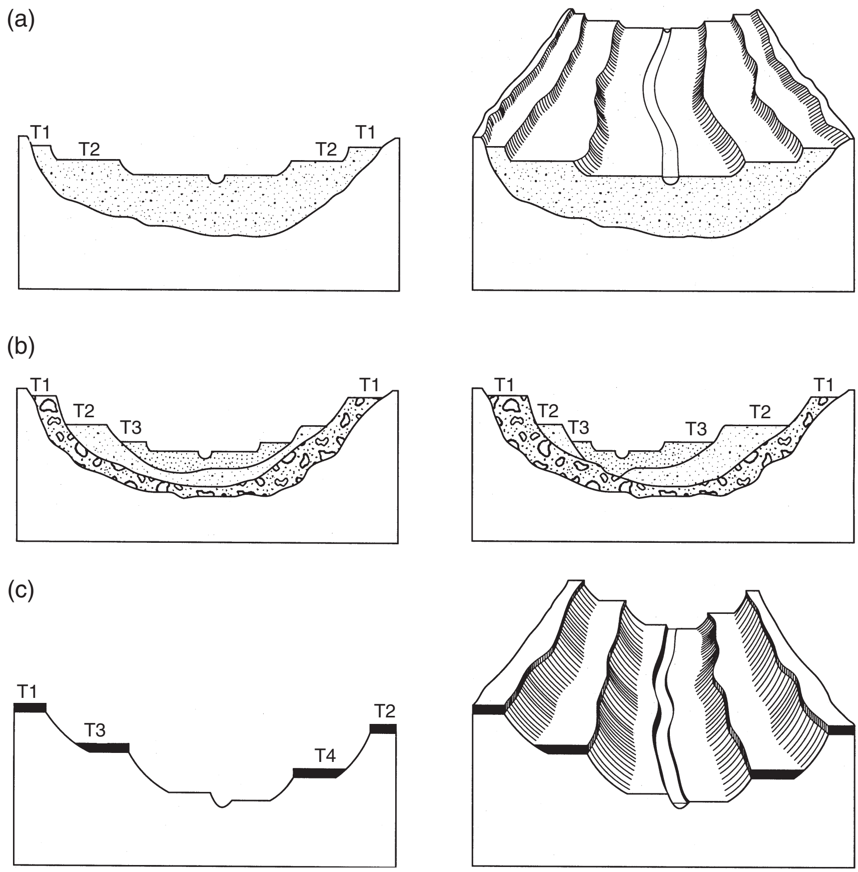

Figure 164 Valley cross section and block diagrams illustrating (a) paired depositional terraces, (b) depositional terraces formed from multiple episodes of valley fill and entrenchment, and (c) unpaired strath terraces

incised to form the terrace. Like depositional terraces, erosional terraces are remnants of an older valley floor, into which the river has cut down. Erosional terraces scoured from bedrock are called *bench, strath*, or *rock-cut* terraces (Howard *et al.* 1968: 1,117). The term 'strath' is a Scottish word meaning wide valley. Rivers carve straths by lateral corrasion, with abrasive alluvial sediments eroding the channel bed from side to side as meanders migrate downstream and/or the meander radius increases. The underlying bedrock surface, as nearly planar as channel beds and valley bottoms are today, lies parallel to the surface of the thin cap of alluvium (Figure 164c). The thickness of the alluvial cap indicates the depth of scouring in the former river regime: the alluvium must be thin enough so that the river could have been in erosional contact with the bedrock valley bottom. A thick valley fill, on the other hand, protects the valley bottom from erosion. The maximum thicknesses of alluvial lag deposits, usually gravel, on strath terraces depends on the size and energy level of the river. Mackin (1937: 828) reported 2.5 m of gravel with a cover of silt in the Shoshone Valley (Wyoming, USA); Wegmann and Pazzaglia (2002: 734) suggest 3 m as a typical maximum depth.

Not all erosional terraces are strath terraces with eroded bedrock at or just below the surface. A valley floor composed of unconsolidated fill may also become truncated by lateral erosion and subsequently stranded by incision to become a river terrace. Such terraces are called *fill*-cut (Bull 1990: 355) or *fill-strath* (Howard *et al.*

1968: 1,119). *Structural* terraces are erosional benches of resistant bedrock resulting from differential erosion rather than from changes in the river flow regime. Structural terraces abound in the Grand Canyon of the Colorado River, where contrasting erodibilities of the nearly horizontally bedded sedimentary strata cause the resulting canyon walls to appear as rock steps.

Paired and unpaired river terraces

Where terraces occur at the same elevation across a valley, they are called *paired* (Figure 164a); otherwise they are *unpaired*. Paired and unpaired terraces can be erosional or depositional; they can be formed on bedrock or on alluvial fill. Paired terraces represent conditions in which downcutting predominates over lateral cutting. A river flowing across the middle of a thick floodplain would respond to a drop in base level by incision. The result would be a lower channel between a pair of terraces. If the base level remained at the same low position for an extended period of time, the period of downcutting, accompanied by mass movements of channel banks, would be followed by a period of aggradation in which the river might form a new floodplain along the entrenched channel. A second episodic event that renewed the greater erosional capability of the river could initiate a second phase of downcutting, causing the river to abandon the newer floodplain and incise a second, inner set of paired terraces. Such pairs of depositional terraces are assumed to be of equal age.

Unpaired river terraces reflect conditions in which downcutting is slow and lateral erosion is occurring at the same time (Ritter 1986: 269). Lateral migration of the river channel can erode and eliminate some older terraces, leaving an asymmetrical record of depositional or erosional surfaces. Unpaired terraces are unlikely to be of equal age, and likely to be erosional in origin. Strath terraces are typically unpaired.

Terraces as evidence of environmental change

Terraces may record tectonic events, changes in climate and other environmental changes that alter the erosional capacity and sediment load of a river. Dating terrace surfaces enables researchers to calculate incision rates and better understand the climate or tectonic history of a region. The age of a terrace is determined by its relative position in the valley and by other evidence present in its biological, chemical and anthropogenic record. Alluvial particles on a strath terrace are synchronous with the time period of erosion, whereas alluvial particles in a depositional terrace predate the time of terrace formation by the time required for a slug of sediment to travel to that location in the river system and the time lapse between deposition and incision (Bull 1990: 360). Organic matter buried in the terrace can be dated with radiocarbon (^{14}C) dating techniques (e.g. Wegmann and Pazzaglia 2002: 734), and investigators have begun to date clasts on terrace surfaces based on the concentration of cosmogenic isotopes such as ^{10}Be and ^{26}Al (e.g. Hancock *et al*. 1999: 47) (see DATING METHODS). Terrace ages can sometimes be inferred from the environmental setting portrayed in the biological record. The presence of sub-Arctic molluscs and fossil ice-wedge casts in terraces along the River Thames indicates deposition of terrace material during cold periods, interpreted as glacial periods in the Pleistocene (Goudie 1984: 292). Terraces from the late Tertiary or Pleistocene can be distinguished from Holocene terraces, not only by their biota and their position in the landscape, but by the degree of weathering of terrace materials. In some older terraces, formerly unconsolidated terrace materials have become cemented by carbonate, silica or iron oxides (Costa and Baker 1981: 161). Human artefacts, including ruined structures and Roman coins (Judson 1963: 899), have helped date Late Holocene terrace surfaces.

In some investigations, terrace formation is attributed to a single causative factor. Born and Ritter (1970: 1,240), for example, attributed the flight of terraces in the Truckee River above Pyramid Lake (California, USA) to an anthropogenically lowered base level. Alternatively, terrace formation can represent a complex landscape response to one change (e.g. a climatic factor) or the integrated landscape response to a set of changes (e.g. climatic and tectonic). The terrace record of the River Rhine, which records the climatic and erosional history of the region, has also been affected by uplifts in the middle and subsidence in the lower portions of the valley (Fairbridge 1968: 1,131).

Evidence provided by river terraces can be complex and challenging to interpret. Erosional episodes can remove older terraces; in fact, terrace sequences are rarely found completely intact. Terraces in one drainage basin may respond more to local factors than to regional climatic or tectonic controls. As an example of this, Brakenridge (1981: 75) found terrace formation along the Pomme de Terre River in southern Missouri (USA) to not match either the sea-level history for the Gulf of Mexico or the glacial chronology of the upper Missouri–Mississippi basin. In other examples, Ritter (1982: 352) found terraces in one valley in the Alaska Range to have formed after a moraine-dammed lake overtopped a drainage divide, and Wegmann and Pazzaglia (2002: 740) did not find base-level control responsible for terrace formation in the Clearwater River (Olympic Mountains, USA), even though the river flows directly into the Pacific Ocean. They suggest that the Clearwater River operates at or near its capacity for sediment transport and responds to changes in upstream sediment production, some of which is likely to be earthquake-related, with episodes of vertical (terrace-forming) or lateral (valley-widening) incision. Their interpretation is supported by Schumm's (1975: 77) work showing COMPLEX RESPONSE of a fluvial system could lead to small terrace formation without external forcing variables. Bull (1990: 352) emphasizes a scale difference between major terraces, particularly climatically caused aggradation surfaces and large tectonically caused straths, and minor terraces, which develop in response to local factors.

Although terrace tread formation is generally thought to reflect relatively stable conditions over long periods of time, terraces are also known to have formed catastrophically or over intervals of only a few years. From photographic and historical evidence, Born and Ritter (1970: 1,240) documented the formation of at least six well-developed river terraces upstream of Pyramid Lake (California, USA) in the forty-four years since water diversion began to lower the lake level.

Tread surfaces of terraces in the contemporary landscape may not have been disturbed by erosion or deposition since the time of their abandonment by the active river channel. A flight of such terraces presents a CHRONOSEQUENCE of weathering and soils, and terraces of known age present special opportunities for studying soil development (Bull 1990: 352). In inhabited areas, opportunities for the study of terrace soils are commonly constrained by human disturbance. Younger terraces are sought as sources of sand and gravel, and terrace surfaces are often chosen for agriculture, urbanization, and the location of highways and airports. Their relative flatness in high-relief environments and their position above elevations of frequent flooding and poor drainage makes terrace surfaces attractive for human occupation.

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SEE ALSO: floodplain; fluvial geomorphology; sediment load and yield; valley

CAROL HARDEN

TERRACETTE

A miniature unvegetated step-like feature that forms on hillslopes. Terracettes extend across the slope in a parallel manner, though predominantly following the contours of the land. They commonly form on ground that is fairly unconsolidated, particularly pasture, possessing moderate to steep hillslope gradients. Terracettes are rarely greater than 0.5 m in height and depth, with a spacing of about 1 m, and may extend laterally for tens of metres. They may form in a variety of climatic environments. The vertical drop exhibited in a terracette is known as the riser whereas the horizontal platform is called the tread. Additionally, the base of the riser is referred to as the foot of the terracette, while the point where the tread meets the riser is termed the crown.

Terracettes are irregular and often anastomosing forms, and may feature as intermittent steps or a whole network of steps covering a hillside. On a loessic slope of 24°, it has been estimated that terracettes may form on up to 11 per cent of the ground surface, and as much as 40 per cent on a slope angle of 37° (Selby 1993). Further terms for a terracette include pseudo-terraces or false terrace. Angle of slope is crucial for terracette formation. An average slope angle of about 30° dictates the boundary for terracette formation. Below this angle the slope face will not break, but instead will exhibit a distinctive undulating surface. Surficial material (based on grass) will begin to crack at angles greater than 30°, while at angles greater than 50° cracks occur at the back of each tread and small-scale slumping occurs, thus developing the characteristic step-like feature (Selby 1993: 258).

The origin of terracettes is contentious, and several explanations of their development have been produced. However, it is likely that the following mechanisms of formation are interrelated, and that each example is the result of a mix of mechanisms, unique to the site. The dominant mechanism of terracette formations is by soil creep, and occurs when hillslope angles are greater than that of which the unconsolidated mantle material can remain stable, resulting in slippage. Terracette formation is not limited to open grassland, and may develop in forested land as soil is washed downslope and leaf litter accumulates in the wake of tree roots. Additionally, erosion may be provoked or enhanced by the trampling of land by livestock. The hooves of both cattle and sheep can remove and wear down the surficial materials, particularly upon frequently used tracks and walkways (termed cattle or sheep tracks).

Terracettes can also result from near-surface faulting, which may break the overlying land into a series of small terracettes. Additionally, they can also be ablation products, man-made aids for cultivation on hillslopes (lynchets), as products of solifluction and antiplanation, or as miniature fluvial terraces.

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STEVE WARD

TERRAIN EVALUATION

The term *terrain evaluation* has been used to describe a wide range of geomorphological techniques and no single definitive meaning has been established. In its narrowest definition terrain evaluation is regarded as synonymous with mapping of LAND SYSTEMS, a procedure for classifying the landscape by dividing it into landform assemblages with similarities in terrain, soils, vegetation and geology (Mitchell 1973). In a slightly broader definition, Lawrance *et al.* (1993) regarded terrain evaluation for engineering projects as a method for summarizing the physical aspects of a landscape initially through classification and then including an assessment of ground conditions in terms of engineering requirements. Griffiths and Edwards (2001) use the expression *land surface evaluation* as an alternative to terrain evaluation in ENGINEERING GEOMORPHOLOGY and engineering geology because the varied usage had created confusion and led to misunderstanding. However, in practice terrain evaluation and land surface evaluation are synonymous and therefore the definition proposed by Griffiths and Edwards (2001) is the most appropriate to use. This states that it is the evaluation and interpretation of land surface and near-surface features using techniques that do not involve ground exploration by excavation (except using small hand-dug pits or hand auger holes) or geophysics. Based on this definition terrain evaluation can be regarded as integral to the development of the ground model proposed by Fookes (1997) as central to all successful civil engineering construction. The definition would also be suitable when terrain evaluation is used as a technique of APPLIED GEOMORPHOLOGY in planning and environmental studies (see Smith and Ellison 1999). In these studies, the terrain evaluation procedure would include an evaluation of soils, vegetation, land use, materials, drainage and human activity in addition to geomorphological processes and landforms.

The techniques that can be employed in terrain evaluations include: GEOMORPHOLOGICAL MAPPING, geological mapping, engineering geological mapping, remote sensing interpretation, analytical photogrammetry, land systems mapping, natural hazard and risk assessment, and the use of Geographical Information Systems (GIS). The output from a terrain evaluation is usually a suite of maps either held as hardcopy or as a suite of overlays in a GIS. The map data can be classified into three categories:

1 Factual or element maps that record the actual ground conditions (Table 45a).
2 Derivative maps that are obtained by either combining element maps or are based on an interpretation of the element maps (Table 45b).
3 Summary maps that pull together a range of derivative and element maps to identify combinations of hazards, resources or land use issues that act either as constraints to any development or indicate the potential of the land for exploitation (Table 45c).

Table 45 Terrain evaluation map categories

Terrain evaluation map category	Examples of typical maps
(a) Element maps	• morphology • topography • bedrock geology • superficial geology • lithology • vegetation • pedology • land use • geotechnical properties • location of sites of special scientific interest • exploratory holes and wells • hydrology
(b) Derivative maps	• slope steepness that uses topography to classify maps into distinct groups based on morphology and topography • depth to bedrock that utilize data from the geological maps • geomorphology • hydrogeology • depiction of various types of resources, such as sand and gravel or brick clay • foundation conditions for engineering structures • geotechnical zoning, i.e. areas of homogeneous ground conditions • hazards, such as subsidence, landslides, flooding or contaminated land • previous industrial usage
(c) Summary maps	• development potential • potential resources • planning constraints, including statutory protected land • construction constraints

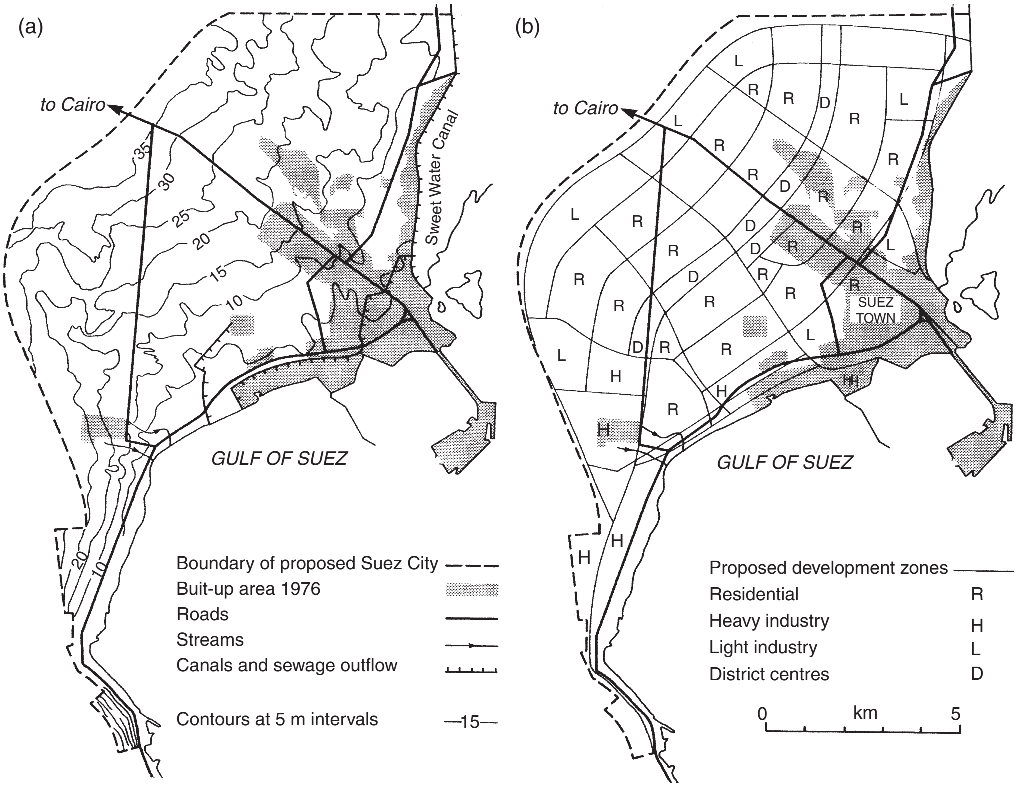

Figure 165 Maps of Suez, Egypt: (a) topography and extent of urban area in 1976; (b) proposed layout of an enlarged urban area (after Jones 2001, reprinted with permission of the Geological Society of London)

An extended legend is normally attached to each of the maps and most terrain evaluation studies would include an interpretative report that explains the basis for the development of the derivative and summary maps.

Terrain evaluation is best illustrated through a case study and Jones (2001) provides a classic example of its use for flood hazard assessment. In this study of a proposed development for Suez New City (Figure 165) geomorphological mapping was initially undertaken based on aerial photograph interpretation and field mapping. This resulted in the production of a geomorphological map, originally at a scale of 1:25,000, that established the distribution of a suite of marine, fluvial and bedrock features in addition to existing areas of urban development (Figure 166). These data were then combined with an evaluation of WADI catchment areas and channel form, based on further aerial photograph interpretation and field data analysis, to produce an interpretative map of flood hazard utilizing an ordinal scaling system (Figure 167). The final hazard map provided the base for both planning development and identifying the areas requiring flood protection.

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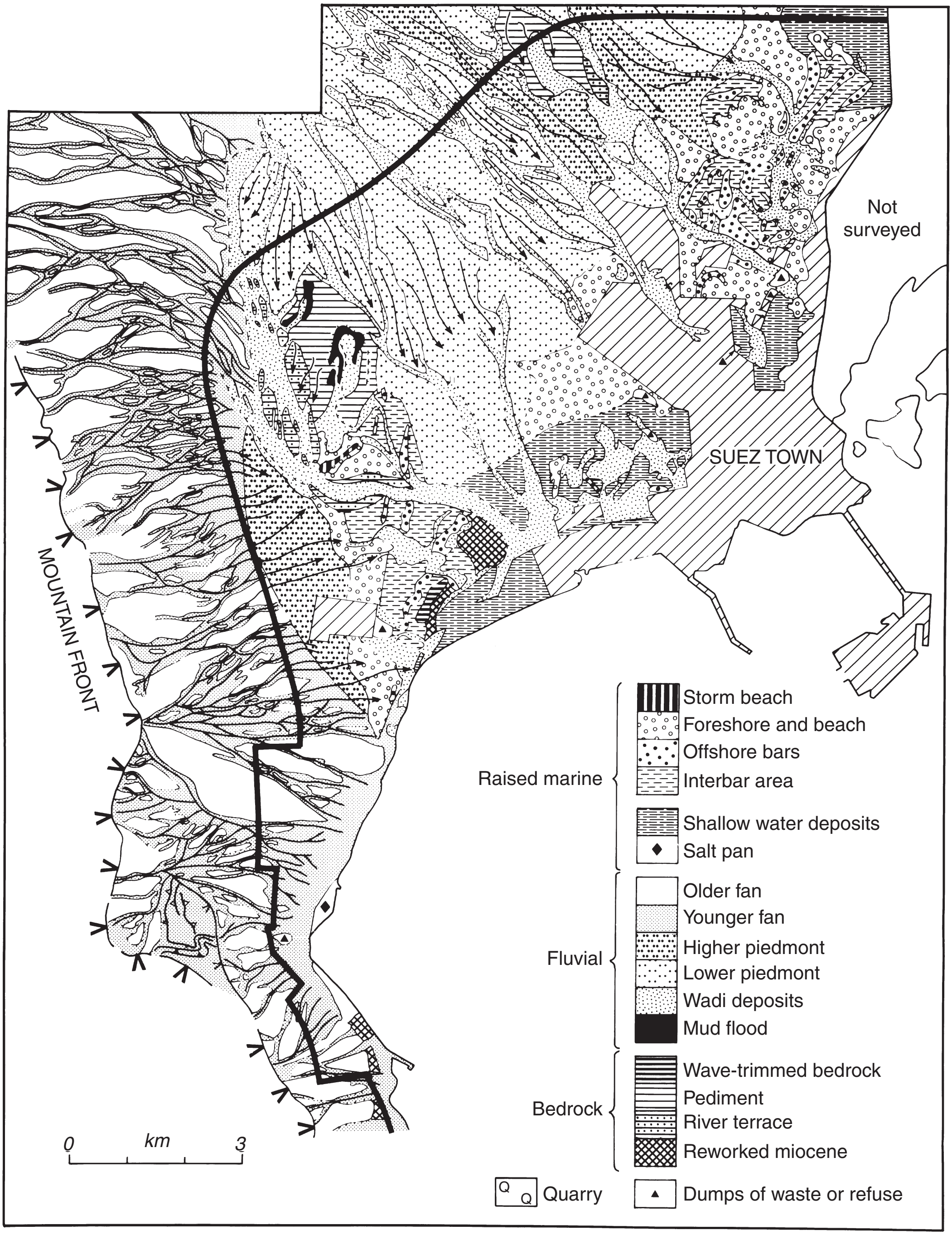

Figure 166 Geomorphological map of the Suez area produced by aerial photograph interpretation and ground mapping (after Jones 2001, reprinted with permission from the Geological Society of London)

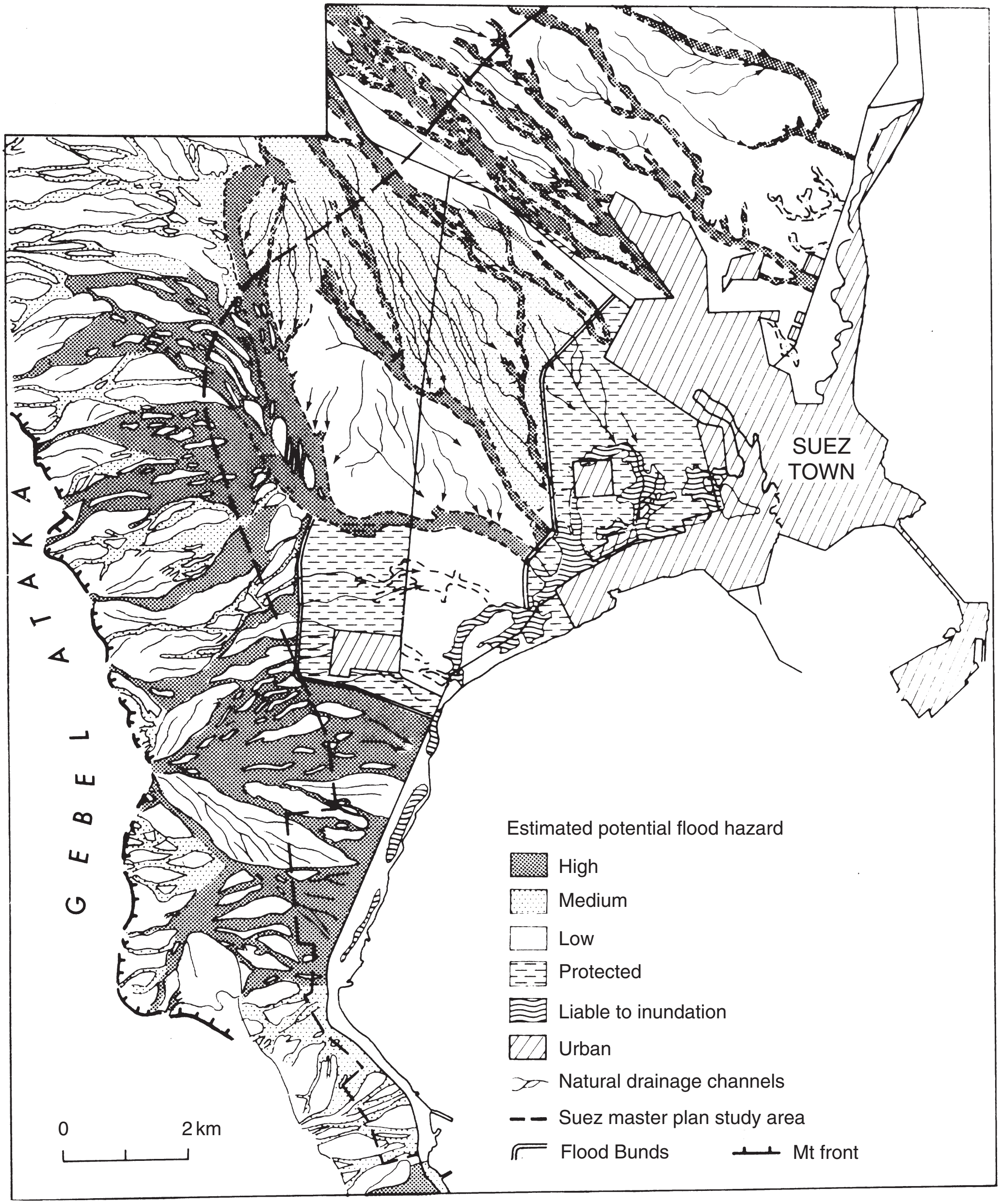

Figure 167 Flood hazard map of the Suez area (after Jones 2001, reprinted with permission from the Geological Society of London)

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JAMES S. GRIFFITHS

THERMOKARST

The term 'thermokarst' was created in 1932 by a Russian, Yermolayev (Czudek and Demek 1970: 103), to describe an uneven morphology (thermokarst terrain) with some similarities to karstic morphology and resulting from soil subsidence due to the melting of ice within PERMAFROST. Later, the same word was used mainly to refer to the processes of soil subsidence resulting from thaw.

A prerequisite to thermokarstic terrain is excess ice within permafrost; in other words, the volume of ice must exceed the volume of the pores that water, under natural conditions, can fill in the soil. If excess ice is present, some change, even a small one, in the surface conditions (e.g. vegetation) can induce the melting of the upper part of the permafrost; such melting recurs during several summers and causes collapse that, cumulatively, can be important. Such collapses will continue until the ACTIVE LAYER has regained a sufficient thickness to protect the permafrost from further melting. Excess ice does not occur everywhere in permafrost. It is mainly found in the low ground of valley bottoms and in coastal lowlands containing abundant silty clays – locations of thaw-sensitive permafrost. Thermokarst also refers to collapse resulting from the disappearance of glacial ice. The forms resulting from melting of ice differ according to the distribution of the ice in the soil and, thus, according to the types of ice. The principal forms of thermokarst are summarized below.

Thermokarst lakes are widespread in the coastal lowlands of Siberia, Alaska and the Mackenzie Delta area (Canada), where excess ice is abundant in the soil. Such lakes are rounded depressions and are rather shallow. They progressively enlarge and their diameter can reach 1–2 km.

In some regions, thermokarst lakes are elongated and under influence of the direction of prevailing winds. They are called 'oriented lakes'. The mechanism leading to such a shape is not quite clear, but it now seems that the longer axis of the oriented lakes is perpendicular to the direction of the prevailing summer winds.

Thermokarst lakes also develop at the expense of ice wedges (see ICE WEDGE AND RELATED STRUCTURES). If a pool appears above an ice wedge (often as a result of a topographic change induced by the growth of the ice wedges), this pool warms the underlying soil and melts the top of the permafrost, i.e. the ice wedge. Because of this melting, the pool grows above the ice wedge, forming linear and polygonal troughs separating centres of the polygons (Plate 139). The centres, which contain less ice, are brought out into relief and evolve into conical mounds called 'thermokarst mounds'. Related to melting ice wedges, beaded streams may develop; these are characterized by narrow reaches linking pools or small lakes. The pools occur at the junction of the ice wedges.

In Siberia, well-known thermokarst forms are the ALASes, thermokarstic depressions with steep sides and flat bottoms covered by grassland, where shallow lakes often exist. Such depressions are generally round or oval, 3–40 m deep and 0.1–15 km long. They occupy 40–50 per cent of the surface of

Plate 139 Thermokarst in Siberia. Vegetation was cut near the road and a part of the active layer was stripped off, causing melting of a polygonal net of ice wedges and consequent formation of thermokarst mounds

the Lena and Aldan terraces in central Yakutia (Washburn 1979: 274) and they result from the melting of exceedingly ice-rich permafrost developed in the silty cover of the terraces.

Melting PINGOs induce formation of closed depressions surrounded by ramparts; such depressions are also thermokarstic. In contemporary permafrost zones, they are called 'collapsed pingos' or 'pingo remnants'. Generally, pingos begin to melt at their top because of the cracks relating to their growth.

Melting PALSAs also give birth to depressions that are visible only with regards to the areas raised by permafrost. When permafrost has disappeared, only very short-lived, shallow depressions remain marked by vegetation different from that on both previously unfrozen peatlands and the remnants of the peaty permafrost landforms.

Lithalsas (the same forms as palsas, but without any cover of peat, and developed in mineral soil), after melting leave, like pingos, depressions surrounded by ramparts. However, unlike pingos, but in the same way as palsas, lithalsas appear as numerous and almost adjacent forms.

On slopes, the most spectacular thermokarst forms are the 'retrogressive thaw slumps'. Described from the boreal forest as well to the High Arctic, they result from a process initiated by thawing of ground ice. It begins by the slide of the active layer on the permafrost table, which acts as a lubricated slip plane for movement and controls the depth of the failure plane. This process produces semicircular hollows opening downslope and usually less than 2 m high (French 1996: 119). Further thawing of permafrost produces steep slopes as much as 8 m high.

A peculiar thermokarstic phenomenon is fluviothermal erosion, namely erosion by the water of rivers, lakes or sea, attacking the permafrost not only by mechanical erosion processes, but also by its warmth which melts the ice. Such erosion is rapid and causes undercutting of riverbanks, particularly in sandy layers. It forms thermo-erosional niches at the floodwater level.

Thermokarstic phenomena have climatic or local causes. Climate warming, by increasing the thickness of the active layer, leads to the melting of the upper permafrost and to thermokarstic phenomena. However, response to climate warming is complex: besides temperature, snowfall also varies and changes in vegetation occur. The immediate response is not obvious. Nevertheless because of the global change, thermokarstic phenomena are finally to be feared.

But more often than not, permafrost is melting because of local causes. Thermokarstic phenomena, for instance, result from destruction or change of the vegetation cover as a consequence of forest fires or human action. Vegetation plays a complex role, generally protecting soil against warmth more than against cold. It acts in winter by trapping snow between branches and impeding insulation of the soil, in summer by protecting the soil by its shade, or by decreasing air circulation, by increasing evaporation, etc.

On areas weakened by excess ice in the soil, buildings, roads, airports, pipelines, etc. pose severe problems that engineers try to solve by impeding the melting of the permafrost. Houses and pipelines are built on piles; roads and airstrips are embanked above the ground in order to lift the permafrost surface; sometimes cooling devices are put into the ground to radiate its warmth to the surface.

Surveying fossil thermokarstic forms in regions where permafrost existed in the past has aroused interest in many scientists. Such forms already mentioned are pingo and lithalsa scars, recognizable by the ramparts surrounding the depressions (Pissart 2000: 344). Traces of thermokarstic collapses have disappeared because of the general subsidence induced by the melting of the whole permafrost. Only remnants of peculiar sediments deposited in those temporary hollows reveals their former existence. However thermokarstic explanations are often invoked without such observations. Their origin remains uncertain.

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ALBERT PISSART

THRESHOLD, GEOMORPHIC

A geomorphic threshold can be defined as the critical condition at which a landform abruptly changes. The change can be the result of an external variable, that exceeds the stability of a landform at an extrinsic threshold, or the change at an intrinsic threshold can be the result of a progressive change of the landform itself.

Extrinsic thresholds have been recognized in many fields. Perhaps the best known is the threshold velocity required to set in motion sediment particles of a given size. With a continuous increase in velocity, a threshold velocity is reached at which sediment movement commences, and with a progressive decrease in velocity, a threshold velocity is encountered at which sediment movement ceases. The best known thresholds in hydraulics are described by the Froude and Reynolds numbers, which define the conditions at which flow becomes supercritical or turbulent. Particularly notable are the changes of BEDFORM characteristics at threshold values of stream power. In these examples, an external variable changes progressively thereby triggering an abrupt change within the affected system at an extrinsic threshold. That is, the threshold exists within the system, but it will not be crossed and change will not occur without the influence of an external variable. The word, threshold, describes the critical range of conditions over which these transitions occur.

Thresholds can also be exceeded when the external variables remain relatively constant, yet a progressive change of the landform itself renders it unstable, and failure occurs at an intrinsic geomorphic threshold. An example is long-term progressive weathering, that reduces the strength of slope materials until eventually there is slope adjustment and MASS MOVEMENT. Another example of an intrinsic threshold is provided by a typical sequence of morphologic changes resulting in the collapse of sandstone-capped cliffs. Beneath a vertical cliff of sandstone is a gentler slope of weak shale. Through time, the basal shale slope is eroded, which produces a vertical shale cliff beneath the sandstone cap. At some critical height, the cliff collapses and the cycle begins again. The episodic retreat of this type of escarpment is the result of the change in cliff morphology under essentially constant climatic, base level and tectonic conditions. Similarly, a meander can increase in amplitude until a cutoff occurs under constant hydrologic conditions.

Field and experimental work supports the concept of geomorphic thresholds, which have been used to explain the distribution of discontinuous gullies in semi-arid valleys. Discontinuous gullies, short gullied reaches of valley floors, can be related to the slope of the valley-floor surface. For example, the beginning of GULLY erosion in these valleys tends to be localized on steeper reaches of the valley floor, which are the result of sediment storage. For a region of uniform geology, land use and climate, a critical intrinsic threshold of valley slope exists above which the valley floor is unstable and subject to incision during floods.

Similar relationships can be established for other alluvial deposits. For example, trenching of ALLUVIAL FANs is common, and the usual explanation for fan-head trenches is renewed uplift of the mountains or climatic fluctuations. However, as the fan grows through continual deposition, the fan-head steepens until it exceeds a threshold slope, when trenching occurs. Experimental (see EXPERIMENTAL GEOMORPHOLOGY) studies of alluvial-fan growth confirm that periods of trenching alternate with deposition at the fan-head. Therefore, the fan-head trenching can occur as a result of the oversteepening of the fan-head, and it is the result of the exceeding of an intrinsic geomorphic threshold.

A similar example is provided by damaging debris-flow events along the Wasatch Mountains in Utah. In 1993, a storm triggered debris flows from some canyons, but not all. It was suggested that debris basins be constructed at the mouths of the active canyons. However, further investigation revealed that the active canyons had been flushed of sediment and were not a threat, whereas the inactive canyons were storing sediment, which at some future time would produce damaging debris flows. The storage and flushing of sediment in these canyons is similar to the storage of sediment and its incision in semi-arid valleys and on alluvial fans as an intrinsic threshold is exceeded.

The identification of an intrinsic geomorphic threshold has significant practical applications.

If, as in the study of discontinuous gullies and alluvial-fan trenches, the critical slope is identified (intrinsic threshold), then unfailed, but sensitive valley floors and fan-heads can be identified. In this way, preventive conservation can be practised, thereby preventing erosion rather than attempting to control it after incision has occurred.

The concept of intrinsic geomorphic thresholds, which involves landform change without a change in external controls, challenges the well-established geomorphic thesis that relatively abrupt landform change is the result of some climatic, base level, or land-use change. Therefore, the significance of the intrinsic geomorphic threshold concept for geomorphologists is that it makes them aware that abrupt erosional and depositional changes can be inherent in the normal development of a landscape and that a change in an external variable is not always required for a geomorphic threshold to be exceeded and for a significant geomorphic event to result.

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STANLEY A. SCHUMM

TIDAL CREEK

A creek is an inlet in a shoreline, a channel in a marsh, or another narrow, sheltered waterway. Creeks occur extensively on MUD FLATS AND MUDDY COASTS, on MANGROVE SWAMPS and on SALTMARSH surfaces (Eisma 1998).

Tidal creeks often have a high drainage density because of the large volumes of water that they drain. Saltmarsh creek densities may be 40 km/km^2 (Pethick 1984). The morphology of the creeks is also often distinctive. Although some may bear a superficial resemblance to dendritic river channel networks, flow along them is bi-directional (French and Stoddart 1992; Pestrong 1965). They have a tendency to taper upstream and flare downstream (Fagherazzi and Furbish 2001), and their discharge is determined by the tidal prism. In areas with a large tidal range or rapid seaward progradation, creek systems may be markedly linear in form. In areas with cohesive sediments creeks have steep edges, whereas in sandier areas they tend to be shallower and wider.

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A.S. GOUDIE

TIDAL DELTA

Tidal deltas are large sand bodies formed within, or in the vicinity of, tidal inlets. The latter may be associated with barrier island chains (see BARRIER AND BARRIER ISLAND) and the entrances to coastal lagoons (see LAGOON, COASTAL) or estuaries (see ESTUARY). Flood-tidal deltas form landward of the inlet mouth, under the influence of flood-tidal currents. Ebb-tidal deltas occur seaward of the inlet, predominantly under the influence of ebb-tidal currents and wave action.

The major morphological features of flood-tidal deltas typically include (after Hayes 1980): a seaward-dipping flood ramp, up which landward sand movement occurs through the migration of sand waves under the action of flood currents; subtidal flood channels, which extend into the inlet and which dissect the partly intertidal landward portion of the delta (the 'ebb shield'); marginal ebb-aligned spits; and spillover lobes formed by the action of ebb currents over the lower parts of the ebb shield.

Ebb-tidal deltas are usually comprised of: an ebb channel, maintained by strong tidal currents; linear bars, formed through wave-current interactions along the margins of the ebb channel; a terminal lobe formed at the distal (seaward) end of the ebb channel, where the tidal current diminishes; sandsheets (or 'swash platforms') formed by wave action adjacent to the ebb channel characterized by migrating swash bars; and marginal channels dominated by flood-tidal currents.

Studies of inlet morphometry have shown that the morphology of tidal deltas is related to tidal prism (itself a function of both tidal range and inlet geometry), the configuration of the inlet and adjacent shoreline (including the offshore bathymetry), wave climate, and the rate of littoral sediment transport. In micro-tidal areas, flood deltas are often better developed than their ebb counterparts, owing to the dominance of landward, wave-driven, sediment transport. Ebb delta morphology is generally more variable than that of flood deltas, owing to the importance of regional and local contrasts in wave climate (Boothroyd 1985), and due to the tighter coupling of delta processes with wider coastal morphodynamic behaviour. Ebb delta volume increases with tidal prism, and decreases with inlet width/depth ratio and wave energy. Under conditions of low wave energy, ebb deltas are typically more elongated and extend further seaward. These controls are interactive so that, for example, wave energy can be modified by the presence of a headland, which also influences both the tidal prism and the intensity of tidal flows. For a sample of seventeen natural inlets in North Island, New Zealand, Hicks and Hume (1996) showed that over 80 per cent of the variation in ebb delta volumes could be successfully predicted by an empirical equation incorporating spring tidal prism and the angle between the ebb channel and the adjacent shoreline. Other factors accounting for some of the observed variation in delta volume include wave energy, sediment grain size (finer sands are less likely to be retained in the vicinity of strong tidal current jets and are thus associated with smaller deltas), and the supply of sediment through littoral drift.

FitzGerald *et al.* (2002) have drawn attention to marked contrasts in the occurrence and morphology of tidal deltas along the coast of New England, associated with a large degree of variability in tidal range, wave energy, sediment supply and inlet origin and geometry. Flood-tidal deltas in meso-tidal inlets tend to have a classic horseshoe shape and a significant intertidal area. Those in micro-tidal inlets tend to be predominantly subtidal and digitate or multi-lobate in form, owing to the limited ability of weak ebb currents to rework their deposits. Ebb-tidal deltas are best developed in moderately sized mixed energy environments. In wave-dominated inlets, they are either absent or small and entirely subtidal.

Although flood-tidal deltas act as long-term sediment sinks, ebb-tidal deltas are more dynamically coupled to the morphodynamic adjustment of the adjacent coast. Important processes include partial wave sheltering by the delta sand body; wave refraction around the delta, causing trapping and storage of beach sediments; and sediment recirculation within the delta (Oertel 1977). Ebb-tidal deltas interrupt the continuity of alongcoast sediment movement, and bypass sand across their inlets in discrete pulses. Hicks *et al.* (1999) have shown that such inlet/ebb-delta processes can be an important source of interdecadal variability in beach behaviour. Years when the delta is accumulating sand are associated with erosion of beaches on the downdrift side of the inlet. Conversely, in years when the delta releases sand, the same beaches experience accretion associated with a migrating sand pulse.

Tidal deltas have historically been exploited as a sand resource (including mining to supply BEACH NOURISHMENT schemes). Questions are now being asked over the sustainability of this practice and its implications for beach stability. Furthermore, the correspondence between delta volumes and tidal prism means that their sand-trapping function is potentially sensitive to sea level rise. Inlets with extensive intertidal areas might experience a significant increase in tidal prism with a rise in sea-level, thus leading to increased sand storage. This may have adverse consequences for the stability of adjacent beaches, especially downdrift of the inlet.

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J.R. FRENCH

TOMBOLO

A sandbar, barrier or spit that joins an island with a mainland or another island, resulting from longshore drift or the migration of an offshore bar toward the coast. Tombolos are constructive features (though ultimately ephemeral due to wave erosion), occurring along shorelines of submergence that are protected from large waves and where islands are common. Sediment supply is predominantly derived from the islands, yet some may also come from erosion of the shoreline, fluvial materials, underwater reefs and offshore glacial deposits. Several types of tombolo exist including single, double, multiple, forked, parallel and complex tombolos, all of which are reflective of the coastal system (e.g. wave mechanisms) from which they are derived. For example, double tombolos (two ridges extending to shore), often form in areas with seasonal shifts in longshore drift. Tombolos can restrict flow between the sea and intertidal zone, forming a lagoon (see LAGOON, COASTAL), and altering the local ecology. An example of a tombolo is Chesil Beach, which extends northwestwards from the Isle of Portland to the coast of Dorset, south England.

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SEE ALSO: bar, coastal; barrier and barrier island; coastal geomorphology

STEVE WARD

TOR

Linton (1955: 470) describes a tor as ‘a solid rock outcrop as big as a house rising abruptly from the smooth and gentle slopes of a rounded summit or broadly convex ridge’. Tors are in fact large, free-standing, residual masses of rock (Plate 140). The word derives from the Old Welsh word *twr* or *twrr* meaning heap or pile. Tors are most common – and most well known – in granitic rocks (e.g. Haytor Rocks, Dartmoor, England), but also occur in coarse sandstones, schists, dacites and dolerites, among other lithologies. Although perhaps best known in south-west England (Devon and Cornwall), tors occur on all continents. In Africa, tors are often known as castle koppies (or kopjes). The rocks in which they occur vary considerably in age, but it is thought that most tors formed during the Tertiary or Pleistocene. Tors may occur in any position in the landscape, but are most common in summit and spur locations (Gerrard 1978; Ehlen 1991). They may occur as single, massive exposures (e.g. Middle Staple Tor, Dartmoor), or consist of groups of individual outcrops clustered together (e.g. Great Mis Tor, Dartmoor). The latter configuration is most common in summit positions. They may also appear to be piles of very large, loose boulders (core stones), but a rock core anchored in bedrock is usually present in the centres of such exposures.

Plate 140 Middle Staple Tor, Dartmoor, south-west England. This tor is approximately 18 m long, 8 m tall and 10 m deep

Occasionally, the individual blocks in large summit tors form a pattern such that there is an elongate, open space, called an avenue, at the crest (e.g. Hound Tor, Dartmoor). Summit tors tend to be the largest, and those along valley sides tend to be the tallest (e.g. Vixen Tor, Dartmoor). Spur tors are most often the smallest. Sizes range from about one metre in height and a few metres in length and width to tens of metres in each dimension. Fallen rock debris, called clitter on Dartmoor, is often present at the bases of tors as well as for some distance downslope. Much of the clitter has been moved downslope by periglacial processes (e.g. stone lines on the west side of Middle Staple Tor, Dartmoor).

It is generally accepted that tor shapes and locations in all lithologies are controlled by JOINTING. Each tor typically contains three major joint sets (and two to five minor ones), one horizontal or gently dipping set defining the top, and two vertical or steeply dipping sets defining the sides of the tor. Diagonal joints may also be present (e.g. Great Tor, Dartmoor), but they are not common. Tors have a variety of appearances depending upon the arrangement of the joints that form them. They may be quite massive blocks of rock, either very rounded or blocky in appearance (Plate 140). They can be lamellar in form (Haytor Rocks, Dartmoor). They can also be tall and narrow, almost pinnacles (e.g. Bowerman's Nose, Dartmoor). These variations in appearance are caused by changes in the distributions of the different types of joints in the tor. If vertical and horizontal joints occur with about equal frequency, the tor will be composed of joint blocks of approximately equal size and will appear massive and blocky. If horizontal or gently dipping joints occur in significantly greater numbers than vertical or steeply dipping joints, the tor will be lamellar in form. If horizontal or gently dipping joints are rare, the tor will be tall and narrow. Logan stones, joint-bounded, precariously balanced boulders that move when touched, can be found in association with all types of tors, but are most common in shorter tors where horizontal joints are closer together than vertical joints, producing an elongate, rectangular block.

Origin of tors

Various theories have been proposed for the origin of tors. One theory suggests that they are the product of atmospheric weathering – that wind, rain, freeze – thaw, salt crystallization and insolation round the shapes of exposed, angular rock outcrops produce rounded, bouldery tors (e.g. Palmer and Neilson 1962). This theory is generally discounted except in certain unusual and specific cases (e.g. Selby 1972).

A second theory for the origin of tors assumes they are the product of a two-stage process (the two-stage theory). In this theory, based on Linton's (1955) work on Dartmoor in south-west England, tors form in the subsurface by chemical weathering along joints, and are subsequently exposed by erosional stripping. Once exposed, they retain their rounded shape. Linton noted a buried tor in a small quarry near Two Bridges in central Dartmoor as unrefutable evidence of this theory. Others (e.g. Thomas 1974) have expanded Linton's theory to include multiple phases of weathering and denudation. The theory suggests that weathering progresses most rapidly where the distance between joints is narrow, and more slowly where it is wide. Once the weathered mantle, which is thickest where the joints were most closely spaced, is removed by erosion, a rock outcrop with relatively widely spaced joints remains. Ehlen *et al.* (1997) showed that this was in fact the case in the Granite Mountains, Wyoming, and Ehlen and Wohl (2002) provided additional evidence favouring this theory in their study of BEDROCK CHANNELS in the Colorado Front Range. The weathered mantle formed from granitic rocks is called growan, GRUS, or saprolite. This theory has received significant support from many workers over the years, such as Eden and Green (1971) in their work on Dartmoor and C.R. Twidale in his many papers primarily on granite landforms in Australia. The genetic nature of the two-stage theory, however, has made it difficult for many to accept.

The third theory (the scarp retreat theory), proposed by King (1949) and based on his work in southern Africa, is that tors are the product of lateral planation and pediment formation. In a later publication, King (1958) accepts that sub-skyline tors could result from chemical weathering and exhumation, and states that his 1949 theory refers only to skyline (i.e. summit) tors. Ollier and Tuddenham (1961) in their work on Australian INSELBERGS and Ojany's (1969) on Kenyan inselbergs, among others, have provided support for King's scarp retreat theory.

The final theory of tor formation (the periglacial theory) is that proposed by Palmer and

Radley (1961). Their studies of sandstone tors in north-east England, where there is no evidence of deep chemical weathering, suggest that the tors were isolated from free faces along joint planes, and then rounded and fretted by subsequent atmospheric denudation. Palmer and Nielson (1962) studied the Dartmoor tors and suggested that – because core stones are lacking, there is no evidence of a DEEP WEATHERING PROFILE, and atmospheric weathering can account for the rounded shapes – the Dartmoor tors have a periglacial origin. They proposed a three-stage theory of periglacial tor formation. The application of this theory to the Dartmoor tors, however, is generally discounted.

The favoured theories among those described above are the modified two-stage theory first presented by Linton and King's scarp retreat theory, and much work has been done since they first published their work to support one or the other theory, as noted above. But perhaps the most reasonable approach to the origin of tors is that tors form by different processes in different environments (i.e. the principle of equifinality) as suggested by Brunsden (1964) and Thomas (1974), among others.

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SEE ALSO: exhumed landform; granite geomorphology; inselberg; rock control; salt weathering; spheroidal weathering; weathering

JUDY EHLEN

TOREVA BLOCK

Large masses of relatively stratigraphically coherent rock that have slipped down a cliff or mountain side upon normal listric faults, and has rotated backwards toward the parent cliff (Reiche 1937). The blocks can measure beyond 600 m in thickness and lateral extent, and are end members for this landslide type. Some blocks lie close to the parent cliff whereas others may have slipped several hundred kilometres from their sources. Their emplacement age remains uncertain, though probably in different (humid) climates in the Pleistocene.

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SEE ALSO: mass movement

STEVE WARD

TRACER

Tracer techniques enable geomorphologists to quantify the movement of Earth materials (whole systems, individual particles, water) and provide data that enable them to model the movement of these materials through a range of Earth systems. Applications have focused on four major research areas.

1 Studies of whole system behaviour (e.g. measurement of soil creep, mass movement and mudflows on hillslopes; measuring rates of movement and internal deformation of glaciers, measuring the surface deformation of volcanoes) (Meier 1960; Carson and Kirkby 1972; Anderson and Finlayson 1975; Goudie 1990).
2 Studies of coarse sediment transport on hillslopes, in fluvial systems and in littoral zones (e.g. determination of entrainment thresholds, transport distances, particle size and shape controls on sediment transport) (Sear *et al*. 2000).
3 Studies of fine particle transport and sediment provenance (e.g. airborne particles, hillslope erosion and soil redistribution, quantifying sediment-associated contaminant movement and determining the provenance of a range of dated fluvial, limnic, aeolian and marine deposits) (Foster and Lees 2000).
4 Tracing water movement and flow pathways (e.g. through soil profiles, groundwater and cave systems and estimating flood wave travel times in rivers) (Leibundgut 1995; Kranjc 1997).

While by no means exhaustive, this brief list of examples demonstrates the remarkable breadth of tracer applications in geomorphology.

Whole system tracer studies have made use of inert or active surface and subsurface markers (e.g. three dimensional Global Positioning Systems; steel/aluminium rods or spheres; distinctly painted stones or pebbles) whose movement can be monitored directly by satellite tracking, by field re-survey or by using remotely sensed images (e.g. aerial photographs).

Studies of coarse sediment transport may use 'passive' natural materials that are marked in some way to allow identification, recovery and measurement (e.g. painted pebbles, pebbles drilled with bar magnets or impregnated/coated with radionuclides). Alternatively, a range of natural or artificial materials may contain radio transmitters to allow rapid location even if the particle is buried.

Studies of fine particle movement may use 'exotic' materials that have characteristic signatures (e.g. fine-grained magnetic powders; fluorescent micro-spheres) that can be detected by field survey or by field sampling and laboratory analysis. Natural properties of environmental materials can also be used if their signatures (e.g. geochemical, mineralogical, mineral magnetic, stable isotope, radionuclide) characterize a distinct source of origin (e.g. geological or lithological units, soil types, topsoils/subsoils).

Tracing water movement and flow pathways has made use of fine-grained neutrally buoyant materials that move in the same way as the liquid (e.g. *Lycopodium* spores) or a wide range of (fluorescent) soluble dyes (e.g. Rhodamine) that can be added to cave or groundwater systems or to the surface of a soil profile and which may be detected at low concentrations.

Whichever method is appropriate to the research problem, a number of assumptions underpin the use of all tracers whatever the field of study. Before looking at an example, we need to ask:

What makes a good tracer?

There are a number of factors which need to be taken into account in order to ensure a successful outcome in tracer experiments:

THE TRACER DOES NOT INTERFERE WITH OR ALTER THE PROCESS BEING MEASURED

In many cases this is difficult to achieve. For example, digging soil pits and excavating or drilling holes in glaciers, installing metal pins to estimate rates of soil creep or ice deformation, or installing piezometer tubes in a mudflow, directly disturb the immediately surrounding environment. While widely used, these and other intrusive methods of measurement are subject to unknown and largely unknowable errors.

THE TRACER IS REPRESENTATIVE OF THE SYSTEM BEING MEASURED

Artificial coarse and fine tracers may differ in size, shape and density from the natural materials being investigated. All these factors are important. For coarse particle studies, for example, entrainment and settling velocities are functions of the mass, density and shape of the particle. In fine particle studies (especially in the silt and clay-sized fractions), particle interactions are important and many natural materials move as aggregates rather than as individual particles. Replacement of natural by artificial materials may poorly replicate particle interactions.

THE TRACER CAN BE RECOVERED AND IDENTIFIED

Historically, poor tracer recovery in coarse sediment studies has posed a major problem, especially if particles have become buried. Poor recovery leads to problems in interpreting results in a statistically meaningful way. Tracers marked with surface paints, for example, are often difficult to relocate even if only buried to a few centimetres depth. Placement of pebbles in high-energy environments can lead to abrasion and loss of paint cover, while brightly coloured pebbles on a beach undoubtedly attract the attentions of young children often leading to redistribution in a manner no natural geomorphological process could explain. More recent developments in tracer technologies have used magnetic or radio-tracers emplaced in locally derived material so that even buried particles have a better chance of being found and do not attract unwarranted attention. It is essential that, whatever the marking system used, the tracer will last the lifetime of the project, that the identity is resistant to removal and that each individual particle has an unequivocal identity (Sear *et al*. 2000).

THE TRACER IS TRANSPORTED AND DEPOSITED IN THE SAME WAY AS THE MATERIAL BEING STUDIED

In many situations, coarse particles or other large material, can be identified by marking the surface with paint on which a code is added for later identification. Painted surfaces, however, may change the surface roughness and porosity characteristics of the original material leading to changes in buoyancy over relatively short timescales. Alternatives include the use of exotic (non-local) materials in order to aid identification and recovery but again differences in shape, density, porosity, buoyancy and/or surface roughness may lead to errors in experimental results. While magnetically or radio-tagged natural or 'exotic' particles have improved recovery rates, tracer physical characteristics may not exactly match those of the natural materials whose behaviour they are manufactured to represent. Fine sediment studies that use natural tracer characteristics may also prove problematic since erosion and sediment transport processes are particle-size selective. In consequence, concentrations of many natural tracers (e.g. heavy metals, nutrients, radionuclides, mineral magnetic signatures) increase with a decrease in particle size and an increase in particle specific surface area and for which a correction factor is often required. Additional considerations must be given to changing environmental conditions during transport. This is especially true in aquatic systems where changes in pH, redox potential (Eh) and salinity may drive adsorption and/or desorption reactions during sediment transport (Horowitz 1991).

LONG-STERM STORAGE

Once deposited, fine-grained aeolian, marine, fluvial and limnic sediments can undergo a range of complex transformations. Interruptions in loess or floodplain deposition, for example, often leads to periods of soil development while changes in the trophic status of lakes and estuaries may again lead to changes in pH and Eh driving the release of many natural tracers into the water column. It cannot therefore be assumed that natural tracer properties remain unchanged with time (e.g. Foster *et al*. 1998)

An example of fine sediment tracing for interpreting erosion processes

Erosion processes operating in a grazed paddock in New South Wales, Australia were analysed by Wallbrink and Murray (1993) using a rainfall simulator and two radionuclides that adsorb strongly to sediment and label different parts of the soil profile in different ways. Figure 168a and b show the vertical distribution of ^{137}Cs and ^{7}Be in a typical soil profile. ^{137}Cs has a thirty-year half-life and has been detected in the environment since atmospheric testing of thermonuclear weapons in the early 1950s (Higgitt 1995). Following the international treaty banning atmospheric nuclear weapons testing in 1963, the southern hemisphere has received little ^{137}Cs fallout. ^{7}Be is continuously produced in the upper atmosphere by cosmic ray bombardment but has a short half-life (53 days) in comparison with ^{137}Cs. Figure 168c shows how the ^{7}Be and ^{137}Cs activities of sediment produced by four different erosion processes (sheet, gully floor, gully collapse, rill erosion) could produce suspended sediment with different combinations of the two signatures. Gully collapse would result in low activities of both nuclides since the gully wall does not receive ^{7}Be fallout and the depth penetration of ^{137}Cs in the soil profile is less than 10 cm. The majority of the gully wall sediments are not labelled with either radionuclide. By contrast, sheet erosion would produce sediment with high activities of both radionuclides. Gully floor sediments would be labelled with ^{7}Be (since it is produced continuously) but would have no ^{137}Cs (since in this case the gully developed after 1963). Rill erosion would produce sediment strongly labelled with ^{137}Cs, but the shallow depth penetration of ^{7}Be in the soil profile would lead to a dilution of ^{7}Be activities as it mixes with sediment derived from deeper in the soil profile. By collecting suspended sediment during rainfall simulation experiments generating surface runoff, Figure 168d shows that the model of high ^{7}Be and ^{137}Cs activity is supported by the experimental results.

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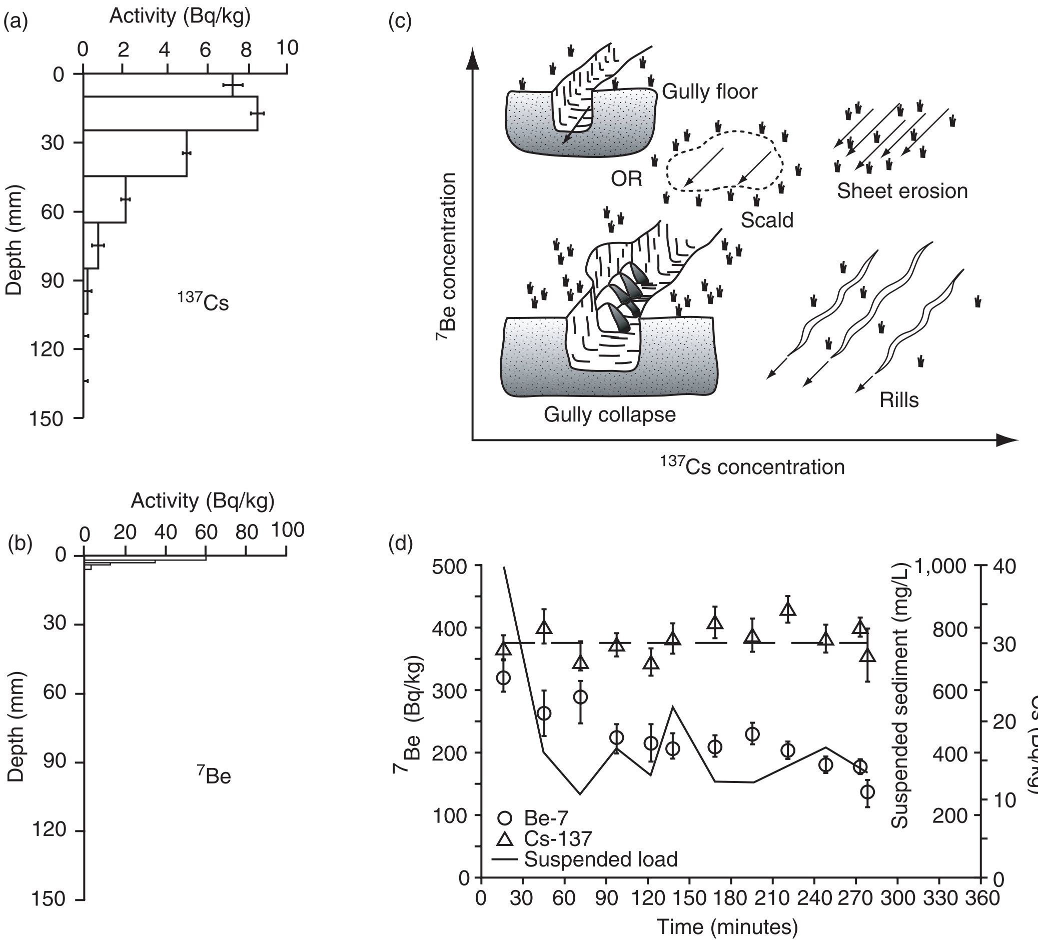

Figure 168 The distribution of (a) ^{137}Cs, and (b) ^{7}Be in the soil profile; (c) a model of how sediment would be labelled with ^{137}Cs and ^{7}Be depending on the erosion process involved; and (d) the results of a rainfall simulation experiment generating surface erosion

Source: Wallbrink and Murray (1993). © John Wiley and Sons Limited. Reproduced with permission.

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IAN D.L. FOSTER

TRANSGRESSION

A transgression is the movement of mean sea level (MSL) in an upward direction, while any lowering of MSL is called a regression. A movement of MSL, as the datum for wave and tidal activity, generates the potential for change in areas above and below the datum. Changes in the position of MSL are usually relative. A transgression is not solely due to MSL moving, as it is possible that the land surface is also moving relative to MSL. Long-term MSL changes can be due to both eustatic (changes in volume of sea water in the oceans) and isostatic (vertical movement of land) causes. A transgression is specified when the net balance between these two processes resolves in a relative rise in mean sea level (RSLR). Eustatic changes generally relate to climatic changes and their control on (1) unit ocean volume contraction/expansion due to atmospheric-ocean heat exchange (the steric effect); (2) evaporation rates of sea water; and (3) variations in the hydrological cycle by which free water is either locked up as terrestrial ice, or terrestrial ice melting to release water back into the oceans.

Quaternary transgressions usually relate to warming phases by which glacier ice melts and sea surface temperatures rise, e.g. the early to mid-Holocene in which atmospheric warming contributed to a major transgressive phase occurring in the mid-latitudes (Pirazzoli 1996). Although the Late Glacial was marked by rapid positive eustatic change, it was also a period of rapid upward crustal lift in the mid-upper latitudes, due to the release of terrestrial ice pressure through ice melting. The near-exponential early rapid isostatic rise in land level meant that eustatic change was more than exceeded, in effect inducing a regression or relative fall in MSL. This was only reversed as the isostatic rate diminished and the eustatic effect maximized some time in the mid-Holocene, to induce a well-recognized transgression (e.g. the Flandrian in north-west Europe). The British Isles can be crudely characterized as showing two main zones of response in the mid-Holocene; a northern zone in which the transgression peaked above present-day MSL (*c.*5–6 ka BP) and then switched to a regressive phase; and a southern zone showing a continuing transgressive phase to the present day but associated with a decelerating RSLR rate since the mid-Holocene. Local isostatic effects tend to modulate a regional eustatic signal and complicate the general picture at any one place. Contemporary concerns with the effects of accelerating global climate change are reflected in the forecasts for rising MSL over the next century, with RSLR up to five times the current rates being forecast. This will result in a major new transgressive phase, regardless of current crustal changes, as these modern eustatic changes are linked to global climate change, while the endogenetic changes required to generate crustal isostatic responses are unaffected.

Transgressions are a common element of geological-scale change. In recent decades, litho-stratigraphies have been interpreted via sequence stratigraphy by which rising and falling sea levels have been used to link subaerial erosion sources to submarine deposition sinks. The timescale of such deposition relates to transgressions ($>10^6$ years duration) as a consequence of: (1) slow orogeny and oceanic basin volume reduction; (2) massive sediment transfers due to landscape denudation causing isostatic readjustment; and (3) probable climate change. The speed of the Holocene transgression (millennia-scale) stands in contrast to these earlier episodes, and emphasizes the idiosyncratic conditions associated with Quaternary deglaciation-induced transgressions.

A transgression leads to inundation of the coastal zone, the onshore extent of which depends on the overall coastal slope angle. The sedimentary expression of the transgression is dependent on both the non-rectilinearity of the coastal slope setting the template for deposition, and availability of free sediment. The penetration distance of the transgression is not solely that of inundation, as the transgression carries wave and tidal activity that reworks the sediment mantle beyond the initial flooding limit. Bruun (1962) has attempted to specify the predictive relationship between RSLR and shoreline retreat, through what is now controversially termed the BRUUN RULE.

The concept of the 'Erosion Front' defines the spatial variation of coastal activity experienced from the quiescent leading edge of the front that moves up estuary, through the high energy breaking wave zone, and the trailing edge that has tidal reworking of near shore and beach face (Carter *et al*. 1992). The basal extent of the transgression is identified by an erosional surface also known as a ravinement. Much interest is centred on the way in which a transgression aids sediment-reworking into distinctive coastal morphologies, e.g. the development of both sand and gravel barriers and associated back-barrier intertidal sediment stores and marshes. Some dune investigators believe that transgressive conditions were requisite for the major coastal dunes associated with the Holocene, though a counter argument exists in which dunes are also associated with regressions. Local excess deposition can generate an apparent regressive signature by outweighing transgressive tendencies, though to achieve this sediment has to be derived from elsewhere alongshore leading to overall shoreline retreat. Key issues still to be resolved relate to transgressive migration rate and coastal morphology stability. Can saltmarsh growth match fast RSLR? Can barriers rollover and maintain longshore continuity under fast RSLR (Jennings *et al*. 1998)? These are critical issues for coastal communities dependent on maintaining sustainable natural coastal morphologies as coastal defences in the face of future extreme rates of RSLR.

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JULIAN ORFORD

TREE FALL

The falling over of trees during high winds is a significant factor in the translocation of material, churning of soils (Schaetzl 1986) disruption of strata, and the development of mound and pit micro-topography (Denny and Goodlett 1956). Deep-rooting trees affect topography to a greater degree than shallow-rooting trees, and when they are blown over leave deep pits and high mounds (Veneman *et al*. 1984). In mixed hardwood forests in Ithaca, New York, USA, mounds were typically 0.48–0.60 m high and pits 0.20–0.41 m deep (Beatty and Stone 1986). Windthrow appears to be a more prevalent phenomenon on deeper soils where there is a sharp contrast between the fine soil material and the underlying stony horizons (Boyd and Webb 1981). Tree fall is also a major contributor to the development of log steps and LARGE WOODY DEBRIS in forest streams (Marston 1982).

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A.S. GOUDIE

TRIMLINE, GLACIAL

Reconstruction of the geometry of former ice sheets and ice caps from geomorphological evidence is based on the identification of trimlines, defined by the maximum level to which GLACIERS or ICE SHEETS have eroded or 'trimmed' bedrock or debris in a valley hillslope (Ballantyne and Harris 1994). The sharpness of this boundary depends on the effectiveness of GLACIAL EROSION, the degree of frost WEATHERING after its formation, and the downslope MASS MOVEMENT during and after DEGLACIATION. The formation of weathering boundaries and glacial trimlines are open to four possible hypotheses (Figure 169):

1. Summit blockfields (see BLOCKFIELD AND BLOCKSTREAM) are formed by *in situ* rock weathering over a longer timescale than the Holocene, representing a glacial trimline marking the maximum altitude to which glacial erosion has eroded or 'trimmed' a pre-existing cover of REGOLITH/frost-shattered debris.
2. Frost weathering on high ground may reflect more pronounced breakup of rock at high

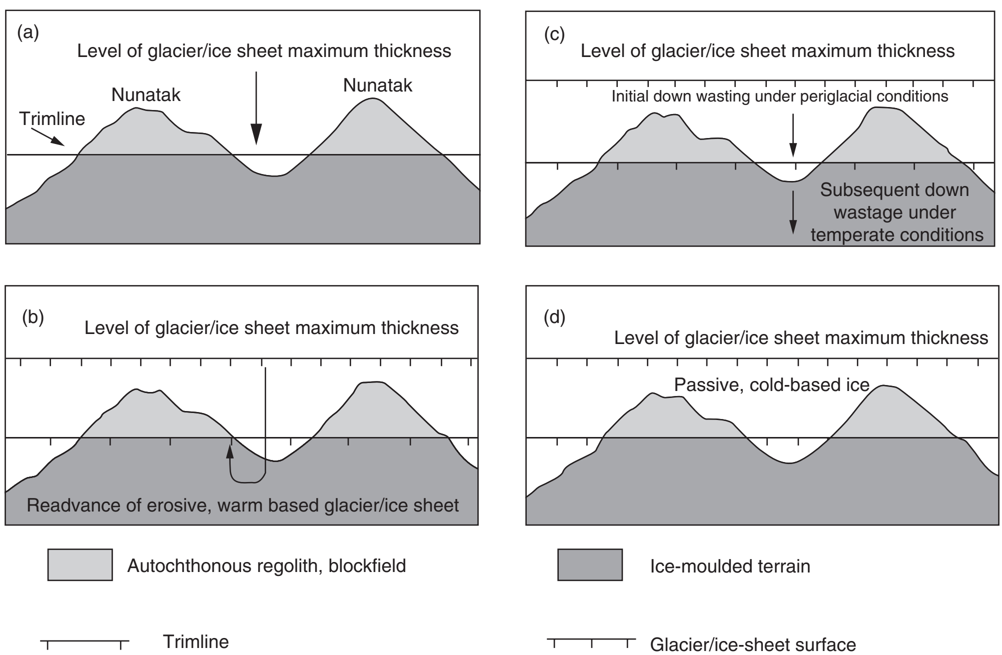

Figure 169 Four hypotheses of formation of glacial trimlines: (a) a glacial trimline representing the upper surface of an ice sheet at its maximum thickness; (b) a trimline cut by a glacial readvance during overall ice-sheet downwastage; (c) trimline formed during an initial period of ice-sheet downwastage under periglacial conditions; (d) weathering limit representing a thermal boundary between cold-based (temperature below pressure melting point) ice and a warm-based (temperature at the pressure melting point) ice within a former ice sheet (modified from Ballantyne *et al*. 1998)

altitudes under periglacial (see PERIGLACIAL GEOMORPHOLOGY) conditions, particularly during downwasting of an ice sheet and subsequent valley glaciation.

3 The initial stage of downwastage of ice sheets/ice caps may be accompanied by frost (see FROST AND FROST WEATHERING) shattering of exposed rock which ceases when the climate warms. The limit between frost-weathered and glacially abraded terrain represents the upper limit of glacier ice at the time of this thermal transition.

4 The weathering limit may represent a thermal boundary within a former ice sheet or ice cap, with *in situ* frost-weathered debris surviving under a cover of cold-based (basal temperature below the PRESSURE MELTING POINT) ice on high ground, whereas lower areas experience scouring by warm-based (basal temperature at the pressure melting point) glaciers.

Various approaches have been adopted to test these hypotheses. These involve analyses of weathering characteristics of bedrock and soils above and below the weathering limit/trimlines, and reconstruction of their altitudinal trend. Techniques employed are SCHMIDT HAMMER measurements of rock surface hardness, measurements of surface ROUGHNESS, measurements of differential relief of adjacent minerals, depth of open horizontal dilation (stress-release) joints using a graduated probe, studies of clay mineral assemblages and mineral magnetic signatures, and COSMOGENIC DATING.

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ATLE NESJE

TROPICAL GEOMORPHOLOGY

Tropical geomorphology has usually addressed conditions in the tropical forests and savannas, and although many areas within the tropics (23.5°N and S lat.) fall within the arid and semi-arid climates, these environments will not be included here. However, near-tropical climates extend to at least 30° lat., along the humid east coasts of Africa, Asia and the Americas and these areas have much in common with the tropics *sensu stricto* (Figure 170).

The study of geomorphology in the tropics cannot be separated from the wider history of the subject. During the early twentieth century most reports on tropical landscapes arose from geological investigations on behalf of former colonial powers: Britain, France, the Netherlands. Many observations concerned the extent and products of rock weathering (Falconer 1911; Scrivenor 1931). Reports also commented on unusual landscapes: extensive plains and great waterfalls in east and south Africa; high granite domes (or *inselberge*) in east and west Africa; the tower karst of South-East Asia. In addition, the widespread occurrence of *laterite*, long known from Buchanan's early work in India, attracted a lot of comment. Because the authors came from Europe, the exotic nature of tropical landforms was frequently emphasized.

In Davis's (1899) paper on 'The cycle of erosion' the evolution of landscapes was seen from the perspective of a humid temperate 'normality' and although Davis later addressed contrasts with semi-arid regions in western USA, the tropics were largely ignored. For fifty years the Davisian view was dominant in Britain, France and the USA and tropical landforms were seen as 'climatic accidents' (Cotton 1942) and peripheral to mainstream interests. In Germany, tropical landscapes were considered within a 'climatic geomorphology' (Thorbecke 1927; Büdel 1948), which led to specific hypotheses for landform

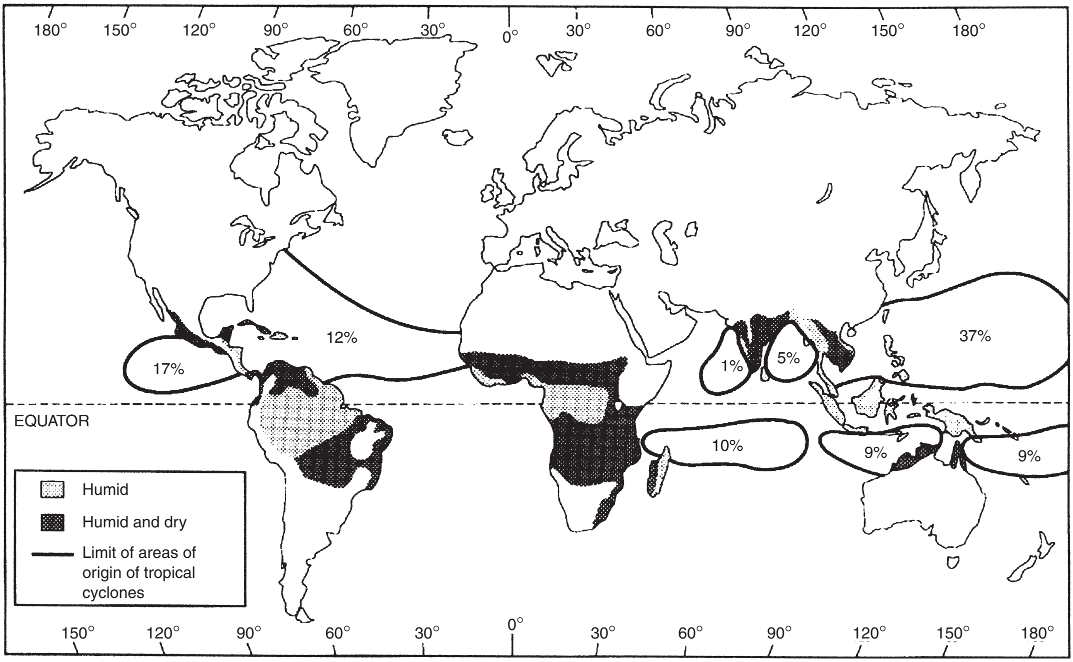

Figure 170 Tropical climates according to the Köppen system (modified by Trewartha), and occurrence of tropical cyclones shown as percentages (After WMO (1983))

development in the different climatic zones. In contrast, the parallel retreat of hillslopes, first proposed to explain some semi-arid landscapes in the USA, was extended by King (1953, 1957) as a universal model for hillslope development. King (1962) applied his ideas to all the Gondwana continents, and by implication, to most of the tropical world. This construct had enormous impact because it proposed a single hypothesis of slope development that was linked to 'continental drift' and the emerging theory of plate tectonics.

King's views contradicted the concept of a 'climatic geomorphology', but had little influence in Germany, where W. Penck (1924) had previously linked the extension of plains to tectonics, in a way that did not conflict with climatic control over geomorphic processes. In particular, the importance of a deeply weathered mantle in explanations of humid tropical relief was central to Büdel's (1957) hypothesis of *Doppelten Einebnungsflächen*, or 'double surfaces of levelling'. This paper led geomorphologists in the tropics to rediscover earlier accounts of weathering and relief development, especially in Africa (Falconer 1911; Wayland 1933; Willis 1936). These authors shared the view that plateau landscapes in the humid tropics were lowered incrementally by stripping and renewal of the SAPROLITE cover. Wayland (1933) called the resulting landscapes 'etched plains' (subsequently *etchplains*), the term later adopted by Thomas (1966, 1994) and Büdel (1982) (Figure 171). Falconer (1911) argued that much of the rocky relief in north Nigeria was due to differential weathering, and Willis (1936) argued that the prominence of granite monoliths or BORNHARDTs was due to repeated cycles of weathering and lowering of more susceptible rocks over geologic time. The isolation of bornhardts by selective weathering was also argued by Rougerie (1955) in Ivory Coast and in Brazil by Birot (1958) who stressed that *les dômes crystallines* were not inselbergs, as often described, but are found in zones of dissection. The importance of DEEP WEATHERING, followed by stripping was advocated by Ollier (1959, 1960) for Uganda, and new attempts were made to document the depth and distribution of the weathered mantle by Thomas (1966).

Linton (1955) drew on ideas about tropical weathering to explain 'the problem of tors' in

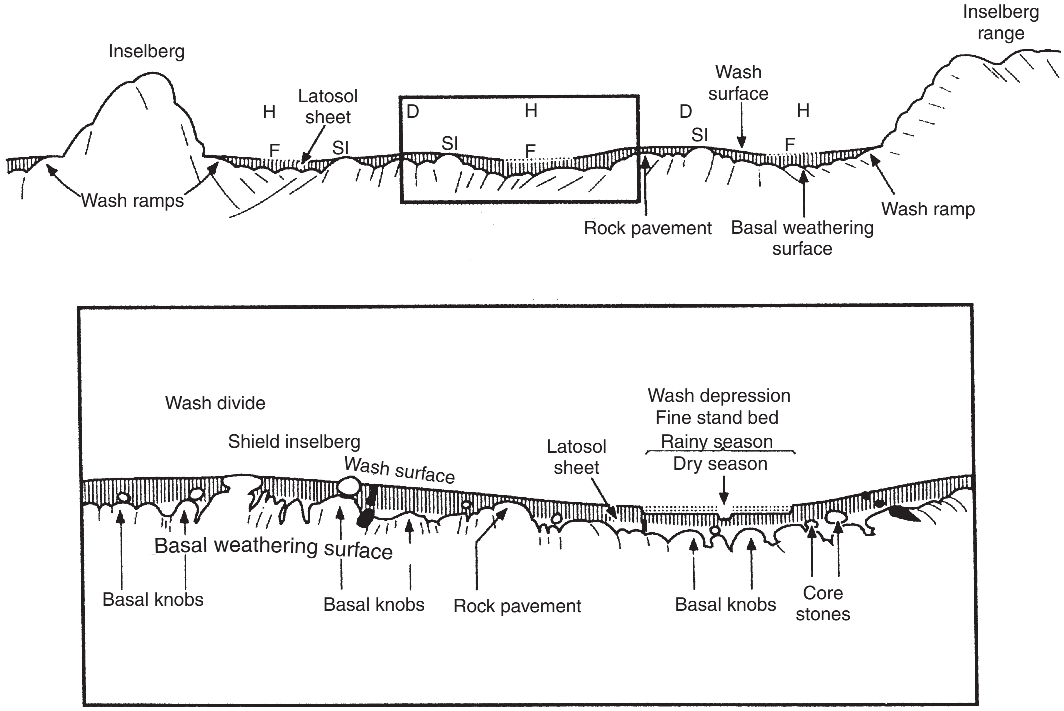

Figure 171 Characteristics of an active etchplain as described by Büdel (1982), based on studies of the Tamilnadu Plain in south India. H: wash depressions; D: wash divides; SI: shield inselbergs; F: fine sand in rainy season riverbed. Details of the wash divide and wash depression are shown in the expanded box. Lowering of the landscape is achieved by rock decay at the basal weathering surface and removal of sands and clays from the wash surface by seasonal runoff

Britain, leading to a clash with King. These ideas were applied to tor landscapes in North America, and to help explain the relief of glaciated areas (Feininger 1971). The notion of changing climates and the existence of 'relief generations' in world landscapes, was reviewed in a European context by Bakker and Levelt (1964), and became essential to a revised 'climato-genetic' geomorphology (Büdel 1968). Stoddart (1969) in an influential review of climatic geomorphology was scathing. He thought that the concept had its roots in Davisian theory, and considered that observations of landform differences between climatic zones were 'methodologically trivial' (p. 213). This view was widely accepted by those who pointed out that the same physical laws apply everywhere, and that the processes of chemical weathering and water flow are similar in all climates.

Although the Davisian paradigm prevailed for half a century, by 1950 proponents of the new 'process geomorphology' were directing attention away from geomorphic cycles in order to understand the mechanisms of landform change, rediscovering the work of G.K. Gilbert (1877, 1914). In the tropics, Branner (1896) had reported from east Brazil the roles of high rainfall and the presence of organic acids in tropical weathering; recognized the involvement of kaolinized saprolite in generating landslides under a natural forest cover, and the importance of soil fauna such as termites. By the 1950s geomorphologists in the tropics had begun to study these processes. White (1949) emphasized the importance of landslides in Oahu (Hawaii), while in West Africa studies of chemical dissolution of rocks, runoff and erosion arose from French research (Corbel 1957; Fournier 1960; Rougerie 1960). In South America, Tricart (1956, 1959) focused attention on fluvial processes, and the impacts of climate change. His 'Le modelé des régions chaudes, forêts et savanes' (Tricart 1965) was the first 'modern' study of tropical geomorphology. Tricart and Cailleux (1965, English trans. 1972) also published their 'Introduction à la géomorphologie climatique' in the same year complaining (pp. xiii) that 'the study of tropical geomorphology... has been delayed by the uniformist ideas of "normal erosion"', which led to the systematic neglect of the tropics. This neglect was compounded by problems of access, fear of disease and, in the rainforests, a lack of inter-visibility.

By this time development involving new roads and construction projects had exposed the reality and extent of deep weathering in the humid tropics. In Hong Kong, detailed studies of weathering (Ruxton and Berry 1957) and analyses of intense tropical rainfall (Lumb 1975) showed unequivocally the importance of both to an understanding of landslide occurrence. The devastating effects of tropical storms was also reported from Rio de Janeiro (Jones 1973). On the other hand, the high permeability of many tropical soils was documented by Nye (1954, 1955) and subsequently by many others, who found that infiltration rates in *ferrallitic* soils (*oxisols, ultisols*) over gneissic rocks exceeded 200 mm h^{-1}, greater than expected rainfall intensities. Nye did not observe surface runoff in his experiments at Ibadan, Nigeria, but others have measured significant runoff on moderate slopes at intensities of 60–70 mm h^{-1} (Morgan 1986). Outside the equatorial zone, the seasonal concentration of tropical rainfall combined with the high intensity of individual falls convinced many writers of its potential for serious soil erosion (Fournier 1960; Roose 1981). Studies of sediment yield have recorded great variation (Douglas and Spencer 1985; Thomas 1994). On undisturbed plots under rainforest, yields of 0.1–10 t km^2 yr^{-1} have been recorded, but small erosion plots left bare of plant cover can return sediment yields of 5,000 t km^2 yr^{-1}. Small catchments in deforested highland areas have recorded from 1,500–3,000 t km^2 yr^{-1}, but larger catchments, which have floodplains for major sediment storage, produce far less sediment per unit area and the average for Africa is just 35 t km^2 yr^{-1}, rising to 380 t km^2 yr^{-1} in Asia. Rates are more influenced by catchment relief and rainfall amount than by other factors (Milliman and Syvitski 1992).

Tropical rivers are as varied as in other regions. Large rivers traversing the plateaux and plains of the southern (Gondwanaland) continents appear distinctive because of the fine calibre of their bedload (usually sand) and the interruption of their thalwegs by rock outcrops, which create rapids and waterfalls. Opinions formed on the basis of these characteristics hold that the thoroughness of tropical weathering reduces most rocks to sand and clay, or causes their total dissolution. According to writers such as Büdel (1957), Birot (1958) and Tricart (1965) this leads to a lack of abrasive tools for channel cutting; valleys are consequently open with shallow slopes, and river channels are interrupted by resistant bands of rock, which they do not erode effectively. However, wherever tropical rivers traverse hill ranges or fault scarps coarse bedload material becomes abundant and deep valleys and gorges result. Deep valleys are also cut into saprolite.

Great seasonal variations in river discharges are found in monsoon areas, and appear to lead to distinctive channel morphologies. The Auranga River, India was described by Gupta and Dutt (1989) as having a braided sandsheet with shallow (30 cm) channels in the long dry season, but is converted during the monsoon rains (1,500 mm in 4 months) to a wide meandering channel with sand and gravel point bars. By contrast rivers draining the equatorial Papua New Guinea mountains have a very low seasonal variability in flow and are subject to almost instantaneous runoff from steep slopes kept in a near saturated state by frequent rainfalls. Rainfall intensity is lower in this environment and most sediment reaches the river via frequent landslides (Pickup 1984; Pickup and Warner 1984).

Justification for a 'geomorphology of the tropics' lies in the balance of processes and their outcomes in terms of weathered materials and soil properties, erosional forms and sediments. Paradoxically, the major rock landforms, characterized by the granite domes or *bornhardts* that previously received such emphasis, owe more to petrology and structure than to specific climatic parameters. Nonetheless many illustrate the principles of selective deep weathering and stripping of regolith over very long time periods. Although it has been argued that lateritic (ferrallitic) weathering is more a product of time than of climate (Taylor *et al.* 1992), these factors are not independent. Advanced weathering occurs beneath ancient landsurface in non-tropical Australia, but is found more widely in the humid tropics, where some Neogene formations are also affected (Thomas *et al.* 1999). Weathering rates are, however, greatly influenced by water movement through the profile, so that dry climates lacking a seasonal water surplus have a much reduced potential for deep and advanced weathering. Bourgeon (see Pedro 1997) showed that deep clay-rich kaolinitic 'alterites' pass into shallow, sandy materials (arènes, grus) along an east–west transect of southern peninsular India ($P > 2{,}000\,mm\,y^{-1} - 700\,y^{-1}$), where smectite increases in proportion to kaolinite as the climate becomes drier.

Properties of the saprolite are fundamental to the understanding of tropical soils, as much for engineering as for agriculture. The permeability of tropical saprolite is increased by Fe_2O_3 adhesion to kaolin platelets, which produces larger aggregates. But these form 'cardhouse' fabrics that can collapse under loading, causing settlement of building structures. On the other hand, 2:1 lattice clays (smectites) found in seasonal climates, experience remarkable expansion as water content increases. This reduces permeability and causes 'heave' in many drier areas, forming GILGAI. Iron enrichment of the saprolite can arise from the flux of Fe^{2+} ions that are subsequently oxidized to Fe_2O_3. The Fe is mobilized as Fe^{2+} under conditions of low pH or by chelation involving organo-metallic complexes. This is likely to occur within poorly drained depressions in humid climates or under forest. But the fixation of Fe needs a rise in pH accompanied by drying of the deposit. The term *laterite* is often used to describe materials with high Al/Fe sesquioxide content (see Aleva 1994). It is a form of FERRICRETE that may also form in transported sediments; it becomes indurated and resistant to erosion if subject to wetting and drying, and may appear as hill cappings and valleyside benches in the landscape. Seasonality of rainfall contributes to this process, but in many cases the duricrusts have formed due to falling water tables, resulting from climate change and/or dissection. Beneath well-drained sites in the rainforest Fe does not become concentrated, but Al_2O_3 rich *bauxite* (mainly gibbsite) forms often as irregular nodules. Beneath undisturbed forests, the dominant exports are ions in solution, leading to chemical sediments offshore (Erhart 1955). In equatorial environments, where water tables are high, almost complete dissolution of silicate minerals is possible, leading to the formation of quartz-dominated 'white sands' (Thomas 1994).

In the mid-Miocene, there was a marked reduction of rainfall in the southern tropics, probably associated with the growth of the Antarctic ice sheet. Large sandsheets, such as the Kalahari Sands that penetrate the Congo Basin rainforests from the south may have originated at this time, and it has been argued that weathering systems were 'switched off' by this event, but many equatorial areas were humid throughout the Neogene. Quaternary climate change in the tropics, involved temperature reductions of 5–6 °C at the Last Glacial Maximum (LGM), and lowland areas experienced rainfall reductions of 25–50 per cent, and increased seasonality. Rainforests were converted to deciduous woodlands and/or savanna mosaics over large areas. Many small streams in the tropics left few traces of sedimentation during the LGM; other formerly meandering rivers became braided. Alluvial fans formed along many escarpments, probably due to diminished stream power and increased sediment load. Today, tropical landscapes are sensitive to environmental changes because the thick saprolites and stores of Quaternary sediment are subject to rapid erosion under intense rainfalls, if vegetation is cleared.

Accidents of history led to tropical areas lying beyond the immediate experience of geographers and geologists resident in the temperate zones of Europe and North America. This was not the case for the arid, glacial and periglacial environments, and it has led to persistent neglect of tropical geomorphology. Those who argue against the recognition of rainforests and savannas as distinctive environments for geomorphology fail to recognize the linkages between abundant rainfall, warm soil conditions, the productivity of humid tropical ecosystems, and the formative environments for soils and sediments. Ecologists and pedologists have long recognized that system outcomes depend on rates of biogeochemical cycles: of growth and decay; leaching and accumulation. Under tropical forests, rates of weathering and bioturbation can compensate for soil loss and erosion. Little known biota enter into weathering processes and organic acids can alter classic geochemistry to enhance the mobility of many cations. The frequency of intense rainfall in many areas is greater than in non-tropical climates, and the incidence of cyclones with high wind velocities is regionally important. These characteristics increase the hazards of flooding, inundation and landsliding, even in tectonically quiescent areas.

If geomorphology had developed from a tropical perspective it might have emerged with a different emphasis. But our views are also influenced by the available database, and the most serious problem in relating a geomorphology of the tropics to studies elsewhere remains the paucity of data about conditions in tropical environments.

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MICHAEL F. THOMAS

TROTTOIR

Trottoirs (surf ledges) are narrow, subhorizontal erosional SHORE PLATFORMs with a veneer of Vermetid gastropod tubes (see VERMETID REEF AND BOILER) and encrusting coralline algae, and surfaces that often consist of tiers of VASQUEs. Trottoirs occur at, or a little below, mean sea level in tropical seas and in the warmer parts of the southern and eastern Mediterreanean. They have been attributed to CORROSIONal cliff erosion in the spray zone and possibly the protection afforded to the wave-battered seaward edges of the platforms by Vermetid tubes, algae and other organic encrustations. On Curacao and other islands in the southern Caribbean Sea, trottoirs, up to 10 m width, are restricted to exposed areas where there are thick encrustations, and they are replaced by notches (see NOTCH, COASTAL) in less exposed areas. It has been suggested that trottoirs may eventually attain a state of DYNAMIC EQUILIBRIUM with the processes operating on them as their increasing width reduces the rate of cliff erosion.

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ALAN TRENHAILE

TSUNAMI

A tsunami is a Japanese word that describes a 'harbour wave' and it is used within the scientific community to describe a series of waves that travel across the ocean with exceptionally long wavelengths (up to several hundred kilometres between the wave crests in the open ocean). As the waves approach a coastline, the speed of the waves decreases as they are deformed within shallower water depths. During this process of wave deformation, the height of the wave increases significantly and as the waves strike the coastline they often cause widespread flooding across low-lying coastal areas and on many occasions cause loss of life and widespread destruction of property. Tsunamis are frequently described in the media as tidal waves. However this view is completely wrong since tsunamis have nothing to do with tides or weather. Most tsunamis in the world occur in the Pacific region. For example, during the past ten years, damaging tsunami floods have taken place in Nicaragua (1991), Flores, Indonesia (1992) and Okushiri Island, Japan (1993). The most recent destructive tsunami disaster took place in Papua New Guinea (1998) where several thousand people lost their lives.

Frequently tsunamis are described as seismic sea waves that are produced as a result of a sudden displacement of part of the seafloor. Usually a seismic disturbance associated with an offshore earthquake will cause rupturing of the ocean floor and when this happens the overlying water column is disturbed. It is during this phase of disturbed motion that tsunamis are often generated. Underwater earthquakes in tectonically active areas of the world (e.g. the Pacific rim) are the most common cause of tsunamis. For example, as recently as May 1960, a large underwater earthquake took place off the coast of Chile. The tsunami travelled across the Pacific Ocean. After twelve hours it struck the coastline of the Hawaiian island chain, while after a further twelve hours it reached the coast of Japan where it destroyed all in its path. A huge tsunami was produced as a result of a large earthquake beneath the seafloor west of Portugal on 1 November AD 1755. One observer who witnessed this tsunami was William Borlase who observed its arrival on the shore of Mounts Bay, Cornwall. He noted:

> the first and second refluxes were not so violent as the third and fourth (*tsunami waves*) at which time the sea was as rapid as that of a mill-stream descending to an undershot wheel, and the rebounds of the sea continued in their full-fury for fully two hours . . . alternatively rising and falling, each retreat and advance nearly of the space of ten minutes, till five and a half hours after it began.

Another observer of this tsunami described how, at Larmorna Cove, Cornwall:

> the sea on this occasion rushed suddenly towards the shire in vast waves, with such impetuosity, that large rounded blocks of granite from below low-water mark were swept along like pebbles, and many of them deposited far beyond high water mark. One large block, weighing probably 6 or 8 tons, was borne repeatedly to and fro several feet above the level of high water, and at length deposited about ten feet above that level in the stream, where it still lies.

Tsunamis can also be generated by underwater landslides and by landslides occurring above sea level and moving downslope into the sea (see SUBMARINE LANDSLIDE GEOMORPHOLOGY). The occurrence of such slides is quite common on a geological timescale. In 1929, an offshore earthquake led to widespread sediment slumping and the generation of turbidity currents across the seafloor off the Grand Banks of Newfoundland as well as a tsunami that locally reached up to +30 m at the coast. In the North Atlantic region a very large tsunami was generated approximately 8,000 years ago by one of the world's largest underwater landslides (the Second Storegga Slide). The tsunami caused flooding along parts of the Norwegian coastline up to levels +20 m above sea level and along the UK coastline where the highest flood levels reached up to +6 m above sea level. One of the most widely described processes of submarine slide generation is through the release of methane gases (clathrates) contained within ocean-floor sediments. It is believed that the sudden release of such gases can cause local slumping and sliding. The second process arises from low-magnitude earthquakes which in themselves have no destructive effect, but which are of sufficient intensity to induce shaking of seafloor sediments thus causing downslope slumping and sliding of sediment. Generally speaking, underwater landslides generate tsunamis with much less energy than those produced by earthquake-triggered faulting on the ocean floor. The size and energy of landslide-generated tsunamis decreases rapidly with increased distance from the source area.

Occasionally, tsunamis can be generated by the impact of meteorites onto the ocean surface. Perhaps the most famous tsunami associated with a large meteorite impact was that which took place *c.*65 million years ago. The so-called K-T impact, usually associated in the geological literature with the extinction of the dinosaurs, also created a huge tsunami. Traces of the tsunami waves can be found in areas of Mexico and Texas. Most authoritative accounts consider that, given the proportion of the Earth's surface area represented by the Pacific Ocean, a meteorite capable of striking the Pacific Ocean and generating a tsunami is in the order of 1:400,000 years.

Tsunamis can also be produced through the collapse of a volcanic crater into the sea during a major explosive volcanic eruption. Although such a phenomenon may have occurred *c.*18,000 years ago in the case of the volcanic eruption of Santorini in the Aegean Sea, such events are extremely rare. Similarly, tsunamis can be produced as a result of the collapse of the flank of a volcano into the sea. Such tsunamis may be generated by rockfalls, rockslides, rotational slips or debris flows that originate on steep slopes and move into the sea. Such an event happened during the recent eruptions on the island of Montserrat. Some authors have proposed that certain hillslopes on the flanks of some of the now dormant volcanoes in the Canary Isles may have collapsed into the sea in the past and produced large tsunamis.

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ALASTAIR G. DAWSON

TUFA AND TRAVERTINE

Tufa and travertine are terrestrial freshwater accumulations of calcium carbonate, whose formation often involves a degree of organic involvement. The names tufa and travertine can be used synonymously, but often tufa is taken to refer to a softer, more friable deposit whilst travertine refers to a harder, more resistant material frequently used as a building material. Tufa derives from tophus or tufo, used in Roman times to describe crumbly white deposits (Ford and Pedley 1996; Pentecost 1993). Travertine derives from the Latin *lapis tiburtinus*, or Tibur stone, and was originally used to describe the massive, hot spring deposits around Rome (Pentecost 1993). Some travertines are of hydrothermal origin and thus contain very limited plant material, whilst tufa and travertines precipitated by cool water typically contain the remains of micro- and macrophytes, invertebrates and bacteria. Different usages of the terms tufa and travertine are found in different countries, and there is as yet no standard international terminology although Ford and Pedley (1996) and Pentecost and Viles (1994) suggest two alternative schemes. Tufa and travertine are distinguished from CALCRETE, SPELEOTHEMS and STROMATOLITES by their environment of formation, but they share many similar characteristics and may grade into one another in some circumstances.

Tufas and travertines form in freshwater environments where thermodynamic and kinetic characteristics favour the precipitation of calcium carbonate from carbonate-rich waters. Such conditions arise where carbon dioxide is removed from the water through turbulent degassing, evaporation or biological uptake. Suitable conditions for precipitation of tufa and travertine are often found in or near KARST areas where dissolution of limestone provides high levels of dissolved carbonate, or where thermal waters rich in carbon dioxide originate in areas of recent volcanic activity. Despite difficulties in obtaining reliable dates of tufas using ^{14}C, most tufas and travertines which have formed from cool water appear to originate from the late Quaternary to early Holocene. This is certainly true in Britain and much of Europe, although in some parts of the world such tufas and travertines are still forming rapidly today. There has been much debate over the major controls on tufa and travertine formation and in particular the role of organisms in their genesis. Certainly, aquatic plants and micro-organisms can aid deposition of tufa – by providing precipitation nuclei, by removing carbon dioxide from the water and perhaps also by direct precipitation of calcium carbonate – but in some environments physico-chemical controls on precipitation outweigh any biological involvement. In many fluvial tufas within forested catchments inorganic modes of precipitation dominate in the turbulent upper reaches, whilst further down tree debris contributes to barrage architecture and macro- and microphytes, and even insect larvae can aid deposition (Drysdale 1999). Hydrothermal travertines are precipitated through largely inorganic processes, although Robert Folk (1994) has suggested that tiny nannobacteria may also play a key role.

Tufa and travertine are found on all continents except Antarctica and there are several very impressive deposits around the world, such as the barrages of the Plitvice Lakes in Croatia, the travertine complex of Antalya in south-west Turkey (Burger 1990) and the Huanglong ravine terraces in north-west Sichuan, China. Good summaries of the occurrence and nature of the major tufa and travertine deposits in Britain, Europe, China and the world are provided by Pentecost (1993), Pentecost (1995), Pentecost and Zhaohui (2001) and Ford and Pedley (1996) respectively. Many occurrences of large tufa and travertines have still not yet been properly documented in the international literature, such as the extensive deposits within the arid Naukluft area of south-central Namibia (Plate 141).

Tufas and travertines form in a range of different geomorphic settings, including fluvial, lacustrine, paludal and spring environments. Within rivers, tufas and travertines can form spectacular barrages often with waterfalls cascading over them, and with clastic tufa accumulating behind the barrage. In some fluvial environments, suites of lakes become created between barrages. Much smaller accumulations of tufa and travertine also occur in many streams, producing fluvial crusts and oncoids. In lacustrine environments, similar oncoids and crusts occur as well as larger reef-like accumulations. In marshy environments, low relief muddy tufas tend to develop, often mixed with marls and chalks. Around springs, mounds and terraces can develop, and where springs debouch on steep slopes such deposits can form huge prograding cascades. Tufas and travertines can produce great geomorphological change within a catchment, as they influence river flows through the creation of barrages and can armour slopes and change the course of springs.

Plate 141 A large cone-shaped tufa, *c.*400 m across and nearly 100 m high, developed on the edge of the Naukluft Mountains at Blasskrantz in central Namibia. It has formed where a seasonal stream cascades over a steep slope, permitting degassing to take place

Plate 142 A small tufa dam formed across a creek in the tropical Napier Range of the Kimberley district, northwestern Australia

The fabric of tufas and travertines reveals much about their mode of formation, and can also contain useful palaeoenvironmental information. Where organic influence in tufa and travertine precipitation is debated, the petrology of the deposit can provide helpful insights. Martyn Pedley has provided a useful concept of tufa and travertine as phytoherms, with different facies found in different parts of the deposits and a clear organic role played by biofilms in the formation of many zones (Pedley 1992). Some tufas and travertines possess clear laminations which can be of great use in environmental reconstruction. Tufas precipitated in association with cyanobacteria from the genus Phormidium for example, tend to display seasonal banding with alternate sparitic, light layers (representing inorganic deposition in autumn and winter) and micritic, dark layers (formed as a result of cyanobacterially mediated deposition in spring and summer).

The fact that many major cool water tufas and travertines date from the late Quaternary and early Holocene has led several authors to conclude that there is a strong climatic influence on their formation (see for example Goudie *et al.* 1993). Alternatively, or as well as such climatic control, human impacts may also be responsible for a recent decline in the accumulation of tufas and travertines in some parts of the world. Isotopic and trace element contents of tufas and travertines can be used to reconstruct palaeoenvironmental conditions within different parts of dated sequences, thus throwing light on both the environmental conditions and their influence on tufa deposition rates. Andrews *et al.* (1997) illustrate the utility of stable isotopes of oxygen and carbon in reconstructing palaeoenvironmental conditions in Holocene tufa deposits, and Matsuoka *et al.* (2001) show how trace element contents within laminated tufas can provide high resolution records of climate and catchment conditions. As Ford and Pedley (1996) conclude 'tufas have an untapped potential to provide the best land-based opportunity for accessing shorter-term Holocene environmental change'.

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HEATHER A. VILES

TUNNEL EROSION

Tunnel erosion is an insidious form of land degradation that is initiated in subsoil and/or substrata and remains inconspicuous until considerable damage has occurred. This type of water erosion is found in earthworks as well as hillslopes and in the latter case refers to the hydraulic removal of subsurface material causing the formation of underground channels in the natural landscape (Boucher 1990). Tunnelling takes place primarily when the shear stress applied by flowing water enlarges an existing macropore or passageway (Bryan and Jones 1997). Corrasion of the tunnel perimeter by ensuing flow and the impact of vehicular, livestock and/or human traffic cause localized collapse of the land surface, producing sinkholes which can entrain surface runoff. The natural bridges of soil remaining between sinkholes are destroyed when the cavities merge, leaving a gully. Debris fans are most conspicuous on hillslopes, beginning from points on the land surface where the translocated sediment is debouched, and consisting of relatively coarse material which can no longer be transported by the hydraulic head. Finer particles are washed further downslope. The initial length of tunnels is uncertain but they may be a series of interconnected macropores and can extend to several hundred metres. The diameter of these features typically ranges between several centimetres and a few metres, the latter proving to be most frequent in semi-arid environments.

Tunnelling has been recorded under a wide range of climates in many different materials on landscapes subjected to diverse land-use histories, but owing to occurrence in uninhabited forest in Papua New Guinea and Eocene palaeopiping in the USA, it should be seen as a naturally occurring process. Sources of tunnelflow include rainfall, snowmelt, irrigation water and ground water. Macropores, desiccation and structural cracks, surface depressions, decayed tree roots and the burrows of insects, crabs, moles, rodents and rabbits have been shown to be points where surface water can infiltrate directly to the subsoil. The materials eroded have included soils that remained highly stable on wetting in the UK (Jones 1981), glaciolacustrine silts and permafrost in Canada as well as ignimbrite and pumice in New Zealand. Many reports of tunnelling in semi-arid climates, especially southeastern Australia, have been associated with nonsaline sodic texture-contrast soils that slake and/or disperse readily on contact with water. An important characteristic of tunnel erosion is the requirement for a layer(s) of material of relatively low permeability which act(s) as a barrier to further vertical percolation. This material can be breached by cracks, joints, tree roots and well-connected macropores. A hydraulic gradient is required to generate flow and the most common outlets occur on the hillslope proper where a hydraulic head can be seen, or comprise various types of free faces such as gully walls and stream banks.

Whilst most hydrologic research has been conducted in humid climates where ephemeral, seasonal and perennial systems are observed (e.g. Wales, UK, Jones 1981, Bryan and Jones 1997), few data are available for semi-arid environments. On grazing land in Victoria, Australia, flow in a shallow tunnel system (generally less than 1 m deep from the soil surface) responded rapidly to rainfall, and it was clear from the typically short lags between peak rainfall and peak runoff that the soil matrix had been largely bypassed. The recession was also rapid and ground water was not a component of the hydrograph (Boucher 1995). Similar characteristics were documented for shallow tunnels in an area of badlands in western Canada (Bryan and Harvey 1985) and a loess plateau in northern China. However, at least partly owing to internal collapse, the relations were more complex for deep-seated systems which were 20 m to 30 m deep in the badlands area, whilst the inlets ranged between less than 0.5 m and over 20 m in both depth and diameter in the latter catchment. The proportion of discharge passing through tunnels to the stream was estimated as up to 10 per cent and 80 per cent respectively. Therefore, tunnel erosion can be a rapid and significant form of soil and water loss, and the economic implications need investigation. Generally, reclamation of land in southeastern Australia should combine deep ripping of soil to destroy the established flowpaths, chemical amelioration of dispersive soils with gypsum in order to displace the sodium and generate an electrolyte effect, and revegetation with grasses which use up excess water.

Bryan and Jones (1997) suggested that tunnelling and piping are distinct processes which are often difficult to distinguish in practice, and that all subsurface erosion is usually combined as piping. However, from research in Australia and China it appears that the terms are still used interchangeably.

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SEE ALSO: pipe and piping

STUART C. BOUCHER

TUNNEL VALLEY

Tunnel valleys are examples of large-scale, erosional glacial landforms that are formed subglacially. Morphologically, tunnel valleys are overdeepened, elongate depressions with steep, often asymmetric sides, cut into bedrock or unconsolidated glacigenic sediment. A characteristic feature is an undulating long profile, which climbs over bedrock rises and contains overdeepened areas along its floor. Long reaches of the flow path can be against the regional gradient. In dimensions, tunnel valleys can reach up to 4 km in width and over 100 km in length, and depths of erosion can be as great as 400 m below sea level. Commonly they trend oblique to the modern drainage, and many contain subglacial landforms such as drumlins, eskers or gravel dunes. In plan view, tunnel valley systems vary from individual, straight segments, to integrated, anastomosing networks of sinuous valleys. Sedimentary infills of tunnel valleys are diverse, and a wide variety of sediments associated with different depositional environments (glaciterrestrial, glacimarine, glacilacustrine and temperate) have been recorded, which reflect changing conditions during, and subsequent to, valley formation. Sediment gravity flow deposits and glacifluvial sands are particularly common.

There are three main theories of tunnel valley genesis. The first argues that they form time-transgressively, by subglacial meltwater erosion during deglaciation, at or close to the ice margin. The second theory also interprets tunnel valleys as the product of subglacial meltwater erosion but argues that the discharges involved were catastrophic channelized floods, and that tunnel valleys within anastomosing networks form synchronously. Finally, the third theory argues that tunnel valleys cut into unconsolidated sediment are due to creep of deformable sediment into a subglacial conduit from the sides and from below. This material is then removed through the conduit by subglacial meltwater.

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COLM Ó COFAIGH

TURBIDITY CURRENT

Turbidity currents or flows are a type of density flow in which movement on a slope occurs due to changed density between a local fluid and a surrounding fluid (Simpson 1987). In a turbidity current it is the suspended particles that cause the density of the flow to be greater than that of the surrounding fluid. The result of this is that the turbulent suspension moves down any local or regional gravity slope.

Turbidity currents in water can originate in a number of ways. One mechanism is for them to originate as sediment slides and slumps caused by scarp or slope collapse (e.g. when a slope is disturbed by earthquake shocks). Another mechanism is direct underflow of suspension-charged river water in so-called hyperpychnal plumes. These can occur during snowmelt floods in steep-sided FJORDs, in front of RIVER DELTAs, and in river tributaries whose feeder channels have extremely high loads of suspended sediments. A third mechanism is the collection of sediment by longshore drift in the nearshore heads of SUBMARINE VALLEYs and canyons (Leeder 1999).

Turbidity flows are not restricted to water. As powder snow avalanches demonstrate, for example, dense suspensions in air may flow downslope.

Submarine turbidity currents can be up to several kilometres wide and several hundreds of metres thick. They can travel as far as 4,000–5,000 km. Erosion can occur at the base of a flow, producing various scour or sole marks including mega-flutes. Deposition from a waning turbidity current produces sediment accumulations called turbidites. Associated with these may be large-scale asymmetric gravel waves and macrodunes and channel levees (Stow 1994).

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A.S. GOUDIE

TURF EXFOLIATION

A denudation process that is particularly prevalent in periglacial areas and leads to the destruction of a continuous vegetation cover through the removal of soil exposed along small terrace fronts. Among the processes that lead, possibly synergistically, to this phenomenon are needle-ice (pipkrake) action, desiccation, rain wash erosion, zoogeomorphic activities and aeolian deflation. It appears to be especially pronounced in high alpine regions, not least in lower latitudes (Grab 2002).

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A.S. GOUDIE

TURLOUGH

A seasonally filled KARST depression, dependent upon water table fluctuations and tidal effects. Turloughs occur in the glacially modified carboniferous age limestones of central and western Ireland, with the term derived from the Irish *tuar loch* meaning dry lake. In times of decreasing water table (dry seasons) the level of the turlough drops, with the contents draining through the porous parts of the basin floor and via connections to swallow holes. In contrast, times of high rainfall (wetter seasons) will produce a high lake level. Theories of development (see Coxon 1986) are based upon interactions of a karstic landscape formed in the Tertiary and subsequent glaciations, with turlough form dependent upon this history. The flooding regime in turloughs varies considerably, as a result of size, depth, local water table, tidal regime and soil conditions. They can be composed of sand, clay, silt, diamicton, peat or marl, or various combinations (Coxon 1986). An example of a turlough is the Carren Turlough of the Burren Region, County Clare, Eire.

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STEVE WARD

U

UNDERFIT STREAM

A stream is underfit when it is much smaller than the size of its valley. Dury (1964) defines this fluvial geomorphological feature as that of a present-day stream flowing in an alluvial plain and describing free meanders far less ample than those of the enclosing VALLEY MEANDERS, which are frequently ingrown meanders of former large streams. Dury (1965) found that the ratio between the bed width of former and present-day channels (*W/w*) gives an index of underfitness, and that this ratio averages 10/1 in lowland England and near the former ice fronts in Wisconsin (USA), 5/1 and about 3/1 in the Ozarks. He also proposed an alternative index, that given by the wavelength ratios (*L/l*).

Because there is a relationship between stream size and size of meanders, it has to be assumed that some cause has operated to reduce the present stream size significantly below its former values. Davis (1913) had defined underfit streams and related their origin to the capture process taking place during the competitive development of rivers. The beheaded stream underwent shrinkage and reduced the size of the meanders which it had formerly processed. Other hypotheses proposed by subsequent researchers include: underflow through alluvium, deep percolation or cessation of meltwater discharge, overspill from glacial lakes or glacial meltwater and tidal scour. During the 1950s Dury undertook a series of field investigations of subsurface bottoms of valleys occupied by manifest underfits in several localities of the English plain in order to verify the hypothesis that, if valley bends are authentic former meanders, former large channels should have been associated with them. He demonstrated the presence of large meandering channels, up to ten times as wide as the existing channel, cut in bedrock and complete with pool and riffle sequences. Further tests on deep buried channels in the driftless area of Wisconsin, in France and elsewhere confirmed his hypothesis, that there is a sensible constancy of the wavelength ratio between valley meanders and stream meanders throughout entire regions. He rejected previous explanations on the grounds that they could only explain a fraction of the total shrinkage and they were of restricted areal application. He concluded that the required explanation of underfitness had to be a widespread one and, in order to meet this requirement, it could be no other than climatic change. In addition, Dury (1977) found that the RIVER CAPTURE hypothesis could be rejected by means of hydrologic empirical power-functional equations which relate discharge to drainage area.

Underfit streams have now been identified widely in western Europe, where perhaps at least 50 per cent of the length of second and higher order streams is underfit, and as far east as the Ukraine. They occur in all the major climatic regions of the USA, including Alaska. They are present also in the coastal drainage of Australia's Northern Territory and on the eastern coastland.

Dury (1964) distinguished several types of underfit streams (see Figure 172). A *manifestly underfit* is a stream which meanders within a more amply meandering valley, the meanders being of the ingrown type. Manifest underfits can be identified immediately from air photographs or from reliable maps. A *variant* is when a manifest underfit is enclosed by sub-parallel scarps of rock, below which weaker formations have been eroded into a broad trough. In the *underfits of the Osage type*, except for being curved around the

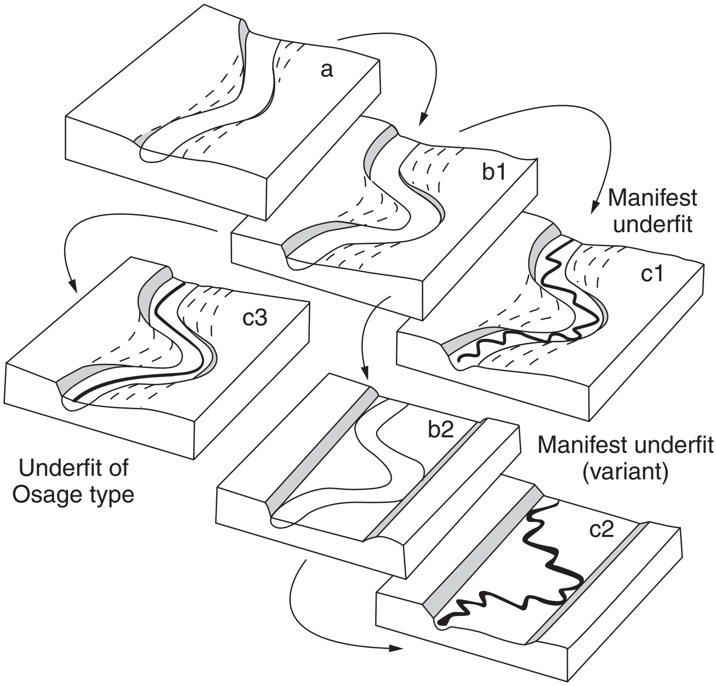

Figure 172 Partial model for the development of underfit streams

valley bends, the channel behaves as if it were straight, but the spacing of the pool and riffle sequences are more closely spaced than the valley meander wavelength would suggest. Osage-type underfits can only be suspected from their plan dimensions and have to be demonstrated from subsurface drilling of the pool and riffle sequence. Although they are named after the Osage River in Missouri, USA, they were first identified by Dury (1966) in the Colo River in New South Wales, Australia.

The development of underfit streams has two aspects: the date of origin of the rivers by which valley meanders have been cut, and the date of the shrinkage by which rivers were reduced to the underfit state. The latest date of origin for ingrown valley meanders is, in many areas, the time when incision of plateaux began. It is known that the incised meandering valleys of the Alpine Foreland and of the Hercynian massifs of Europe were initiated early in the Pleistocene. The high former discharges responsible for shaping valley meanders did not operate throughout the whole interval, but repeated episodes of shrinkage took place. Dates obtained by pollen analysis for the last major streams which are now underfit fall between about 12,000 and 10,000 BP. Evidence from river terraces also places the last main shrinkage well within last-deglacial time, when pluviosity markedly decreased. This appears to have been the time when general underfitness was finally confirmed, except in areas which were still covered by ice or by proglacial lakes.

There is a relationship between stream size and size of meanders. Meander wavelengths of valleys have been estimated to require a BANKFULL DISCHARGE of about 25 times greater than the present, and even 50 or 60 times greater when differences in channel slope, cross section and velocity are incorporated into the estimates. Stream shrinkage is then the result of a significant reduction in discharge at the recurrence interval corresponding to channel-forming flow or to the most probable annual flood. It was suggested by Dury that the channel-forming discharges required to explain the former channel patterns of streams which are now underfit could have been provided by increases in mean annual precipitation of 50 to 100 per cent. Magnitude-area-intensity analysis of precipitation shows that the means of reaching the required increase in precipitation is an increase in the frequency and power of storms.

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SEE ALSO: bankfull discharge; river capture; valley meander

MARIA SALA

UNDRAINED LOADING

If a soil is loaded very quickly it can result in there being no time for the drainage of pore water. There may be no consequential change in volume, but pore-water pressures change and result in differential shear and normal stress at every point in the loaded material. This rapidly reduces resistance and sometimes initiates shear movement, or accelerates movement downslope. The eventual dissipation of surplus pore pressure should reinstate steady-state equilibrium in the soil. The extent of pore-water pressure change will vary from soil to soil, as it is predominantly dependent on the soil's composition and its properties. Undrained loading is therefore important in slope stability analysis, particularly for soils with low permeability and receiving rapid loading, and has been incorporated into several modelling studies (e.g. Baker *et al*. 1993). However, such conditions are hard to accommodate within models, as influential factors such as changes in PORE-WATER PRESSURE are difficult to predict.

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STEVE WARD

UNEQUAL SLOPES, LAW OF

States that slopes will behave differently depending on their declivities (inclination). The law was proposed by G.K. Gilbert (1877: 140) within his paper on the geology of the Henry Mountains, USA. As rain falls on a slope, the amount of work it can do is proportional to the declivity of the slope. The steeper slope is always degraded faster, and will carry the divide towards the gentler slope. Thus, unless there are equal slope declivities, with homogenous material and identical rainfall, unequal slope activity will proceed. Eventually, a state of equilibrium will form (slope symmetry). Gilbert used this law to explain the form of BADLANDS, alongside his law of divides. This states that slopes on one side of a ridge are independent, while the law of equal declivities establishes a relationship between the slopes, the ridge crest, and the other side of the ridge. Thus, the slopes of the whole ridge behave independently over time and a landscape of unequal slopes develops.

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STEVE WARD

UNICLINAL SHIFTING

The gradual lateral shifting of a stream or river down-dip as a result of the slope of the underlying bedrock. When a river passes over a zone of inclined alternating hard and soft strata in a valley, it is usually easier for the channel to follow the strike of the less resistant strata, rather than to cut down into the harder rock. This may then induce a lateral shift in the channel. However, the mechanism by which uniclinal shifting occurs remains vague. Differential erosion, and rock permeability have been suggested as important factors, alongside the initial orientation of the channel. An example of uniclinal shifting is found in the Middle Thames Valley, England. Here an extensive flight of aggradation terraces exist on the northern side cascading southwards towards the London Basin syncline, yet there is almost nothing on the southern side that records the river's former route. This suggests the lateral 'uniclinal shifting' of the channel southwards, believed to have occurred in the Pleistocene (Bridgland 1985). Alternative terms to uniclinal shifting are down-dip shifting and 'down-dip migration'.

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STEVE WARD

UNIFORMITARIANISM

Uniformitarianism is a mode of thought which was conventionally defined as 'the present is the key to the past' (Geikie 1905). The names of James Hutton and Charles Lyell are permanently associated with having brought uniformitarian thinking into the main stream of geology, in opposition to the earlier catastrophist thinking, associated with names like Georges Cuvier and Abraham Werner. Prevailing schools of thought in the late eighteenth century had two main tenets: (a) the general belief that God has intervened in history, which therefore has included both natural and supernatural events and (b) the particular proposition that Earth history consists in the main of a sequence of major catastrophes, usually considered as of divine origin. Uniformitarianism as expressed by Hutton (1788) embodied two propositions that were contradictory to these catastrophist views: (a) Earth history can be explained in terms of natural forces still observable as acting today and (b) Earth history has not been a series of universal or quasi-universal catastrophes but has in the main been a long, gradual development. The most obvious example of the confrontation between catastrophists and uniformitarians at the time is provided by the contradictory views on the origin of valleys. Valleys gradually formed by rivers still eroding the valley bottoms were juxtaposed against valleys that had opened up as clefts through divinely controlled revolution. Although this dichotomy made much sense at the time (late eighteenth and early nineteenth century) (Gillispie 1960), there is considerable controversy over the use of the term uniformitarianism at present (Goodman 1967; Shea 1982; Schumm 1991). Hooykaas (1970) says that the use of uniformitarianism should be restricted to a view that states that geological forces of the past differ neither in kind nor in energy from those now in operation. The past should be reconstructed on the assumption that all geological causes of the past were of the same kind and intensity as those of the present.

Gould (1967) subdivided the confusing issues surrounding the definition of uniformitarianism into two components:

(a) a substantive uniformitarianism, which postulates uniformity of kinds and rates of processes (Hooykaas 1970) and
(b) a methodological uniformitarianism comprising a set of two procedural assumptions which are basic to historical enquiry in any empirical science: the principle of the uniformity of natural laws and the principle of simplicity.

Because (a) cannot possibly be true, and was not what Lyell had in mind (Kennedy 2000) when he popularized the term, and because (b) is standard procedure for historical science, the substantive content of the term is superfluous.

Shea (1982) identified twelve fallacies associated with the use of the term uniformitarianism: that uniformitarianism (a) is unique to geology; (b) was first conceived by James Hutton; (c) was named by Charles Lyell, who established its definitive modern meaning; (d) should be called actualism because it refers to the actual or real events and processes of Earth history; (e) holds that only currently acting processes operated during geologic time; (f) holds that the rates or intensities of processes are constant through time; (g) holds that only gradual, non catastrophic processes have occurred during Earth's history; (h) holds that conditions on Earth have changed little through geologic time; (i) holds that Earth is very old; (j) is a theory or hypothesis and can be tested; (k) applies only as far back in history as present conditions existed and only to Earth's surface or crust; (l) holds that the laws governing nature are constant through space and time. He recommends abandoning the term.

Kennedy's definition is, in the last analysis, the most satisfying

> uniformitarianism is a practical tenet held by all modern sciences concerning the way in which we should choose between competing explanations of phenomena. It rests on the principle that the choice should be the simplest explanation which is consistent both with the evidence and with the known or inferred operation of scientific laws. Uniformitarianism is therefore applicable to both historical inference and to prediction of the future outcome of the operation of natural processes (Goodman 1967).
>
> (Kennedy 2000: 502)

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SEE ALSO: actualism; catastrophism; neocatastrophism

OLAV SLAYMAKER

UNIVERSAL SOIL LOSS EQUATION

The Universal Soil Loss Equation is a method for estimating annual SOIL EROSION on the basis of soil loss from a field or hillslope. It was empirically derived from data collected over a twenty-year period from runoff plots at experimental stations established in the 1930s in the United States by the Soil Conservation Service under H.H. Bennett. The object was to measure soil erosion rates under natural rainfall on different soils, slope conditions, cropping and tillage practices, as a basis for SOIL CONSERVATION recommendations. Eventually data were available for twenty-three soils between the Rocky Mountains and the US east coast. Continuing attempts to develop a reliable equation to predict soil erosion culminated in the USLE in 1958 (Wischmeier *et al*. 1958).

The metric version of the equation is:

$$E = R \cdot K \cdot L \cdot S \cdot C \cdot P$$

where E is mean annual soil loss (t ha^{-1}), R is annual rainfall erosivity (10^7 J ha^{-1}), K is soil erodibility (relative to a control soil without vegetation cover), L is slope length (relative to a standard slope length of 22.6 m), S is slope gradient (relative to a standard 9 per cent slope), C is crop management (relative to a cultivated bare field), and P is a conservation practices factor (relative to a bare surface without conservation measures).

The most complex and critical factor is annual rainfall EROSIVITY, based on regression analysis of rainfall characteristics to determine those most strongly correlated with soil loss from the runoff plots. The most effective measure is a composite measure involving the total kinetic energy (E; J m^2) during a rainstorm and the maximum rainfall intensity recorded over a 30-minute period during the storm (I_{30}; mm h^{-1}). Annual rainfall erosivity is the sum of EI_{30} for all storms during a year, divided by 1,000. Calculations should be based on records spanning at least twenty-five years, but there are not many locations where such long-term records of rainfall intensity exist. Available data show the highest values from humid tropical areas like the Gold Coast of West Africa, where erosivity exceeds 1,700 (Roose 1977), and the lowest values in temperate and arid regions.

Other data were obtained by direct measurement at the original research stations and extrapolation procedures were proposed for areas without complete data. These are described in detail in *Agricultural Handbook 282* (Wischmeier and Smith 1965). This includes a nomograph for estimating soil erodibility (K) from soil texture, structure and organic content, graphical solutions for combined slope length and gradient (SL), and values for 128 crop combinations and cropping practices (C). The C factor can be subdivided for variations in cover protection through the cropping cycle. Finally, values of the conservation factor (P) are provided, based on tests at the experimental stations comparing techniques such as contour ploughing or terracing.

The USLE was designed as a conservation guide for farmers to estimate erosion hazard, to identify the most significant contributing factors and to predict the potential reduction in soil loss from introduction of conservation practices. It has been used widely in the United States, and has been effective in the area where original data are available. Subsequent modifications have been incorporated as understanding of soil erosion processes has increased, including, for example, a seasonally adjusted K factor, to reflect changes in soil structure caused by rainfall and weathering.

The modified equation was published in 1991 as the R (Revised) USLE.

The USLE has been useful for soil conservation in the central USA, and as an aid in instruction about the factors which control soil erosion. Unfortunately, its conceptual simplicity has encouraged use in areas for which it was not designed, such as steeply sloping forested lands, or in areas where inadequate rainfall records are available. It is purely empirical, with no proven physical foundation for extrapolation, and it can lead to extremely inaccurate predictions. Attempts have been made to accumulate appropriate data for wider use particularly in the tropics. Where measured data are available it can be used with some confidence, but elsewhere it is not reliable, particularly as a basis for expensive, socially disruptive or contentious measures. It has also been criticized on theoretical grounds as understanding of soil erosion processes has increased. The K factor, for example, is very simplistic, with no attention to important processes such as surface sealing, or to the effect of soil chemistry on erosion resistance. It certainly cannot be used with confidence to predict important soil erosion effects such as contaminant transport, or nutrient enrichment in lakes or streams. Much research has been directed to development of a sound physically based erosion equation, e.g. WEPP (Water Erosion Prediction Project) in the United States and EUROSEM (European Soil Erosion Model) in Europe. Although promising, these are conceptually complex, require data which are often unavailable, and are not yet sufficiently reliable for widespread use.

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RORKE BRYAN

UNLOADING

'The removal by denudation of overlying material' (Yatsu 1988: 140) was suggested by Gilbert (1904) as a mechanism that produced exfoliation domes in the granites of the Sierra Nevadas, western USA. According to this theory, granites are buried under a deep cover of older rock and so are subjected to compressive stress. That compressive strength was balanced by internal expansive stress competent to cause actual expansion if the external pressure was removed by denudation. This in turn may produce sheeting (EXFOLIATION) that is broadly conformable to topography. However, considerable controversy surrounds the issue of sheeting – does the sheeting produce the topography or is it *vice versa* as the unloading model suggests.

In addition to unloading being a potential cause of joint development, the term unloading is also used to describe the process of weight removal from the crust. *Glacial unloading*, resulting from the wastage of ice caps leads to postglacial faulting and seismicity, and to isostatic compensation. Equally, *erosional* or *denudational unloading* is an important factor in determing uplift and erosion in situations like passive margins (Clift and Lorenzo 1999). *Mechanical unloading*, associated with lithospheric extension, can contribute to flexural uplift of rift flanks (Weissel and Karner 1989).

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A.S. GOUDIE

(URANIUM-THORIUM)/HELIUM ANALYSIS

(Uranium-thorium)/helium analysis in apatite is currently the lowest temperature thermochronometer for providing detailed information on the thermal history of the crust for low (shallow) crustal temperatures. The technique is based on the alpha decay of U and Th in apatite to yield the daughter ^{4}He. (U-Th)/He analysis was the first radiometric dating system developed, by Ernest Rutherford at the beginning of the twentieth century. Independent geological evidence indicated that the ages it returned in its early applications were generally too young, and so the technique was abandoned as a dating tool. It was realized in the 1980s that (U-Th)/He analysis generally returned ages that are too young because the daughter ^{4}He diffuses out of the apatite above *c.*80 °C, which is well below the 'formation' temperatures that analysts were attempting to date in the technique's early applications. This behaviour means, however, that the (U-Th)/He system can be used to date rock cooling below *c.*80 °C.

The analytical procedures for (U-Th)/He in apatite are relatively straightforward, involving heating the apatite grains to drive off the daughter ^{4}He for measurement, followed by dissolution of the grains to measure the amounts of the parent U and Th using an Inducting Coupled Plasma Mass Spectrometer (ICPMS). The standard age equation for radioactive decay of parent element(s) to daughter element is then applied. Sample preparation is also relatively straightforward via standard mineral separation procedures, except for the final stage involving microscope handpicking of apatite grains for the analyses. This careful selection of grains is necessary to avoid the analysis of apatites with inclusions of U-bearing minerals, such as zircon. These U-bearing inclusions generate ^{4}He in the apatite but the inclusions' resistance to dissolution may mean that their corresponding U and Th contents remain unmeasured, thereby confounding the age calculation. A correction to the calculated age must be applied to account for the loss of ^{4}He from the outer edges of the grain by recoil (i.e. complete expulsion of ^{4}He from the grain during the alpha decay). This so-called 'recoil correction' requires that grains being analysed be of standard shape and of known size, which is a further reason for careful handpicking and characterization of the crystals to be analysed.

In the same way as there is a temperature range in which fission tracks are only partially retained (the partial annealing zone in FISSION TRACK ANALYSIS), the temperature range between *c.*80 °C and 40 °C defines the ^{4}He partial retention zone (PRZ). All of the daughter ^{4}He diffuses out of the apatite grain above *c.*75 °C, but at temperatures cooler than this the He is increasingly retained. Partial retention is grain-size dependent, providing the potential in (U-Th)/He analysis for an analytical tool comparable to the track-length distribution in fission track analysis. The potential of this grain-size effect is still to be fully explored.

Geomorphological applications of (U-Th)/He analysis largely focus on the rates of denudation required to bring apatite-bearing rocks to the Earth's surface from the crustal depths corresponding to a temperature of *c.*80 °C, at which the ^{4}He daughter product of alpha decay of U and Th begins to be retained. This denudation may be regional continental denudation required to bring the apatite to the Earth's surface (with the only 'tectonic' component being passive denudational isostatic rebound) or the denudation may be driven by tectonic rock uplift. In the latter case, and on the assumption that cooling through the PRZ is coeval with the tectonic uplift, the (U-Th)/He age corresponds to the age of onset of tectonic uplift. The assumption that denudation is coeval with uplift is extremely important. The assumption is reasonable in settings in which agents of subaerial incision, such as fluvial and glacial processes, are efficiently connected to the 'external' reference plane for tectonic uplift (e.g. a local base level or global sea level), and uplift-driven disequilibrium in the drainage net, due to relative lowering of base level, is rapidly transmitted through the drainage net to trigger incision and denudation throughout the catchment. Not all high elevation uplifting areas are well connected to the external base level which is the reference plane for surface uplift (e.g. the Tibet plateau, the Andes Altiplano). In these cases, the low-temperature thermochronological ages of rocks now at the surface of these landscapes may bear little relationship to the onset of uplift.

A further geomorphological application of low-temperature thermochronology, especially (U-Th)/He thermochronology in apatite, relies on the fact that the thermal structure of the shallow crust (upper few kilometres) is deformed by the long-wavelength Earth surface topography. The shallow crustal isotherms mirror topography of length scales of tens of kilometres. Thus the thermal structure of the crust beneath long-lived major valleys, for example, mirrors these valleys. If (U-Th)/He ages along a transect at constant elevation across these valleys (at a mountain front, for example) mimic the topography then the topography must have existed when the apatites were passing through the PRZ. That is, if the (U-Th)/He ages are older on that part of the transect coinciding with the interfluves and younger where the transect coincides with the valleys, the (U-Th)/He ages effectively provide minimum ages for the long wavelength topography.

PAUL BISHOP

URBAN GEOMORPHOLOGY

Urban geomorphology examines the geomorphic constraints on urban development (Cooke 1984) and the suitability of different landforms for specific urban uses; the impacts of urban activities on Earth surface processes, especially during construction; the landforms created by urbanization, including land reclamation and waste disposal; and the geomorphic consequences of the extractive industries in and around urban areas (McCall *et al.* 1996).

Constraints on urban development

The original founders of towns and cities carefully chose their sites for defensive, strategic, resource exploitation, navigation or cultural reasons. Great attention was given to finding sites with adequate water supplies and protection from obvious environmental hazards. However, the growth of these settlements often led to the spread of urban development on to less suitable ground and overstretched the capacity of the local environment to support the community. Many environments have particular conditions that make conditions for modifying slopes or establishing foundations difficult (Table 46).

Application of geomorphological mapping to the classification of the suitability of land for different types of urban development is now part of the work of geological and soil surveys in many countries. Such mapping considers the steepness of slopes, their colluvial and weathered mantles, their drainage and depth to bedrock and provides guidance on the type of development suited to different parts of the slope.

Knowledge of landform evolution is particularly important, as modern earthmoving can reactivate features inherited from past conditions, such as fossil periglacial landslides in Europe and North America. Loading of peat with urban structures formed after the retreat of ice sheets can result in significant subsidence and building damage. Karstic features formed when sea levels were lower in the Quaternary, but now buried under alluvium, can pose severe problems for the foundations of high-rise buildings.

Many present-day conditions constrain urban development. Widespread soils rich in montmorillonitic clays are subject to 'shrink–swell' cracking clay phenomena which require special foundations if buildings are not to become unstable. Climate change is likely to shift the areas where these problems are severe, if summers become drier. Mobile dune systems and sources of wind-blown sand pose problems for the siting of many structures. ALLUVIAL FANs may normally be inactive with the local stream confined to a narrow channel, but they may be reactivated, flooded and covered with debris if an extreme flood descends from the adjacent mountains. Building in PERMAFROST areas has to isolate the heated structures from the frozen ground and be careful not to disturb the permafrost during the construction process.

Geomorphic impacts during urban construction

Urban construction involves removal of the natural vegetation cover and excavation of the topsoil and often much of the underlying weathered rock and bedrock layers. In new urban developments, small streams are often diverted into culverts or urban drains and minor depressions and valleys are filled in. Steep hillsides may be terraced into a series of home sites by cut-and-fill operations. Major rivers may be embanked and artificially straightened. In extreme cases, as in Palma de Mallorca, Spain and Winnipeg, Canada, large new flood channels may be built around the urban centre to divert flows away from the city. The new features replacing the original landforms are often designed to direct water away from the new developments more effectively, so producing off-site, downstream consequences.

The earthmoving operations during urban construction frequently lead to severe erosion problems and consequent channel modifications

Table 46 Geomorphological problems for urban development

Environment	Chief problems
A Climatic	
Periglacial	Permanently frozen ground and overlying active layer require special types of construction and foundations for buildings and infrastructure
Arid	Water supply problems; wind erosion; flash floods; possibility of salt weathering of building materials and foundations
Humid tropical	Rapid weathering and decomposition of building materials; deep, uneven weathering of most rocks in tectonically stable areas; frequent rain events causing rapid water erosion of exposed ground surfaces
B Topographic	
Mountainous	Risk of unstable slopes, rockfalls, debris flows and avalanches; potential for flash floods
Floodplains	Liable to periodic flooding; variable foundation conditions over former, buried river channels and alluvial deposits
Coastal plains	Storm surge and flooding risk likely to increase with rising sea levels; complex ground conditions reflecting former shorelines and old drainage channels; possible salt penetration in ground water affecting foundations
Coasts with weak rock cliffs	Liable to rapid coastal erosion, cliff undercutting and collapse; eroded debris often deposited in ports and harbours causing dredging expenditure
Islands	Particular storm-surge, rising sea level and salt water penetration risks on low-lying atolls and coastal plains
C Tectonic/lithological	
Active plate margins	Major risks associated with coastal urban developments, especially on Pacific rim, special foundation requirements on filled areas, lake sediments and other unconsolidated materials; major earthquake-triggered landslide hazards; volcanic debris and lahar risks requiring awareness of flow pathways on lower volcanic slopes likely to have urban settlements
Shrink-swell clays	Cracking clay problems likely to be accentuated by climate change
Karst	Buried karst a major problem for foundations of tall buildings and for sinkhole development; need for knowledge of buried karst plains and effects of lowered Quaternary sea levels

Sources: Based on data in Marker (1996); McCall *et al*. (1996); and Bennett and Doyle (1997)

(Table 47, Figure 173). Erosion control guidelines suggest that construction should be carried out in phases to avoid disturbing too much of the land at any one time. No unnecessary clearing should be undertaken. Immediately below any cleared area, detention ponds should be constructed to retain any sediment washed off the site and to hold back stormwater runoff so that peak discharges in streams below are not increased.

Increased sediment loads and peak storm discharges lead to channel modifications (Figure 173) with formerly meandering streams becoming braided, steeper and shallower. Sometimes these changes are controlled by modifying the channel, often with expensive structural works. However, even these are not always successful as siltation of the channel can follow, with large accumulations of weeds and silt building up if the stream receives discharges with high nutrient loadings. Further downstream, rivers may adjust in response to upstream channelization, eroding their banks, developing new gravel bars and threatening bridge abutments and riverine structures.

Landforms of extraction

Meeting the demand for construction materials changes the land surface, by creating pits and

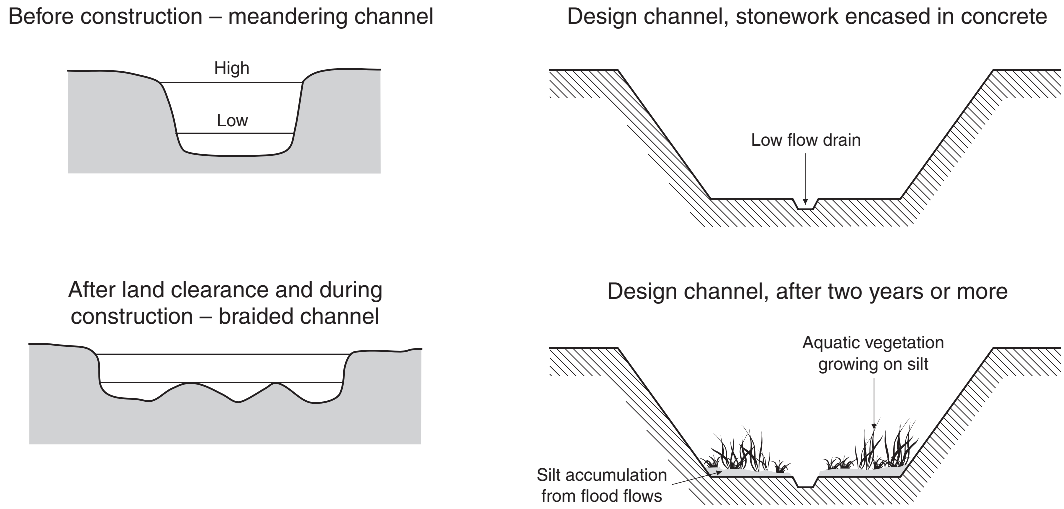

Figure 173 Channel changes to the Sungai Anak Ayer Batu at Jalan Damansara between 1960 and 1990 due to urban development

Table 47 Sequence of fluvial geomorphic response to land use change: Sungai Anak Ayer Batu, Kuala Lumpur

Land cover/land use	Channel condition
Forest	Narrow, meandering with low sediment load
Rubber plantation	Gullying during clear weeding; peak discharge increased; channel slightly widened; later stabilized; few cut-offs
Urban construction	High sediment yield; high peak discharge; metamorphosis to wider, steeper, shallower braided channel
Channelization and stable urban built-up area	Higher peak discharge; less sediment load; channel enlargement downstream; bank erosion, minor channel incision; loss of fine bed material by scour
Post channelization siltation	Where large quantities or organic debris enter concrete channels and are deposited, vegetation can become established and build up deposits that reduce channel capacity

quarries. The largest excavations take up many square kilometres of the land surface, often creating areas of brick pits that sometimes are used for waste disposal, or gravel pits that become peri-urban wetlands, often serving combined recreational and flood control purposes. Not all the filling of former opencut mines is without incident. In the past, escaping methane gas from landfills in old opencast coal pits has caused problems for the houses built upon them. In karstic terrain, the SUBSIDENCE of filled material in chalk pits or in old tin mines overlying cavernous limestone has led to severe damage to houses and urban infrastructure.

Removal of mineral resources and water from beneath the ground leads to subsidence creating new surface topographies and, often, new water bodies. Groundwater pumping has put the historic world heritage buildings of Venice at risk. Built at sea level on the lagoon, Venice has subsided some 22 cm since 1900. Most of that surface lowering occurred between 1950 and 1970. High water (*aqua alta*) has occurred more frequently since 1970. Whilst the people-induced subsidence is part of the problem, higher extreme sea levels related to global climate change may possibly be another factor. In the Los Angeles area, extraction of oil beneath Long Beach created severe subsidence that had to be halted by the injection of water into abandoned wells.

Landforms of deposition

Much modern urban development involves land reclamation and major landform modification (Gupta and Ahmad 2000). In extreme cases, huge quantities of material are moved, for example in the development of major airport sites such as Kansai, Singapore and Hong Kong. At Kansai, the fill material has caused some subsidence of the original seabed, with allowance having to be made for this in the operation and maintenance of the airport. The problems of subsidence of the second-stage runway are expected to be more severe than in the first runway, with a prediction that after fifty years subsidence will have been 18 m compared to 11 m for the first stage. Detailed analyses have been made of the way landing aircraft cause small temporary depressions in the runway that in turn affect the drag on aircraft moving along the runway.

As disposal of solid waste moves from landfill to land raise, new hills appear on the edges of floodplains, above former gravel pits and quarries and on offshore islands. In some urban areas, waste dumps are prominent features of the landscape. Although the older dumps are the result of coal, slate and china clay production, modern land raise features dominate many low relief areas. Whilst much of this waste is deposited in disused opencast mines and quarries, land raise mounds are probably the fastest growing artificial landforms in many countries today. The greatest geomorphological impact of landfill is in river valleys, sections of which are being filled, raising the height of the ground surface well above the former floodplain level. This effectively reduces the flood storage capacity of the floodplain, shifting the flood problems downstream.

Many of the old dumps are being closed or modified, from the huge dumps on the edges of cities like Istanbul and Manila to the managed disposal areas, such as Freshkills on Staten Island, New York, which has been taking nearly all the 17,000 tons of waste the city collects each day. As events at the Payatas tip in Manila have shown, some of these urban waste mounds are unstable, prone to massive slumps and landslides. The loss of life and property that ensues is a challenge to the management of the waste disposal and the application of geomorphology to the construction of land raise mounds.

Urban regeneration itself involves creating new landforms as the old buildings are demolished and construction and demolition waste is used for on-site fill or is taken short distances to sites that have to be raised to be above known flood levels. Many historic city centres have been so rebuilt that the average level of streets is now above that of the entrances to medieval buildings. These changes in landform may often be individually small, but collectively they are the result of two of the main human drivers of global environmental change: increasing urban development and the mining and quarrying which supplies minerals for industry and the construction materials required to build all the infrastructure, homes, offices and factories of cities. Urban geomorphology is thus a key element in supplying the guidance needed to achieve a better quality of urban life and working towards more sustainable use of resources.

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IAN DOUGLAS

URSTROMTÄLER

Broad (2–25 km wide), pronounced depressions which are aligned parallel with the margins of the Pleistocene ice sheets in Europe and North America; also called *ice marginal valleys, ice-marginal streamways*. They were first studied in North Germany (Wolsted 1950). These forms can be traced across the North European Plain from Russia to the North Sea. Marshy, flat-floored trenches usually have steep erosive external scarps, 20–40 m high and, often, systems of lateral terraces (Galon 1961). Their courses relate to particular stages of the Scandinavian ice sheet limits. Their origin is usually explained as the product of both erosion and sedimentation by the conjoined waters from proglacial *meltwater channels* (see MELTWATER AND MELTWATER CHANNEL) and rivers which drained the ice-free areas in the south. Almost certainly, large volumes of water must be involved in their production; at least some might be derived from catastrophic outbursts from ice-dammed lakes. Jahn (1975) considered that intense thermal erosion of river-banks under permafrost conditions was also a generic factor.

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JACEK JANIA

V

VALLEY

'A depression sloping in one direction over all its length' (Von Engeln 1942: 7), which tends to be longer than it is wide. Valleys have a range of sizes and a multiplicity of names – gully, draw, defile, ravine, gulch, hollow, run, arroyo, gorge, canyon, dell, glen, dale and vale (Huggett 2002: 193).

Valleys are normally regarded as the products of a range of fluvial processes such as corrasion, abrasion, pot-holing, cavitation, corrosion and weathering. They widen by lateral stream erosion and by weathering, mass movements and fluvial processes on their sides. They lengthen by such processes as HEADWARD EROSION or by building new land (e.g. RIVER DELTAs) at their bottom ends. In planform they develop networks (see HORTON'S LAWS) and they have a variety of DRAINAGE PATTERNs, including ALIGNED DRAINAGE. Some valleys develop in bedrock (see BEDROCK CHANNEL) while others develop in superficial materials, such as alluvium. In general big rivers have big valleys, but there are cases where small, misfit rivers occupy large valleys. This can be due to river capture, which can divert large amounts of water from that valley into another river system. Alternatively, it can be due to major climatic changes that have decreased the flows of water through the valley meander systems (Dury 1997) while some valleys may have no channels in them at all (see DRY VALLEY). Many valleys are accordant to geological structures, while others are discordant as a result of antecedence or superimposition.

There is a huge diversity to valley forms (see ARROYO; BEHEADED VALLEY; BLIND VALLEY; BOX VALLEY; BURIED VALLEY; CANYON; DAMBO; DELL; MEKGACHA; TUNNEL VALLEY; WADI). While most valleys are of subaerial type, there are also SUBMARINE VALLEYs. Some valleys have regular cross sections, while others, for structural or aspect-related microclimatic reasons, display asymmetry. Equally, normal fluvial valleys are often perceived as having a tendency towards V-shaped cross profiles (though this is far from universal), while glacially excavated valleys are often perceived as giving U-shaped cross profiles with truncated spurs.

The origin of valleys has proved troublesome. In the early nineteenth century they were sometimes regarded as the result of the Noachian deluge (see DILUVUALISM). There was also a great deal of debate as to what extent they were essentially tectonic features, related to fracturing of the Earth's crust. It was not easily recognized that they were the result of rain and rivers. These debates are well reviewed by Chorley *et al.* (1964). However, as Kennedy (1997: 67) has pointed out,

> *Any* process which creates topographic irregularities will cause the subsequent concentration of any available surface moisture and, potentially, a stream-and-valley. Moreover, since valleys are exceptionally durable features . . . we must face the fact that many networks will contain portions which owe both their ultimate origin and also their persistence to *different* processes.

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A.S. GOUDIE

VALLEY MEANDER

Meanders which are usually cut in bedrock and usually have a greater wavelength than that of the contemporary river pattern. These meanders shape valleys winding rather symmetrically between hills, and they are much wider than the meanders of the river flowing in the alluvial plain or alluvial meanders. The two types of meanders tend to be geometrically similar, the only real difference being that meanders in bedrock are commonly ingrown, whereas meanders on a floodplain are not. Entrenched meanders, which have cut vertically down without enlarging themselves in the lateral and axial directions, are uncommon. They represent one end of a series which extends through the intermediate range of normally ingrown meanders to meanders on a floodplain.

Davis (1906) described bedrock meanders in relation to lateral corrasion and placed their origin during the youth stage of the cycle of erosion. Dury (1954, 1977) thoroughly studied meandering valleys in order to ascertain whether they were cut by a stream larger than the present-day one. From many examples, he showed that valley meanders were produced during periods of higher runoff and higher peak discharges, that is, before stream shrinkage which led to contemporary UNDERFIT STREAMS, and that these larger discharges in past times were associated with Pleistocene climate.

The relation of meander length to valley width in valley meanders in rock shows more scatter compared to those in alluvium, but it is apparent that the length is directly proportional to the channel width in both cases. In the rock meanders wavelength is 15 to 20 times valley width. On the other hand, study of individual bends of valley meanders suggests that differences in geologic structure and lithology lead to differences in wavelength of meanders in rock.

Because of the difficulty of visualizing how a channel could maintain a regular sinuous pattern while cutting across hard-rock strata, it has often been assumed that the meandering pattern was initiated in an overlying sedimentary cover and superimposed on the tougher rock below as the river entrenched itself into the strata. Very often these presumed overlying strata have been eroded away, and hence the hypothesis is difficult to verify. In most cases there appears to be no need for such a two-cycle hypothesis. Other meanders in bedrock suggest that the river was antecedent to uplift. That is, the river appears to have maintained its course, trenching the structure as the latter formed. In the absence of stratigraphic evidence it is impossible to distinguish between an antecedent river and one which was superimposed from an overlying cover.

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SEE ALSO: underfit stream

MARIA SALA

VASQUE

Limestone and AEOLIANITE SHORE PLATFORMS in Mediterranean and tropical regions commonly consist of a series of low terraces formed by wide, flat-bottomed pools or vasques. The pools are separated from each other by narrow, winding ridges that can be: built by calcareous algae, Vermetids (see VERMETID REEF AND BOILER), or even Serpulids (see SERPULID REEF); the residual CORROSIONal pinnacles of lapiés; or a combination of the two. Plates-formes à vasques are covered at high tide and washed by breaking waves at low tide, with the return flow cascading into successively lower pools (Plate 143).

Plate 143 Rimmed pools, vasques, developed in eroded aeolianites at Treasure Beach near Durban, South Africa

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ALAN TRENHAILE

VENTIFACT

A term introduced by Evans (1911) to describe wind-faceted stones, the surfaces of which are flattened such that they intersect at sharp angles. They include the brazil-nut shaped 'Dreikanter' of German workers. For their formation three conditions are required: strong, generally unidirectional winds; the presence of loose materials (sand, dust, snow, etc.) that are available for transport in suspension or saltation; and the presence of pebbles, boulders and bedrock outcrops projecting into the wind stream. However, there has been considerable debate as to the relative importance of abrasion by dust and by sand (see Breed *et al*. 1997 for a review) and to the precise mechanisms that produce flattened surfaces on three or more sides of many ventifacts.

Ventifacts have been noted on a wide range of lithologies, including basalt, granite, dolerite, aplite, andesite, chert, marble, dolomite and limestone. They occur in a range of exposed environments, including deserts, periglacial and coastal settings. They also occur on Mars (Bridges *et al*. 1999). Some ventifacts are relicts of former tundra conditions subsequent to ice recession but before vegetation establishment in the Late Pleistocene. Such ventifacts have been used to infer palaeowind directions (Schlyter 1995), with strong easterly winds being present in Denmark and southern Sweden. In Ireland, some coastal ventifacts may have formed during the Little Ice Age of the Late Holocene, when there were increased offshore winds, waves, sediment fluxes and periods of sand dune construction (Knight and Burningham 2001).

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A.S. GOUDIE

VERMETID REEF AND BOILER

CORNICHES and other organic REEFS in the northwestern Mediterranean usually consist of calcareous algae and Serpulid (see SERPULID REEF) worms. Higher temperatures in the southern Mediterranean are favourable for large Vermetid populations, but whereas they only form veneers on TROTTOIRS in fairly easily eroded limestones and sandstones, there are purely constructional Vermetid corniches on fairly resistant substrates. Vermetids also contribute to the development of boilers or cup reefs in the Mediterranean and western Atlantic. Boilers, which are awash at low tide, are up to 12 m in height and a few tens of metres in diameter. They resemble MICROATOLLS with a central depression or micro-lagoon, up to a few metres in depth, surrounded by raised rims. Boilers in Bermuda consist entirely of coralline algae, Vermetid gastropods and the encrusting coral *Millepora*, but similar forms in the Mediterranean are merely veneers of Vermetids and algae over eroded AEOLIANITE blocks.

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ALAN TRENHAILE

VISOR, PLINTH AND GUTTER

CORROSIONal notches (see NOTCH, COASTAL) at the cliff foot sometimes have protruding visors above them and plinths below them. They have been described on AEOLIANITE SHORE PLATFORMs in southern Australia (Hills 1971), and from western Australia, Hawaii, Bermuda and northwestern India. The visor consists of a band of hardened, indurated rock, which may form when fresh rain water deposits calcium carbonate where it comes into contact with rock that is saturated with sea water. This might explain why the height of the visor declines as it is traced into sheltered areas, although it is questionable whether seawater saturation in the high and supratidal zones can be maintained during low tidal periods. The plinth is a slight prominence which is attached to the outer edge of the notch base. Hills proposed that the plinth develops at the height to which water is drawn by capillary action above the platform surface. The gutter, or moat, is a channel, occasionally found at the base of a ramp (see RAMP, COASTAL), that has been eroded by sand, pebbles and small boulders.

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ALAN TRENHAILE

VOLCANIC KARST

Karst-like features occur in volcanic terrains and they have been classified into four types (Reffay 2001). These are shown in Table 48.

The type *pseudokarst sensu lato* consists of structural landforms in lava flows that are unrelated to denudational processes. They include lava tubes, lava speleothems, and pseudodolines and shafts generated by collapse of lava flow roofs. The type *pseudokarst sensu stricto* is created by piping processes in loose volcanic material (e.g. pyroclastic deposits). The type *orthokarst* develops as a result of dissolution of carbonatites. The type *parakarst*, develops as a result of dissolution of minerals other than carbonates.

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A.S. GOUDIE

VOLCANO

A volcano can be defined as the site on the surface of a planet or moon through which gaseous, liquid and/or solid materials are expelled or erupted, usually through the action of internal thermal processes. Eruptions often discharge magma – molten igneous material composed of a silicate or other liquid, with variable quantities of crystalline phases and gas bubbles, though this can be transformed dramatically in violent, explosive eruptions into a stream of rock fragments and hot gases. An eruption can also result from steam explosions that blast out near-surface

Table 48 Volcanic karst classification

Pseudokarst, S.L.	Lava	Tunnels, speleothems, etc.	Structural. Related to lava emplacement
Pseudokarst; S.S.	Pyroclastics; Andosols	Pipes, holes, canyons, dry valleys	Suffosion
Orthokarst	Carbonatites	Closed depressions, megalapiés	Dissolution of carbonates
Parakarst	Basalt	Closed depressions, lapiés, speleothems, travertine and sinter	Dissolution of Ca, Mg, Na and silica. Precipitation of $CaCO_3$ and SiO_2

Source: Modified from Reffay (2001)

rocks without any accompanying fresh magma. The erupted products typically accumulate around the eruptive vent or vents, and can, if eruptions are sustained or repeated, construct mountains of very considerable volume. Olympus Mons, the highest volcano in the Solar System, rises some 24 km above the surrounding martian plains, and has a volume of around $3 \times 10^6\,km^3$ (Plate 144). Volcanism, past and present, is one of the fundamental geological processes of the Solar System.

On Earth, volcanoes are broadly distinguished as either active or extinct. The term ‘active’ is used in different senses. Often it is used to indicate a volcano actually in eruption. However, it is also applied to all volcanoes known to have erupted in the Holocene period (last 10,000 yr). This broader definition obviously includes many volcanoes that have not erupted for centuries or even millennia, and which may actually be extinct (incapable of future eruption) but it helpfully covers many more volcanoes, which may experience long repose periods (the intervals between eruptions), and which can be considered dormant and capable of further eruption. Around 1,500 volcanoes are known or suspected to have erupted during the Holocene, and are therefore classed as active (Siebert and Simkin 2002). Of these, some 550 erupted in the historic period. Every year, an average of fifty to seventy volcanoes erupt, though some of these are volcanoes with long-lived eruptions spanning years or decades. On an average day, at least twenty of Earth’s volcanoes will be erupting as you read this page. It is important to point out that all these figures are for volcanoes on land. Accurate figures for submarine eruptions are not available, although it is known that seafloor volcanism dwarfs the magma output of subaerial volcanism by a ratio of as much as ten-to-one.

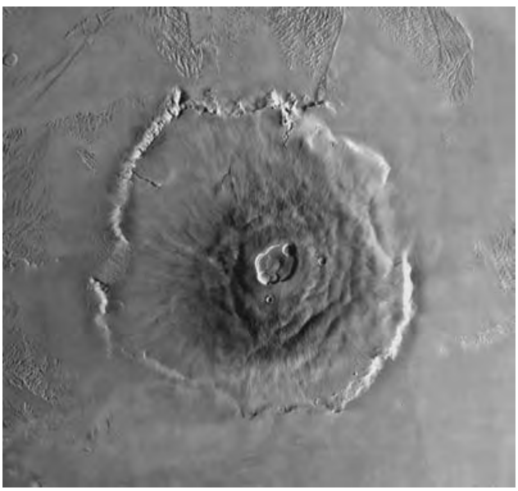

Plate 144 Olympus Mons – highest volcano in the Solar System. Note the overall shield shape, broad summit region crowned with nested calderas, and the prominent 550 km circumference, several km high basal scarp, whose origin has been the subject of much debate. Image processed by J. Swann, T. Becker and A. McEwen, and archived by NASA/NSSDC

Many dormant volcanoes discharge gases and liquids at the surface. In the case of hot springs, the flux of steam and gas is subordinate to that of liquid water. Geysers are spectacular examples of hot springs, Old Faithful in Yellowstone, Wyoming (USA) being perhaps the most famous. When gaseous emissions predominate, the term fumarole is usually applied. Emission temperatures of fumaroles therefore typically exceed the local boiling point of water. Long-lived fumarole fields are sometimes termed solfataras or soufrieres. Fumarole emissions are very often composed of both magmatic and hydrothermal gases, the latter evolving through complex chemical and physical interactions between magmatic fluids, meteoric water, seawater and rock. The emissions of some volcanoes discharge into crater lakes, which form by the condensation of the gases in the lake, as well as the capture of precipitation. The ‘black smokers’ associated with oceanic ridges are another important manifestation of subaqueous volatile discharge.

Some volcanoes have been in apparently continuous eruption for as long as records exist. For example, there is no evidence for any significant hiatus in the ongoing eruption of Stromboli (Italy) in over two millennia. The following volcanoes have been erupting more or less continuously for decades: Stromboli and Etna (Italy); Erta ‘Ale (Ethiopia); Manam, Langila and Bagana (Papua New Guinea); Yasur (Vanuatu); Semeru and Dukono (Indonesia); Sakura-jima (Japan); Santa Maria and Pacaya (Guatemala); Arenal (Costa Rica); Sangay (Ecuador); and Erebus (Antarctica). According to the records of the Smithsonian Institution’s Global Volcanism Program, the median duration of an eruption is around seven weeks. Most eruptions end within three months.

Materials erupted

Excluding some exotic but rare instances, most volcanoes erupt silicate magma of one sort or another. On eruption, these materials can be divided into lavas, which flow (with widely varying degrees of ease) across the surface in partially molten form, and pyroclastics (literally 'broken by fire'), which are expelled from volcanoes as solid fragments. Pyroclasts may be lofted in buoyant eruption columns to considerable heights in the Earth's atmosphere (exceeding 40 km altitude in exceptional cases), and can then be transported hundreds or thousands of kilometres by air currents. When they sediment to the surface they typically form ash fall deposits. These are often characterized by superposed beds of well-to-moderately sorted ash. If ash deposits harden, they are often called tuffs.

Although the term 'ash' is used widely in a loose sense, strictly it refers to pyroclasts with a diameter of 2 mm or less. The very finest ash is composed of shards, minute fragments of volcanic glass shattered apart by violent, explosive eruptions. Larger fragments (up to 64 mm across) are termed lapilli, and still larger ejecta are referred to as blocks (if solid when ejected) or bombs (partly molten when ejected). The very largest blocks and bombs tend to separate from an ascending eruption column and follow ballistic trajectories back to the surface, dropping relatively close to the vent. Fluid, coarse pyroclasts that accumulate around the vent are often termed spatter. Lapilli of mafic or intermediate composition (low or moderate silica content) that display a fragmented, bubbly texture are termed scoria. Highly bubbly pyroclasts, usually of more silica-rich (silicic) composition, are often called pumice. A common term pertaining to all pyroclastic material regardless of size is tephra.

Pyroclastic eruptions do not always produce stably convecting eruption columns in the atmosphere. They can also result in fountain collapse, in which all or part of the jet of pyroclasts leaving the vent region is unable to entrain and heat sufficient air to become buoyant. It then sinks back to the surface, where it can feed pyroclastic flows that move across the ground as a density current. Generally pyroclastic flows follow topography closely but they can gain sufficient momentum to overcome topographic barriers. The largest silicic eruptions on Earth are predominantly pyroclastic flow eruptions with volumes of a few thousand cubic kilometres (masses exceeding 10^{15} kg). Their deposits are often termed ignimbrites, and some anneal during compaction to form an excellent source of building stone. Pyroclastic flows consisting of ash-sized material are also known as ash flows, and their sediments as ash flow deposits. The finest particles can separate from a pyroclastic flow, and rise as a convecting plume to considerable heights in the atmosphere, ultimately settling out to form co-ignimbrite ash deposits.

Eruption styles

We have already referred to two very generalized eruption styles – lava (effusive) and pyroclastic (explosive) eruptions. Various schemes exist for classifying eruptions in finer detail. While they all retain some descriptive value, confusion can arise from inconsistent use of the terminology, and the fact that an individual eruption can display many different phenomena in rapid succession or even simultaneously (Plate 145). There are two basic approaches – one to describe an eruption on the basis of contemporaneous observations (i.e. the physical description of the eruptive phenomena), the other to characterize and interpret the deposits or impacts of an eruption. Clearly, the latter is the best or only available option for many eruptions in the past, and considerable efforts have gone into developing theoretical frameworks

Plate 145 Mount Etna (Italy) in eruption in August 2001. Note that several aligned vents are active simultaneously. The one disgorging dark ash clouds is at 2,550 m above sea level and produced a new cinder cone, now a tourist attraction. The vent below it is emitting a lava flow, picked out by the curtain of whitish gases rising above it

relating the depositional record to eruption physics and the atmospheric transport of ash clouds (Sparks *et al*. 1997).

The least ambiguous physical descriptors of eruptions are magnitude, intensity and duration. Intensity describes the mass eruption rate (e.g. in kg s^{-1}), and is a particularly useful parameter for explosive eruptions, as it is closely related to the height reached by the eruption column. Integrating intensity through the whole duration of an eruption yields the total erupted mass, or magnitude (in kg). This is also a useful first-order measure that enables intercomparison of sizes of different lava and/or tephra eruptions.

Eruptions do not always expel fresh magma. They can be driven by the sudden expansion of a liquid changing phase into a gas – for example, when ground water comes into contact with magma and flashes to steam – or by the instantaneous decompression of pockets of gas that have accumulated at some position within a volcano or its basement. These phreatic blasts can excavate considerable volumes of rock between the explosion source and the Earth's surface, leaving impressive holes in the ground. If new magma is also expelled in such explosions, they are termed phreatomagmatic. Such hydrovolcanic phenomena are quite common when a dormant volcano awakens – the volcano is effectively clearing its throat to make way for the passage of new juvenile magma.

A suite of more subjective terms to describe eruption style is also in common usage. These derive from particular historic eruptions (for example, 'vulcanian' refers to the 1888–1890 eruption of Vulcano; 'plinian' to Pliny the Elder's observations of the AD 79 eruption of Vesuvio, recorded by his nephew) or characteristic styles of individual volcanoes ('strombolian' refers to Stromboli volcano's propensity for modest pyrotechnic displays). Unfortunately, because one volcanologist's idea of a vulcanian eruption can be rather like another's image of a plinian, the terminology is not ideal but since it remains in widespread use it will be briefly outlined here.

Gas-rich eruptions of low viscosity magma can be sustained, generating the fire fountains characteristic of Hawaiian activity. Strombolian activity is typified by more discrete, fairly instantaneous explosions capable of propelling bombs a few hundred metres above the vent. These eruptions are formed as large bubbles of gases burst at the surface of a conduit filled with magma. Vulcanian eruptions are more violent. Here, build-up of gas pressure, often in a volcanic pipe blocked for decades or centuries since the last eruption, suddenly blows out a dense column of blocks and ashes, often composed of more old lava than new. These sometimes turn out to be throat-clearing eruptions that clean out the volcanic conduit ready for a more substantial plinian eruption. Plinian eruptions typically last for hours or days. The eruption plumes soar to heights of 20 to 40 km, and ash, gases and aerosols can circle the globe in a matter of weeks. Plinian eruptions produce well-sorted pumice and ash fall deposits. One important factor, as explosive eruptions crank up in scale, is the relationship between the duration of magma discharge and the rise time of the plume in the atmosphere. The physics of eruption columns diverge for discrete vs. sustained eruptions, with the latter capable of significantly higher ascent for a given intensity. The most intense known eruption, based on studies of its deposits, is the *c*. AD 181 outburst of Taupo in New Zealand. With an estimated intensity exceeding 10^9 kg s^{-1}, its ash cloud would have climbed to around 50 km above sea level (Carey and Sigurdsson 1989). Occasionally, as at Mount St Helens (USA) in 1980, volcanic explosions are directed more or less horizontally, and are termed lateral blasts.

Hydrovolcanic eruptions are sometimes referred to as surtseyan events. Lava dome eruptions (see LAVA LANDFORM), which often show sudden switches in behaviour between slow effusion of lava that accumulates in the dome, and explosions and dome collapses that feed pyroclastic flows, are termed peléean. Fissure eruptions have yet to earn a special name but they are quite common in some volcanic regions, especially Iceland, where magmas can rise up to the surface in vertical sheets (dykes) of considerable length. The discharge of magma quickly focuses on a number of discrete but aligned points, and the total length of the system can be up to 10 km or more.

Types of volcanoes

The landforms constructed by volcanism are very diverse, reflecting the supply rates of melt from the mantle, the storage, evolution and transport of magma in the crust, and the tectonic environment, as well as external factors such as presence of liquid water. Perhaps the simplest volcanic construct is the cinder or scoria cone. These are often monogenetic volcanoes formed as the result of a single eruptive episode. They seldom exceed 100–200 m in height and are typically composed of mafic scoria. Many much larger shield volcanoes, characterized by low angle, convex-upwards slopes and a broad summit region (Plate 144), are dotted with adventive cinder cones, which develop where branches from the central magma conduit break the surface on the flanks of the volcano. The shield profile reflects the generally low viscosity of the erupted lavas, and their ability to flow for considerable distances before solidifying. Shield volcanoes are usually crowned by nested and intersecting CALDERAS and collapse pits formed by subsidence. Calderas can also develop during large explosive eruptions as the crust founders above the emptying magma chamber.

When there is abundant ground water, hydrovolcanic eruptions can blast out wide depressions enclosed by low, circular or elliptical rims of ejecta. These features are known as maars and they may show little or no juvenile material. Tuff rings are distinguished from maars by being built on to the substrate rather than excavated into it, and typically contain more juvenile tephra. They have gentler slopes (2–10°) compared with tuff cones, which have gradients of 20–30°.

Most volcanoes are polygenetic, the result of numerous eruptive episodes. Simple cones, also called strato-volcanoes, are usually composed of interlayered lavas and tephra produced in countless eruptions over the lifetime of a volcano, which can be anywhere from a few thousand to many hundreds of thousands of years. They typically range in height between 1,000 and 3,000 m, and are often crowned by comparatively small craters (a few hundred metres in diameter). Mount Mayon (Philippines) is a good example, and justly famed for its near-perfect radial symmetry and convex-upwards slopes. Volcanoes may be subjected to major gravitational collapses during their history (see next section) but regrow by further eruption. The remodelled edifices that result are termed composite cones.

Clusters of overlapping volcanoes are sometimes called volcanic complexes, though the term is used rather loosely. Another case of more distributed volcanism is the volcanic field, formed where a single magmatic system has fed many discrete, usually monogenetic eruptions. The Michoacán–Guanajuato volcanic field in Mexico, composed of some 1,400 individual cinder cones, tuff cones and maars peppering a 200 × 250 km area, is a fine example.

Submarine volcanism has only recently received much attention, thanks to advances in deep-sea exploration. The geomorphology of ocean ridge volcanoes displays some of the characteristics of their subaerial counterparts in extensional tectonic environments but eruption in water at high pressure suppresses explosive activity and causes lava surfaces to solidify rapidly. Similar considerations apply to subglacial eruptions, common in Iceland. Here the weight of ice and presence of meltwater result in formation of tuyas in the case of central vent eruptions, or móbergs when fissure eruptions take place. Catastrophic releases of meltwater, jökulhlaups, associated with subglacial eruptions, are responsible for some of the highest discharge rates ever measured, and are capable of transporting massive quantities of sediment. In Iceland the deposits form plains known as sandur.

Erosional features

Because of their physical prominence, volcanoes are prone to all the usual agents of erosion. There can even be feedbacks between constructive and destructive processes that strongly affect the history of a volcano and its geomorphology. Measures of the degree of erosion of a volcano by wind, rain or ice, can be usefully applied to assessment of the relative age of activity. Rivers can slice rapidly through pyroclastic deposits on a volcano's flanks, while fresh tephra can even more quickly infill them. Once erosion gets the upper hand, volcanoes can exhibit well-developed planezes, triangular facets on the flanks of the cone delineated by the intersection of gully heads.

The most devastating destructive events that affect volcanoes are large-scale gravitational collapses, or volcanic landslides. The largest flank failures, sometimes called sector collapses, are a common feature of many composite cones and oceanic volcanoes, and they have generated the largest debris avalanche deposits on Earth (Moore *et al*. 1989). These are often identifiable from their characteristic hummocky topography, even when traces of the avalanche scar have been obliterated.

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SEE ALSO: caldera; lava landform

CLIVE OPPENHEIMER

W

WADI

A fluvial valley in a dryland region. Some wadis are relict landscape features that developed in former pluvials either as a result of increased overland flow or because of groundwater sapping (see DRY VALLEY). They include the *Megakcha* of the Central Kalahari, an area where there is currently little or no surface runoff. In the Sahara some of the wadis are enormous palaeodrainage systems up to 1,400 km in length (Pachur and Peters 2001). Others are more stubby features that some authors attribute to groundwater sapping (e.g. Luo *et al*. 1997).

Other desert wadis are sporadically active systems. This is, for example, the case in the Negev, where runoff and sediment yields can be high. One reason for this is the nature of rainfall events in the region. Rainfall intensities can be high (Schick 1988). In the Nahel Yael catchment, over a seventeen-year period, intensities exceeding 14 $\mathrm{mm\,hr^{-1}}$ accounted for nearly one-half of the total rain (223 mm out of 449). Of this intense rain, 37 per cent fell in intensities exceeding 2 $\mathrm{mm\,min^{-1}}$. Extreme flooding in the wadis can follow major rainfall events, as was demonstrated by the storms that afflicted southern Israel and Jordan in 1966 (Schick 1971).

Not all desert rainfall occurs in intense storms. A major contributing factor in runoff generation is the nature of some of the desert surfaces. For example, with dry conditions and a limited vegetation cover, silty soils, associated with loess deposits, rapidly become sealed under the influence of raindrop impact, and so have diminished infiltration capacities. Even on moderate slopes, silty soils generate substantial runoff (Evenari *et al*. 1983). Another runoff generating surface type results from the presence of organic crusts. These contain Cyanobacteria which partially plug soil pore space, particularly when they swell after they are moistened by rain (Verrecchia *et al*. 1995). Yet another important type of surface for runoff generation is bare rock. Available data indicate that the threshold level of daily rainfall necessary to generate runoff in rocky areas is a mere 1–3 mm. This compares with 3–5 mm for stony colluvial soils and more than 10 mm for stoneless loess soils. Because arid areas have a greater exposure of bare rock than semi-arid their wadis may generate more runoff and sediment (Yair and Enzel 1987).

Studies of experimental catchments over several decades have indicated high rates of sediment yield. Of particular note is the magnitude of bedload transport that has been recorded in the Nahal Yatir and neighbouring areas. Reid *et al*. (1998) showed that although wadi channels are only hydrologically active for about 2 per cent of the time (*c*.7 days per year) and only have overbank flow for about 0.03 per cent of the time (3 hours per year), the bedload flux is remarkably high. Indeed, the Nahal Yatir is about 400 times more effective at transporting coarse material than its perennial counterparts in humid zones (Laronne and Reid 1993). The explanation for this (Reid and Laronne 1995) is that its bed is not armoured (see FLUVIAL ARMOUR) with coarse material. The unvegetated nature of the desert watershed provides ample supplies of sediment of all sizes and this, together with the rapid recession of the flash flood hydrographs and the extended periods of no flow, discourages the development of an armour layer. Therefore, the flux rates are not sediment-supply limited as they so often are in perennial stream channels.

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A.S. GOUDIE

WATER-LAYER WEATHERING

Water-layer weathering refers to the accelerated geochemical weathering that occurs on SHORE PLATFORMS immediately above the platform water level. The weathering is a result of a number of interrelated processes that require an unsaturated or alternately wet and dry environment in which to operate. These processes include the combined actions of wetting and drying including thermal expansion in some rocks, the chemical action of salt spray, salt crystallization and the removal of solutions through rock capillaries. Additional processes acting in the same environment can include wave and rock abrasion, biological processes including rock boring, and frost shattering. The main role of waves however is to remove the debris provided by the weathering processes. Over time positive relief features on the platform are weathered down to the water level. Below the water level (the level of saturation) the lack of drying and free oxygen precluded the above processes.

The type and extent of individual processes will be dependent on latitude/climate, exposure to sun or orientation, lithology and rock structure, tide range and level of wave energy on the platform. Water-layer weathering will progress most rapidly on platforms exposed to regular wetting by seawater or spray, in warm temperate to tropical environments which favour drying, and in weaker and more porous sedimentary rocks which increase the depth of activity and aid removal of debris.

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SEE ALSO: chemical weathering; shore platform; wetting and drying weathering; weathering

ANDREW D. SHORT

WATERFALL

A waterfall can be distinguished from cascades, cataracts or other sharp descents in a stream profile, by the free fall of water over a very steep rock face. The greatest single fall is 986 m at Angel Falls in Venezuela; Tugela Falls in South Africa drops 948 m; and a descent of 800 m occurs in several drops at Yosemite Falls. Although they have much smaller descents, Victoria Falls (123 m), Niagara Falls (62 m), Iguazu Falls (70 m) on the Parana River and Khone Falls on the Mekong River (22 m) are notable for their great discharges.

Many waterfalls have been initiated by tectonic uplift along continental margins, or by local tectonic disruption along a fault (see FAULT AND FAULT SCARP) or rift scarp, as at Thingvellir in Iceland. They also commonly occur where a stream profile has been steepened by glacial erosion of a valley side, as at Skykje Falls in the Hardanger Fjord of Norway, or where streams flow over sea cliffs. Some small waterfalls, like the upper Suha Falls (25 m) in Slovenia, are essentially constructional features resulting from the deposition of tufa on the bed of a stream. The great majority, however, are the result of the differential erosion of rocks of varying strength.

The example most commonly cited is Niagara Falls, probably because of the fine account written by G.K. Gilbert. There, a dolomite caprock is undercut by failure in the shale beneath it. Other notable examples of caprock waterfalls are Gullfoss and Dettifoss, in Iceland, though in these cases the fall is due to the variable resistance of basalt, and of interbasaltic sediment and breccia. But not all waterfalls are undercut. Many have vertical faces, and others are buttressed outward at the base. These types of falls are not just short-lived features, where a stream is temporarily held up by a resistant vertical barrier. Many of them have retreated considerable distances. Vertical or buttressed forms are widespread in south-east Australia, where they occur in crystalline and in sedimentary rocks; Dangar Falls and Fitzroy Falls are typical examples.

Since Gilbert's account of Niagara, undercutting of waterfalls has been widely attributed to the erosive power of water swirling back into the recessed fall face. However, in most cases this does not occur, because, especially during flood discharge, the water descends well out from the fall face. Most undercuts are quite dry, except when spray is carried in by up-valley winds. Spray may promote weathering of rock in an undercut, especially if it freezes and fractures the rock. Seepage is also important. When saturated, a rock may have only about half its normal strength. Moreover, the abrupt drop in elevation at a large waterfall may greatly increase the pore-water pressure caused by seepage within the rock at the base of the falls. For example, the large undercut at Belmore Falls, in south-east Australia, occurs not behind the falling water, but to one side of it, where the main line of seepage drops 60 m from the valley above.

Water flowing over the fall face can pluck joint-bounded blocks from the rock on the lip of the falls. During flood discharge, when velocities are great, erosion on the lip may also be caused by processes like CAVITATION. The main effect of the falling water is the excavation of a plunge pool at the base of the falls. The claim that falling water has little erosive power and that the erosion is primarily the result of the impact of transported boulders has been repeated many times, but is not true. The kinetic energy of debris-free water falling over a dam can result in serious erosion at the base. For example, water over the Kariba Dam on the Zambezi River scoured a hole 50 m deep in just four years. The maximum, or limiting, depth of erosion varies with the height of the fall and with the magnitude of the discharge, together with the strength of the rock into which the pool is cut. Debris carried over many falls, or scoured from the plunge pools, forms a protective armouring of bed at the downstream end of the pool.

The basic requirement for a waterfall to develop is that the rock incised by a stream has sufficient strength to stand in a steep face. As stress related to the weight of rock is greatest near the bottom of a steep face, a waterfall cut even in an essentially homogeneous rock, such as granite, is likely to fail at its base, and thereby retreat upstream. Even where weaker rocks occur on an undercut waterfall, they commonly are quite fresh, and rupture by brittle fracturing in response to high stress. For example, stress in the gorge walls below Niagara Falls is causing the shale beneath the dolomite to bulge outwards and to generate vertical fractures, which weaken the rock face.

The rates at which waterfalls retreat depend largely on the magnitude of their discharge and the resistance of the rock over which they tumble. It also depends on their planimetric shape, for horizontal stresses normally are such that a waterfall in a narrow slot is less stable than one flowing over a broad AMPHITHEATRE. Rates of retreat range from about 1 km per 1,000 years at Niagara and Victoria Falls, to about 0.1–2 m per 1,000 years at smaller waterfalls in south-east Australia.

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R.W. YOUNG

WATERSHED

The term watershed in British English is used for a drainage divide – watershed boundary (American English) or catchment divide (Australian English) – that defines a water parting or a line, ridge or summit of high ground separating the water flow in two different directions draining in two drainage basins or catchments. In American English and by definitions of several international organizations the term has been changed to signify the region drained by, or contributing water to, a stream, lake or other body of water and is often used synonymously with the term drainage basin or catchment (Bates and Jackson 1980). A watershed can direct water from a surrounding watershed to a point such as a watershed outlet or a sinkhole. The total area of the watershed in a horizontal projection above a discharge-measuring point (watershed outlet) is the watershed area, catchment area or basin area. The watershed can be subdivided into several subwatersheds or subcatchments that conduct water from the surrounding hillslopes from one or the other side into a channel or drain water from a tributary channel at a confluence into a larger channel of a higher order or magnitude (see STREAM ORDERING).

One source of local watershed leakage is seepage or underground flow of water from one drainage basin to an outlet in a neighbouring drainage basin during a flood. Sometimes stream flow in a karst area will flow into an underground channel or sinkhole or directly into the sea. This may also occur when a water body is involved, such as subterranean flow or seepage from a stream, swamp or a lake (drainage lake) into a neighbouring watershed, or above the surface, e.g. stream bifurcation as in deltas or braided river systems (see DRAINAGE PATTERN). In such cases it is very difficult to exactly locate the surface and subsurface watershed boundary. In mountainous regions one therefore defines a watershed that runs along the crest of the highest range as a normal watershed in contrast to an anomalous watershed that does not behave that way. The location of the watershed boundary and size is dynamic over time and space depending on the migration of the watershed boundary caused by, e.g. water body level (overtopping the divide), temperature (thermokarst), sedimentation (alluvial fan), isostatic movements or disruptions by an earthquake or MASS MOVEMENTs, that temporarily or permanently changes the flow direction (Fairbridge 1968).

A watershed can be considered as a dynamic environmental system unit that has the functional organization of interacting hydrologic and geomorphic processes, e.g. precipitation, evapotranspiration, infiltration, RUNOFF GENERATION, EROSION, transport and sedimentation. Watershed management is the administration and regulation of the aggregate resources of a drainage basin to control and regulate water quantity and quality driven by these processes, e.g. for the production of water and the control of erosion, stream flow and floods. Watershed managers use Geographical Information Systems (GIS) and a DIGITAL ELEVATION MODEL (DEM) to delineate watershed characteristics such as watershed boundary, drainage network and contributing subwatersheds for inventory and planning purposes. They also process geo-spatial information on weather and climate, topography, soils and geology, vegetation and land use as well as infrastructural data processed in a GIS to link them with environmental models to analyse and simulate the complex dynamic hydrologic and geomorphic processes for decision-making purposes (see MODELS).

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SEE ALSO: channel, alluvial; drainage basin; hillslope, form; hillslope, process

CHRIS S. RENSCHLER

WAVE

Water waves are periodic undulations of the air-water interface (Figure 174) defined by their height (H), wavelength (L), period of oscillation (T) and speed of propagation (C). They can be *progressive* (e.g. *wind waves, tsunami*) or *standing* (e.g. lake *seiche*, ocean *tide*) and either actively *forced* (e.g. *sea*, tide) or propagate freely (e.g. *swell, tsunami*). The tide is forced by the periodic tractive forces generated by the gravitational attraction of the moon–sun system on the Earth's hydrosphere. TSUNAMI (Japanese for harbour waves) or seismic sea waves result from disturbance of the ocean by an earthquake, an explosive volcanic eruption or a mass failure of ocean sediments. Wind waves are found wherever wind blows over a significant body of water (e.g. marine or lacustrine) and constitute the most important global source of water wave energy. Wind wave size (e.g. height, length, period) depends upon the wind speed, the length of open water over which the wind blows (fetch) and the duration of time that the wind blows. When the oscillatory fluid motions under waves interact with a solid boundary an oscillatory boundary layer is formed; thus, all waves have the potential to generate stresses at the boundary and carry out geomorphological work.

Wave spectrum

By convention, complex irregular wave fields are analysed using Fourier Analysis, which provides a 2-dimensional spectrum of heights and periods. If the direction of wave propagation is included a 3-dimensional *directional spectrum* is produced (Figure 175). Water waves are classified by their frequency (1/T) or period of oscillation, the generating force and the force restoring equilibrium of the water surface (see Kinsman 1965).

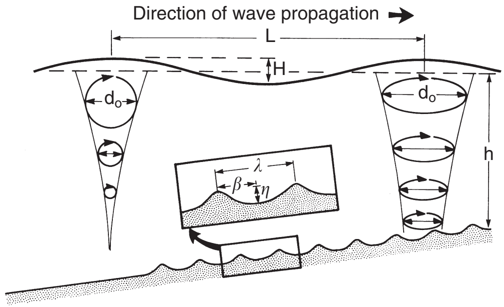

Figure 174 Schematic of wave form and water particle motion in deep and shallow water, where the wave form interacts with a rippled bed. Note: L = wave length; H = wave height; d_o = orbital diameter; h = water depth

TIDES

Tides are the characteristic semi-diurnal or diurnal rise (*flood*) and fall (*ebb*) of the Earth's mean sea level, caused by the periodic components of the gravitational tractive forces generated by the moon and sun; there are at least 360 tidal components leading to modulations of the water surface. Tidal waves are forced standing waves in the world's oceans that are resolved by Coriolis into a series of *rotary standing waves* (Kelvin waves) around the continental margins. These *amphidromic systems* have a central *amphidromic point* of no sea-level change and a *tidal range*, which increases outward to a maximum at the shoreline. Lines of equal tidal range are denoted by co-range lines, which encircle the amphidromic point. Tidal propagation is denoted by *co-tidal lines* (denoting points of equal *phase lag*), which radiate out from the amphidromic point. Such systems occur on all ocean coasts, even where tides are small. Tides are modulated at a range of timescales depending upon the changing relative positions of the Earth, moon and sun. Tides are largest (*spring tides*) when the sun, the moon and the Earth are in alignment (*conjunction* produces slightly larger tides than *opposition*); tides are smallest (*neap tides*) when the moon is in quadrature. Tidal waves are strongly influenced by interaction with the coastal boundary in terms of both their *amplitude* and *phase*, as tides attempt to penetrate into the embayments and estuaries around the coast where friction is important. In the Bay of Fundy, Canada, the tidal range may reach 18 m at springs, since the semi-diurnal forcing by the primary lunar component (M_2) is close to the *natural period of oscillation* of the basin and the shape of the basin causes a distinct topographic forcing. Tides are classified by range: (1) macrotidal >4 m; (2) meso-tidal 2–4 m; (3) micro-tidal <2 m (Davies 1973; Davis and Hayes 1984). *Surges* or meteorological tides are super-elevated tide levels (up to several metres) resulting from the combined effect on the water surface of barometric pressure differentials, wind stress on the water surface and wind wave set-up.

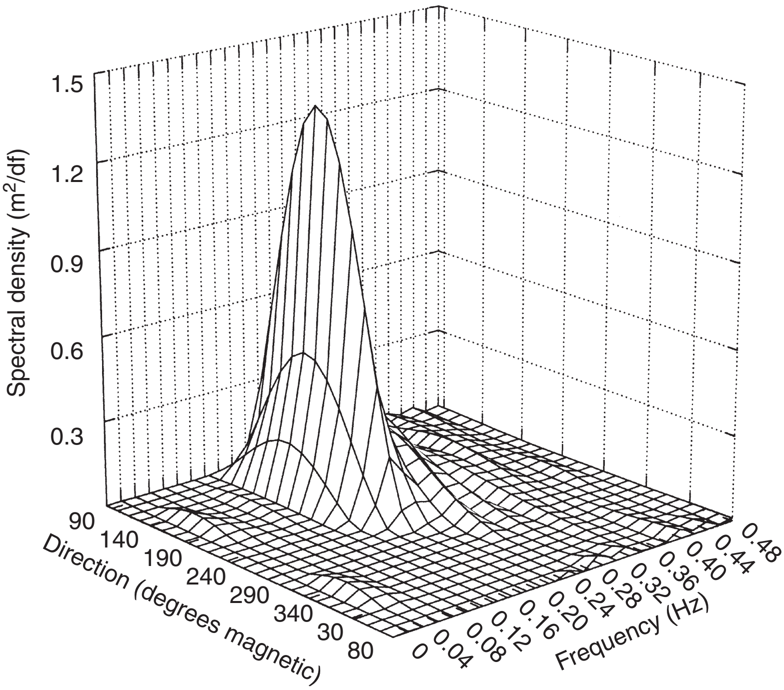

Figure 175 Directional wave spectrum recorded from southern Lake Huron, Canada in ~8 m water depth ~2 km offshore, recorded using a Falmouth Scientific 3DACM Wave Recorder. The record represents a 20-min sample of water surface elevation collected at 2 Hz. Note: a standard frequency spectrum can be obtained by simply projecting the data onto the left-hand frame. These waves have a peak period of ~5 s and a peak direction of approach between 160° and 170°

The rise and fall of mean sea level is complemented by horizontal translations of water known as *tidal currents* or *tidal streams*. Because of the long wave period, the currents are generally small, $<0.05\,m\,s^{-1}$; however, where topographic forcing occurs (e.g. estuaries, straits, inlets, etc.), tidally currents can reach speeds up to $6\,m\,s^{-1}$ (see CURRENTS).

TSUNAMI

In the deep ocean tsunami may be only a few centimetres or decimetres in height and may be difficult to observe, but may have periods ranging from 10–200 min, and hence extremely long wave lengths. Hence they can reach speeds of 700–900 km h^{-1}. In shallow water, as a result of wave *shoaling*, tsunami can reach catastrophic proportions, with run up heights of 10–20 m at the shoreline. The large run-up and large current speeds produce extensive flooding, damage to shorelines and man-made structures and even loss of human life.

GRAVITY AND INFRAGRAVITY WAVES

The most important waves outside the zone of shallow-water wave breaking (*surf zone*) are gravity waves (forced by wind and restored by gravity), with periods ranging from 1–25 s. A modulation of wave height is common, and gravity waves propagate as groups of large waves separated by several smaller waves (Figure 176). This modulation forces a secondary wave, the group-bound long wave, which propagates at the group speed and has a period equal to the group period. Waves with periods of ~25 s to several minutes or longer (frequencies of 0.004–0.04 Hz) are infragravity waves (first recognized and called surf beat by Munk 1949). They are most important to water circulation and sediment transport within the surf zone (Bowen and Huntley 1984). In intermediate water depths, infragravity energy results from wave–wave interactions and consists of a mixture of forced and free wave motions (Herbers *et al.* 1995). Several mechanisms associated with wave breaking in the surf zone have been proposed for their

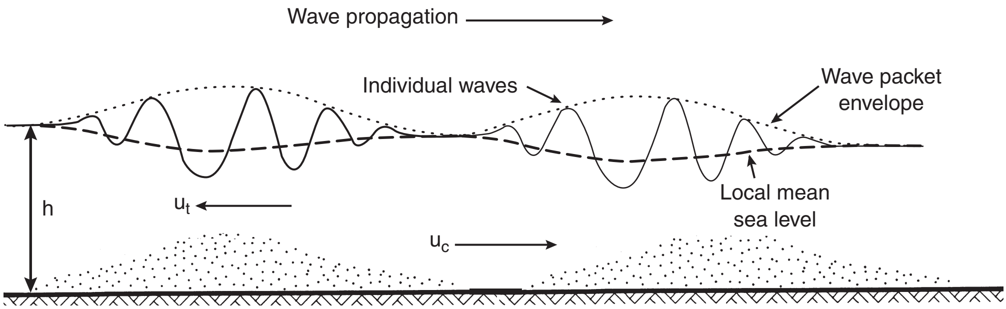

Figure 176 Wave groups and the forced group-bound long wave

generation including: (a) release of the group bound long wave (List 1992); (b) periodic shift in the position of wave breaking (Symonds *et al.* 1982); (c) persistence of groupiness into the surf zone causing a rise and fall of water at the shoreline (Watson and Peregrine 1992). Nearshore infragravity energy can take several forms including standing and progressive *edge waves* (trapped to the shoreline by refraction) and long *leaky waves* (reflected seaward without trapping; Huntley 1976). An important characteristic of infragravity waves in the surf zone is that they increase in amplitude and become more energetic as the offshore wave height increases. In contrast, gravity waves are saturated and thus decrease in importance proportionally during storms. *Shear waves* are a subset of the infragravity field (periods range from several tens of seconds to several tens of minutes) and they form within the surf zone through instabilities in the longshore current. They appear as large fluctuations in the mean current (Bowen and Holman 1989; Oltman-Shay *et al.* 1989) and can contribute significantly to both the cross-shore and alongshore transport of sediment (e.g. Aagaard and Greenwood 1995).

Gravity wave propagation and energy dissipation

Swift (1976) defined a number of *hydraulic provinces* stretching from the deep ocean to the coastline; in the innermost coastal zone (*shoreface*), both the hydrodynamics, sediment transport and morphodynamics are controlled primarily by gravity waves. Propagating gravity waves transfer energy and momentum from offshore to the shoreline. As they move into shallow water, frictional resistance results in a deformation of the orbital motions (Figure 174), a decrease in wave speed and wavelength, and an increase in wave height; this is called *shoaling*. The *specific energy density* (E, according to linear theory = $1/8\ \rho_f\ g\ H^2$, where ρ_f is fluid density, g is the gravitational constant) increases up to a point at which the wave becomes unstable and breaks. During propagation, wave energy may be redistributed laterally through convergence or divergence of the wave *orthogonals*; this is *refraction*. Further, the near-sinusoidal wave form characteristic of deep water becomes increasingly asymmetrical about the horizontal and vertical axes (*wave skewness* and *wave asymmetry*; Figure 177). These nonlinearities are critical to the net transport of water and sediment by waves, as they introduce non-symmetrical motion in the oscillatory velocities. In very shallow water, waves break when $H/h \approx 0.4–1.0$. The wave may be destroyed in a *plunging* breaker, producing a large *roller vortex* of the same order as the water depth, or the wave may continue to propagate as a *spilling* breaker, with collapse of the leading face of the wave. Surf *bores* are solitary waves, propagating across the surf zone controlled only by the water depth. The surf zone is a complex hydrodynamic environment, where interactions occur between incident waves, macro and micro turbulent vortices, secondary waves and quasi-steady currents of various origins (for a review see Kobayashi 1988). Ultimately wave energy is either reflected from the beach or dissipated in the surf zone and the reversing *swash* currents on the beach face. Sediment transport in the *uprush* and *backwash* is constrained by extremely small water depths, large Froude Numbers, large pressure gradients near *bore* fronts and the *infiltration* or *exfiltration* of water from the beach face *water table* (for recent review see Butt and Russell 2000).

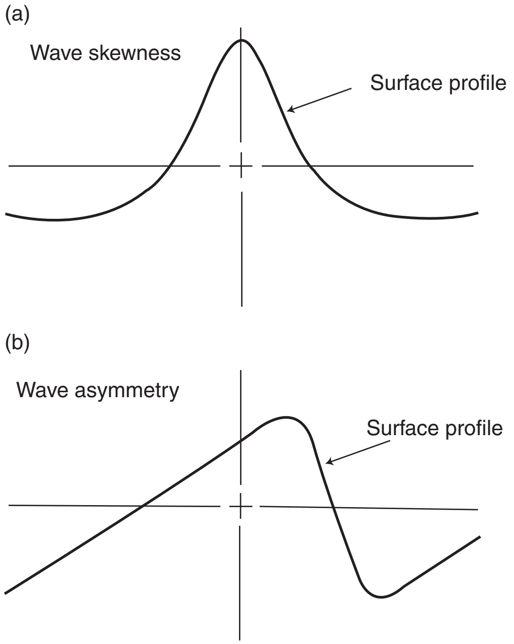

Figure 177 Asymmetrical wave forms: (a) asymmetry about the horizontal axis – wave skewness; (b) asymmetry about the vertical axis – wave asymmetry

Stresses generated by waves

Stresses generated by wave motion can be considered to take two forms: (a) *radiation stress* is the excess flux of momentum due to the presence of waves in the water and operates between the wave crest and trough. The concept was introduced by Longuet-Higgins and Stewart (1964). It explains the generation of the group-forced long wave in deep water and a number of hydrodynamic phenomena in the surf zone, such as wave *set-up* and *set-down* (i.e. the raising and lowering of the still water level at the shoreline), the *near-bed return flow* known as *undertow; longshore currents* (see CURRENTS); (b) the *boundary layer stresses* under waves are significantly larger than those of an equivalent quasi-steady current. Since the boundary layer develops and decays each half-wave cycle, it is much thinner and the resultant velocity gradients much larger. The time-varying bed shear stresses (τ_w), which have units of force per unit area (N m^{-2} in SI units) can also be written in units of velocity using the friction or shear velocity (u_{*w} which is not a real velocity, but is used for convenience and can be related to the fluid turbulence) and the fluid mass density (ρ_f):

$$\tau_w = \rho_f u_{*w}^2$$
$$u_{*w} = (\tau_w/\rho_f)^{1/2}$$

A dimensionless form of the bed shear stress is given by the *Shields Parameter* (θ_w):

$$\theta_w = \tau_w/\{g(\rho_s - \rho_f)D\}$$

where D is the particle size on the bed. The bed shear therefore depends not only on the flow velocity but also the *bed roughness* (grain size and bedforms). When quasi-steady currents are combined with waves (usually the case), these relationships become more complex (Soulsby 1997). However, if during the wave cycle these stresses exceed that for the *initiation of motion* of particles on the bed, entrained material will begin to move and oscillate in place. However, it is rare that a perfect oscillatory motion occurs in nature and rare to find a perfectly horizontal bed where gravity has no effect; thus a net displacement of particles occurs. Usually, there is also superimposed quasi-steady current, either induced by *wave streaming* (mass transport), the *asymmetry* of the waveform, or by some other process, such as *undertow* or *longshore currents*, which will enhance the net motion.

Sediment transport by waves

A simple concept of waves 'stirring' and currents 'transporting' sediment has been used to model transport under waves (Bagnold 1966; Bowen 1980; Bailard 1981); however this is not strictly correct. Waves and currents interact in a complex, non-linear manner and there is strong feedback from the bed materials (particle size, bedforms, bed slope, etc.). This is especially true for sediment transport, where rates have been related to some power of the horizontal velocity, and exponents varying from 3 to 7 have been used (Soulsby 1997). In the presence of ripples or megaripples, there is a complex, periodic shedding of turbulent sediment-laden vortices formed by flow separation at the sharp discontinuity of the bedform crest (Sleath 1984). In general these vortices are released upwards into the flow close to the time of flow reversal, when the oscillatory currents drop to zero. However, the timing of this release is critical as the vortices can be advected either down wave or upwave depending upon the release time and the wave period.

Geomorphological response to waves

The geomorphological response to waves depends on a range of variables including: (a) the timescale; (b) the spatial scale; (c) the geological and bathymetric setting; (d) the tidal range; etc. Forms can range from small sand ripples of a few centimetres in height and perhaps 10 cm in spacing, to the upwardly concave shape of the shoreface profile out to the depth at which waves begin to shoal (called *wave base*). Coastal morphologies are often classified by the nature of the substrate: (a) *rock, cliff* or *cohesive* coasts (see Tsunamura 1992; Trenhaile 1987); (b) *sedimentary* coasts (see Davis 1985), where the grain size is a critical determinant of form (gravel, sand or mud coasts). Sedimentary coasts account for approximately 20 per cent of the world's coastline, of which sandy coasts make up 10–15 per cent, and rocky coasts make up approximately 80 per cent. The vast majority of wave studies have been restricted to sedimentary, especially sandy, coasts, where measurable change is relatively rapid compared to rocky coasts.

ROCK, CLIFF AND COHESIVE COASTS

On bedrock and cohesive coasts exposed to large *wind waves* erosion of the base of cliffs (see CLIFF, COASTAL) or bluffs may occur, producing an indentation or *wave-cut notch*. The material above the notch may become unstable and *mass wasting* processes will induce failure. The resulting debris is then entrained by the waves and transported alongshore and offshore (depending on grain size) by wave-generated currents. The factors governing basal erosion rates are: (1) the *hydraulic* and *pneumatic* action of the waves, modulated by the water level, the upper shoreface slope and any beach materials present; (2) the *resistance* of the bedrock, which is controlled by its lithology, structure and geotechnical properties (Tsunamura 1992). The hydraulic action of waves consist of: (a) *compressive* forces, which vary with the waveform (standing wave, breaking wave, broken wave) at the cliff toe; the maximum compressive pressure is exerted by a wave that breaks exactly at the cliff face; (b) *pneumatic* forces may be induced by compression of air by waves in fractures and joints or other indentations and its explosive expansive release as the wave recedes; (c) *shearing* and *tension* forces develop across the rock surface as the wave uprush and backwash generate boundary forces capable of entraining loosened debris. The total mechanical process is known as *quarrying* or *plucking*. When significant sediment load is available, waves also cause erosion through *abrasion* and *impact* forces by the mobilized sedimentary particles. No field measurements have yet been made of the forces acting against a cliff and most of the information comes from theoretical calculations or laboratory simulations. Major questions concerning rock coasts revolve around the relative importance of waves and tides and subaerial weathering, as well as the extent to which forms may have been inherited from a previous time (Trenhaile 2002). Bryant and Young (1996) cite evidence for erosion of rock cliffs up to 15 m above present sea level by *tsunami* (see also Aalto *et al*. 1999).

Typical coastal forms are: (a) *steep cliffs*; (b) *wave-cut notches*; (c) *sea caves*; (d) *rock arches* and *stacks*; and (c) *shore platforms*, with or without a seaward *rampart*. All involve wave erosion to a greater or lesser degree. Mass wasting processes, such as *rock falls* and *planar* or *rotational slides, earth* and *debris flows*, and *piping* play a major role in maintaining cliffs along both rocky and cohesive coasts. Weathering processes are also critical in making material available for wave erosion and transport. The most characteristic form is a steep cliff and shore platform, with the latter being gently sloping or subhorizontal with a sharp break at the seaward end. Here the tide plays a major role in translating the hydrodynamic zones horizontally, so controlling the location, duration and type of wave action which occurs, but it plays no active role.

SEDIMENTARY COASTS

In cohesionless and very weakly cohesive materials, the shoreface is generally upwardly concave with slopes increasing toward the shoreline. Associated with this is a general shoreward increase in grain size. The increasing slope generates gravitational forces on the sediment to balance the increasing stresses from the landward propagating waves. An equilibrium profile shape was defined empirically by Dean (1991):

$$h = A\, x^{2/3}$$

where h = water depth, A = a constant and x is the distance offshore. Using the suspended sediment transport model of Bagnold, Bowen (1980) showed that if the equilibrium profile is defined as that shape where the net sediment transport over a wave period is everywhere zero, then the exponent 2/3 also applied. Van Rijn (1998) gives an extensive review of the literature on the *profile of equilibrium* concept.

Once waves break in the nearshore, the force balance on the bed is no longer simple. The occurrence of increased turbulence (macro and micro), secondary waves at a range of frequencies, secondary quasi-steady currents, and alongshore and cross-shore pressure gradient currents lead to complex sediment flux patterns, morphological forms and highly variable grain size. On many sandy coasts the upper shoreface profile is characterized by from 1 to 13 sand ridges called bars, which may be 2 or 3-dimensional, shore-parallel or shore-normal and, in the extreme case, form classic crescentic patterns (Greenwood 2003). The generation of barred profiles, their spatial and temporal dynamics have been the focus of extensive research. In 1979, Greenwood and Davidson-Arnott classified bars by their location and the primary forcing agents. Wright and Short (1984) proposed that nearshore bars (their number, form, etc.) are simply sequential changes as wave energy increases and decreases over time from a fully dissipative beach state to a fully reflective state. Using Dean's dimensionless fall velocity parameter (Ω), based on wave height (H_b) at breaking, wave period (T) and sediment size (settling velocity, w_s):

$$\Omega = H_b /(w_s T)$$

Wright and Short (1984) linked this to six distinct beach states; a similar *model* sequence was proposed by Lippmann and Holman (1990). Recently, attention has been directed at the feedback between the shoreface topography and the hydrodynamics, leading to the concept of self-organization of morphological forms (Plant *et al.* 2001; Wijnberg and Kroon 2002).

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BRIAN GREENWOOD

WEATHERING

Weathering refers to a group of processes collectively responsible for the breakdown of materials at or near the Earth's surface. Weathering occurs because the environmental conditions under which most rock materials formed differ substantially from those which prevail near the Earth's surface. Consequently, they undergo a variety of modifications which result in more stable products under the newly imposed conditions of temperature, pressure, moisture and gaseous environment.

From a geomorphological perspective, rock weathering is extremely important. First, weathering processes prepare Earth materials for subsequent transportation by agents of erosion. Second, weathering is an essential component of soil formation at the Earth's surface. Third, weathering processes are directly responsible for land form and landscape evolution. Karst landscapes and their distinctive land form assemblages, for example, are a direct consequence of weathering processes, as are the thick regolith-mantled landscapes of the tropics and subtropics.

Weathering processes

Traditionally, weathering processes are regarded as being physical, chemical or biological in nature. In reality, the three groups of processes act synergistically making it difficult to distinguish clearly between the effects of any single one (Pope *et al*. 1995). For instance, biological processes effect both physical and chemical change within the weathering system.

PHYSICAL WEATHERING

Physical weathering collectively involves the breakdown of rock material into smaller pieces without any change in the chemistry or mineralogy of the rock. Rock material is simply disaggregated as a result of the effects of the generation of forces within the rock mass. These forces are derived either from internal expansion of rocks and minerals, or from growth of materials in voids. The principal physical weathering processes associated with internal expansion include: INSOLATION WEATHERING (thermal expansion), unloading, HYDRATION, WETTING AND DRYING WEATHERING, organic expansion and SALT WEATHERING. Growth in voids is dominated by frost action and growth of salt crystals.

Insolation weathering refers to the breakdown of rock as a result of expansion and contraction caused by frequent temperature changes. As bedrock has low thermal conductivity, the surface of rock material expands more than the interior of the rock and, as a result, stresses are set up that eventually lead to the disintegration of the rock. Its effectiveness is enhanced when rocks consist of a mixture of light and dark minerals thus facilitating marked variation in thermal conductivity with individual minerals expanding and contracting at different rates. Research from engineering and ceramics has demonstrated that relatively small temperature variations over short time periods can most effectively disintegrate rock. Thermal stress has also been shown to be intimately associated with the breakdown of rock in areas affected by fire (Ollier and Ash 1983).

Unloading, or sheeting, refers to the formation of slabs of rock parallel to the ground surface, but separated from the underlying intact rock by joints. Unloading or sheet joints are generally attributed to the reduction in compressive stress in rock masses as a result of erosion of the upper part of the rock mass which then fails in the direction of stress removal. Failure is likely to begin with the extension of an initial crack and continue with the development of sheet joints parallel to the unloading surface. While this process appears to be most effective in brittle crystalline rocks it is also widely developed in massive sandstones.

While hydration begins as fundamentally a chemical process, with the absorption of water along fractures, especially cleavage planes in minerals, its effect is primarily a physical one. Expansion of minerals as a result of the incorporation of water into the crystal lattice can result in the production of tremendous force, especially in confined spaces. The effect of mineral expansion is to disrupt adjacent mineral grains causing loss of grain boundary cohesion and ultimately bedrock disaggregation to produce friable rubble referred to as grus. Similar processes concentrated around the edges of joint blocks where water is most abundant leads to the formation of concentric shells of weathered material referred to as spheroidal weathering or EXFOLIATION. The formation of these concentric shells of weathered material is also sometimes referred to as onion skin weathering or desquamation.

Repeated wetting and drying of rock may be a significant physical weathering process. Rock breakdown by wetting and drying involves the development of internal rock stresses as a result of the progressive formation of ordered water layers which exert forces against confining walls or void boundaries. As the positively charged ends of water molecules are attracted to the negatively charged surfaces of clay minerals and colloids they form a layer of oriented water particles. With each wetting event a new layer of ordered water is added to clay particles, which remain during the drying phase. Failure appears to be most pronounced during the drying phase when negative pore-water pressure is greatest and tensile failure occurs.

Frost weathering (see FROST AND FROST WEATHERING) has long been viewed as one of the most significant physical weathering processes in cold climates. Traditionally, frost weathering has been believed to occur as a result of forces generated in association with the volume increase accompanying the phase change from liquid water to ice. However, the theoretical assumptions underlying the generation of such forces are seldom met in nature. A more realistic model, with growing empirical support, is the segregated ice model of frost weathering in which expansion is the result of the migration of unfrozen water toward growing ice lenses and only secondarily the result of volumetric change accompanying phase change. Frost weathering is thereby the result of enlargement of microcracks and relatively large pores by ice growth accompanying ice segregation (Walder and Hallet 1986).

The growth of salt crystals in voids in rock exerts forces capable of disaggregating rock material when the tensile strength of the rock is exceeded. Salt crystal growth occurs when solutions containing salts are evaporated or in some cases cooled, when water is added to salts and hydration occurs and when salts are heated. All these processes result in substantial increases in volume of salt crystals and accompanying application of force against the walls of voids, leading ultimately to rock disruption and disintegration (Goudie 1997).

Biological organisms are also effective agents of mechanical weathering. Biophysical weathering processes include root wedging, and lichen, algal and bacterial activity. It has been widely suggested that the penetration of plant roots along fractures in rocks is capable of splitting rocks apart as the root grows. Physical weathering by lichens is accomplished in two principal ways. First, by the penetration of hyphae along microcracks and grain boundaries. The penetration and growth of hyphae have been shown to exert tensile stresses which exceed the tensile strength of most rocks. Second, by the expansion of lichen thalli and hyphae due to the uptake of moisture. Such moisture uptake substantially increases the size of the thallus and hyphae thus exerting considerable pressure on the rock surface. The efficiency of this process is seen in the frequent incorporation of rock fragments into the thallus (Barker *et al.* 1997). Algae also appear to be important contributors to the breakup of rock materials. It has been found that upon wetting, the polymer sheath surrounding algal cells increases by as much as twenty times, thus exerting expansion forces sufficient to pry already weakened rock flakes from a rock surface (Hall and Otte 1990).

CHEMICAL WEATHERING

CHEMICAL WEATHERING processes are those which involve the chemical and or mineralogical transformation of rocks and minerals at/or near the Earth's surface into products that are in equilibrium with Earth surface conditions. Several principal chemical weathering processes are recognized, including solution (dissolution), HYDROLYSIS, HYDRATION, carbonation, chelation and redox reactions. Solution refers to the dissolving of minerals in the presence of water and involves the removal of atoms from mineral surfaces thus reducing the stability of minerals and enhancing their vulnerability to subsequent chemical degradation (Blum and Stillings 1995). Hydrolysis refers to the reaction of hydrogen in solution with mineral surfaces and subsequent formation of secondary clay minerals as displaced cations react with hydroxy ions in adhered water on mineral surfaces. Hydration involves the addition of water to a mineral structure to form a new mineral. Redox reactions are reduction/oxidation reactions which involve reactions with oxygen in the atmosphere. OXIDATION involves a loss of electrons while REDUCTION involves a gain of electrons: oxidation of one mineral component is achieved through reduction of another. Iron is the most commonly affected chemical species. Carbonation involves the reaction of minerals with carbon dioxide in the presence of water and is especially important in the chemical weathering of limestones and other carbonate-rich rocks. In the soil environment, where the weathering system is dominated by clay mineral–soil solution interactions, ion exchange reactions occur. Ion exchange involves the transfer of ions between solution and mineral, and generally involves the replacement of ions in the mineral interlayer though replacement within the crystal lattice can also occur. The efficiency of this process is to a large degree controlled by the strength of electrical double layer, an exposed plane of negatively charged oxygen ions and the balancing swarm of exchangeable cations (Jenny 1980).

Plants represent significant contributors to the chemical weathering of rocks and minerals, primarily due to the fact that they produce abundant quantities of organic acids. These organic acids form ring structures around a metal core with multiple bonds between the organic acid and the metal. They are responsible for the chelation of cations such as Fe and Al. In addition, during their formation they commonly release H^+ which further reacts with mineral surfaces. Most of these acids are produced in the vicinity of root tips.

Lichens, algae and bacteria produce abundant organic acids which are responsible for considerable rock and mineral weathering. These organisms are responsible for the production of two principal groups of organic compounds. These include oxalic acid and various oxalates. The former compound possesses high solubility and contributes abundant protons for mineral dissolution as well as producing ring structures for chelation. Oxalates have been shown to enhance mineral grain dissolution through proton donation (Barker *et al*. 1997).

Weathering intensity

Weathering intensity refers to the degree of decomposition of a rock or mineral; it is a measure of the amount of alteration that has occurred. Numerous indices of weathering intensity have been developed and include both descriptive measures which are based on changes in the visual appearance of regolith as it becomes more altered, as well as quantitative measures of the amount of chemical and/or mineralogical change that has occurred to the primary mineral or unaltered rock. Descriptive or qualitative measures of weathering intensity have been widely used in the fields of engineering geology and regolith geomorphology. They typically are descriptive and subjective, based on changes in visual appearance of materials associated with progressive disaggregation or loss of cohesion. These classifications typically contain a limited number of weathering intensity categories, generally consisting of categories such as fresh rock at the unweathered end and soil at the most weathered end with slightly, moderately, highly and completely weathered categories in between. The boundary between the fresh, unaltered bedrock and altered material is referred to as the WEATHERING FRONT. This boundary may be quite abrupt, but more commonly it is highly irregular.

Semi-quantitative measures of weathering intensity involve the assessment of the relative hardness of fresh and various states of altered material. The most widely used of these methods is the SCHMIDT HAMMER which provides an index of hardness based on the amount of resistance to the compressive stress applied to the rock by the hammer.

A large number of quantitative measures of the intensity of weathering are available that are based on chemical and mineralogical characteristics of the weathering rocks. In virtually all cases, these indices compare abundances of non-resistant constituents to resistant constituents. They are expressed as dimensionless ratios which generally decrease as the intensity of weathering increases: as more non-resistant materials are removed. The more commonly used ratios include silica:iron, silica:aluminium, silica:sesquioxides, silica:resistates, bases:alumina, alkalis:resitates, and alkali earths: resistates. In addition several more comprehensive chemical weathering indicies have been developed and still receive limited use including the Parker Weathering Index, and the Reiche Weathering Potential Index.

Several mineral weathering indices exist such as quartz:feldspar ratios and multiple-mineral weathering indices including those for heavy minerals such as zircon + tourmaline:amphiboles + pyroxenes which increase as weathering increases. Several methods involving the characteristics of individual mineral grains have also been developed for evaluating the intensity of weathering including surface micro-textural features of heavy minerals and the degree of etching on ferromagnesian minerals.

The intensity of weathering is influenced by numerous interacting factors which affect both the extent to which primary minerals and fresh rock have been altered as well as the rate at which alteration takes place. Intrinsic factors include the mineralogy and chemistry of the parent material, grain size and shape, porosity, and fracture abundance and openness. Extrinsic factors generally include temperature, moisture abundance and water chemistry. Traditionally it is the role of temperature and moisture that have been emphasized in controlling both rates and intensity of weathering at the global scale as portrayed in the weathering models of Peltier (1950) who used temperature and precipitation as a basis for a global model of the variability of physical and chemical weathering processes. Similarly, Strakhov (1967) portrays the arid/semi-arid and tundra regions of the world as possessing a weathering mantle dominated by no more than disaggregated bedrock with essentially no chemical or mineralogical alteration, while the hot, wet tropics/subtropics carry weathering mantles dominated by the concentration of sesquioxides. In fact, the intensity of weathering is not controlled by tropospheric climate, rather it is controlled by a combination of boundary layer and reaction site temperature and moisture. More specifically, the intensity of chemical weathering is controlled by a complex set of synergistically operating factors that control both weathering intensity and rate including: the availability and proximity of both biotic and abiotic weathering agents; mineralogy, chemistry and petrology of the parent material; structure of the parent material at multiple scales; temperature at the reaction site; hydraulics of water movement; removal of weathered materials; addition/removal of organic and inorganic fines; microtopography of both landscape surface and mineral surface; exposed surface area and the presence of any accreted surface coatings (Pope *et al.* 1995).

Deep weathering

In many parts of the world, weathering profiles reach extraordinary thicknesses. These weathering profiles are commonly referred to as deep weathering profiles. Weathering to depths in excess of 100 m is not uncommon, and cases of weathering profiles extending to a kilometre or more, not unknown. While great depths of weathering have been traditionally recognized from the tropics this does not mean that deep weathering profiles outside the tropics necessarily imply their formation under such climatic conditions. In fact deep weathering profiles can develop in a wide range of climatic settings in which weathering has been prolonged. Deep weathering profiles around the world display a marked variation in the intensity of weathering. On the ancient landscapes of Australia and Africa, deep weathering profiles are characteristically intensely weathered, with the clay mineral kaolinite dominating. These intensely weathered deep profiles simply reflect the availability of abundant moisture and efficient removal of weathered products in solution. It is important to point out that deep weathering profiles displaying little chemical and mineralogical transformation also occur extensively in Australia, Europe and North America. These profiles reflect less efficient leaching and often also reflect weakening of rock strength early in the geologic evolution of the parent material by providing exceedingly deep fracture systems which can be exploited by circulating meteoric waters.

The deep, kaolinite-dominated, weathering profiles of Australia and Africa, commonly are capped by a strongly indurated crust enriched in a variety of chemical cementing agents. Collectively, these indurated crusts are referred to as DURICRUSTs and are most frequently, but not exclusively, cemented by iron, aluminium, silica, calcium carbonate or gypsum.

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JOHN C. DIXON

WEATHERING AND CLIMATE CHANGE

Weathering plays a fundamental role in the carbon cycle and hence influences the role of CO_2 as a greenhouse gas. While volcanoes add CO_2 to the atmosphere, the major long-term process of CO_2 removal is chemical weathering of continental rocks. The prime weathering mechanism involved in the process is HYDROLYSIS. Rates of chemical weathering vary through time in response to changes in temperature and precipitation amounts. They are also affected by vegetation cover, the nature of which is also linked to temperature and vegetation conditions. Thus climate affects the global rate of weathering, but weathering has the capacity of altering the climate by regulating the rate at which CO_2 is removed from the atmosphere. Weathering may act as a negative feedback that moderates long-term climatic change (Figure 178) (Ruddiman 2000).

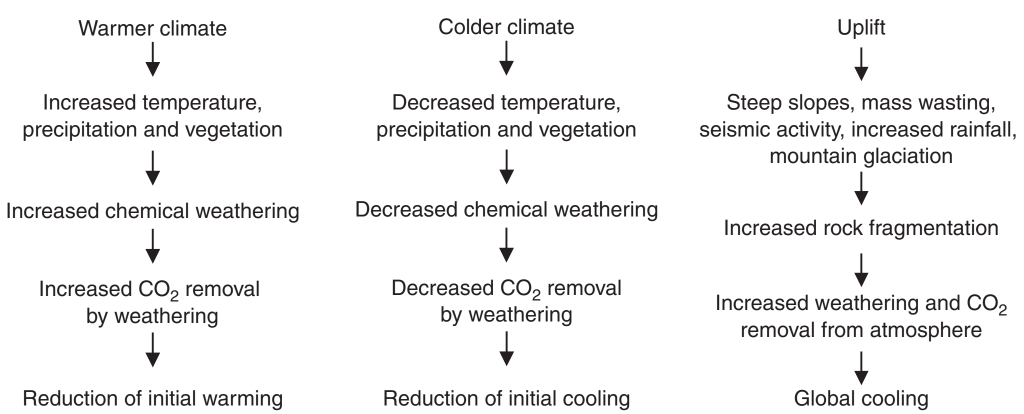

Figure 178 Negative feedbacks between weathering and climate change

However, it is also possible to see chemical weathering not only as a negative feedback that moderates climate changes, but also as an active driver of climatic change. The 'uplift weathering hypothesis' asserts that exposure of fragmented and unweathered rock is an important factor controlling the intensity of chemical weathering. It also asserts that rates of exposure of fresh rock are increased in areas of active mountain building because of seismic activity and mass wasting processes. Furthermore, uplifting areas generate more precipitation and glaciation. Mountain glaciers create pulverized bedrock.

Uplift has been active in the last few tens of millions of years (e.g. the Himalayas, Tibet, Andes, Rocky Mountains, European Alps). It is thus possible that accelerated chemical weathering has contributed to the climatic decline of the Late Cenozoic. It is also possible that differences in rates of chemical weathering between glacial and interglacial conditions in the Pleistocene occurred and contributed to the low CO_2 levels in the atmosphere during cold phases (Munhoven 2002). Karstic dissolution may also be an important component of the global carbon cycle (Gombert 2002).

The relationships between chemical weathering rates and climatic conditions are, however, complex (Kump *et al.* 2000) and the detection of variations in weathering intensity in the geologic record by the study of radiogenic isotope ratios ($^{87/86}$Sr, $^{143/144}$Nd, $^{187/186}$OS) is not without problems.

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A.S. GOUDIE

WEATHERING FRONT

The term pertains to subsurface weathering and is used to describe the boundary that separates solid, unweathered rock and rock that has already been weathered but remains still *in situ*. In reality, the transition between weathered and fresh rock compartments is rarely sharp. Usually there exists a transitional zone, in certain circumstances only a few centimetres thick, within which the change from unweathered mass to disintegrated or decomposed rock takes place. In some lithologies, especially in foliated metamorphic rocks and clastic sedimentary rocks, the transition is highly gradational and a well-defined weathering front may not exist. By contrast, weathering profiles in igneous rocks such as granite tend to have a very clear lower boundary.

The concept of weathering front has evolved from the notion of 'basal platform' introduced by Linton (1955) to explain the origin of TORs. Later it was shown that the boundary is hardly ever planar and the 'basal platform' is better replaced by 'basal surface of weathering' (Ruxton and Berry 1957). This implies a stable position of the rock/weathering mantle boundary and is not adequate to describe weathering profiles with abundant core stones. Therefore the present term 'weathering front' was recommended for use (Mabbutt 1961).

A weathering front is a dynamic feature, which migrates downwards and sidewards over time, as more and more rock disintegrates or decomposes. It should not be visualized as a single, continuous surface present at some depth, as each core stone is separated from the surrounding SAPROLITE by the weathering front. If the thickness of the weathering mantle is highly variable over short distances, the topography of the weathering front will accordingly be very rough.

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SEE ALSO: deep weathering; etching, etchplain and etchplanation; saprolite

PIOTR MIGOŃ

WEATHERING-LIMITED AND TRANSPORT-LIMITED

Weathering and transport limitations have been used as an important concept in geomorphology and are basic to an understanding of hillslope development. Broadly defined, a weathering-limited system is one in which the supply of material determines the flux of mass, while in a transport-limited system, sufficient material is available at any given time for mass movement to occur.

The two terms were originally introduced to geomorphology by G.K. Gilbert in his monograph on the geology of the Henry Mountains (Gilbert 1877). In this classic work the author refers to two landscape types: one in which the overall rate of DENUDATION remains limited by the rate of rock weathering and mass movement processes are very effective in removing deposited materials as they accumulate (weathering-limited); the second is characterized by weathering processes exceeding the rate at which material can be denuded and there is a net accumulation of rock (transport-limited).

The same concept was revisited by Alfred Jahn in 1954, but remained largely unnoticed until translated from Polish into English (Jahn 1968). Jahn uses the denudational balance of slopes to classify slope processes. Although Jahn does not explicitly use the terms weathering-limited and transport-limited in his 1968 publication, he clearly separates slopes into those where the intensity of weathering is lower than that of material transport and those where the opposite is true. He further explains that the denudational balance is in equilibrium when the intensity of weathering equals that of transport.

Eighteen years after Jahn's original publication, Kirkby (Carson and Kirkby 1972) first coined the terms weathering-limited and transport-limited, leaning on Gilbert's and Jahn's early explanations and defining weathering-limited slopes as those in which the transport processes are more rapid than weathering. In contrast, transport-limited slopes are characterized by weathering rates in excess of transport rates which implies the formation of a soil cover. In other words, the thicker the soil cover, the more transport-limited a process can be classified.

Gilbert's, Jahn's and Kirkby's definitions can be expanded to other fields in geomorphology. For example, a distinction was made between weathering-limited and transport-limited watersheds to explain DEBRIS FLOW activity (Bovis and Jakob 1999). In their paper, the authors define transport-limited WATERSHEDS as those in which there is a quasi-infinite amount of material available for transport by debris flows, while weathering-limited watersheds are those in which sediment supply rates to the channel system (by ravelling, rockfall, debris slides, etc.) are low and a channel requires recharge over a given time period before a rainstorm can trigger the next debris flow. This distinction is important in predicting debris flow frequencies as well as magnitudes. In the first watershed type (transport-limited) the exceedance of a climatic threshold will likely trigger a debris flow, while in the second watershed type (weathering-limited) a debris flow will only occur at the exceedance of a climatic threshold, if sufficient material has accumulated in the channel system. A similar separation has been made by Stiny (1910) in his monograph on debris flows written in German and later translated by Jakob and Skermer into English (1997). Stiny distinguished between 'Altschuttmuren' (old rubble debris flows) and 'Jungschuttmuren' (young rubble debris flows). The first are derived from large accumulations of debris (usually morainal material) with a quasi-infinite supply of material. The latter are derived from recently accumulated debris which may be depleted in the course of the debris flow.

Recently, the concept of weathering and transport limitation has been extended to encompass the development of fluvial BEDFORM. For example, during low flow stages of sand and GRAVEL-BED RIVERS, the channel bed may be armoured and sediment from upstream will pass over the armour layer in the form of sediment supply limited (weathering-limited) bedforms such as BARCHANS and sand ribbons (Kleinhans *et al.* 2002). Fully developed RIPPLES and dunes (see DUNE, FLUVIAL) exist when sufficient material exists for the formation of these features (transport-limited). The relevance of this distinction lies in the observation that bedform characteristics cannot be explained or predicted from local hydraulics and sediment characteristics alone. Another application is, for example, to assign controls on WEATHERING rates and solute (see SOLUTE LOAD AND RATING CURVE) fluxes on large rivers (Stallard 1985, 1995) using the same principles.

It is important to recognize that weathering-limited systems and transport-limited systems are not constant in space and time. For example, a large LANDSLIDE in a previously largely forested watershed may expose a large amount of debris which is now susceptible to transport by debris flows or other transport mechanisms. At this point the power of the stream (see STREAM POWER) network to transport the freshly accumulated and newly eroded material may have been exceeded and a previously weathering-limited watershed has now shifted to transport-limitation. Similarly, sudden climatic shifts may result in accelerated weathering rates which may not be matched by transport rates or vice versa. Finally, over time slope gradients may reach thresholds whereby the weathering-limitation may shift to transport-limitation.

The terms weathering and transport limited are somewhat misleading because they imply that the process of mechanical weathering is responsible for providing material to the system. However, other processes such as landsliding or resupply of material by fluvial processes may provide material which has not directly been supplied by weathering. Weathering is also hardly ever 'limited' compared to other geomorphic processes and is responsible for significant redistribution of material. Similarly, the term transport-limited may create confusion because it implies that it is the transport mechanism that limits the flux of mass. Rather than the transport mechanism, it is the availability of material that allows mass movement. It is therefore suggested to replace the terms weathering-limited and transport-limited by the less ambiguous terms 'supply-limited' and 'supply-unlimited'.

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MATTHIAS JAKOB

WEATHERING PIT

Small closed depressions on horizontal and gently inclined rock surfaces. They are also called 'gnammas', 'Opferkessel' or 'pias'. They have been described from a range of silicate rock types, most frequently granites and sandstones. They may be similar in their broad morphology to solutional pits developed in carbonate rocks, to which the term 'kamenitza' is often applied. Goudie and Migoń (1997: Table 1) provides a list of references on weathering pits from a diverse range of morphoclimatic regions that range from polar to desert and humid tropical. The largest examples may be between 10 and 20 m long (Twidale and Corbin 1963).

There is a great deal of uncertainty concerning the processes involved in the development of pits. Chemical processes of solution are usually invoked, but other processes may include hydration, the mechanical action of frost and salt and biochemical weathering. Positive feedback mechanisms related to the ever-growing amount of water available as a pit enlarges may account for a localized high intensity of weathering (Schipull 1978).

Many weathering pits are reported to be partially infilled with debris and organic matter, yet there are also many which are empty, with bare flaky bedrock exposed on the floor. It seems that

Plate 146 A large weathering pit developed in granite in the Erongo Mountains of central Namibia. This example has a completely bare bottom and the process by which debris has been removed from it is still the subject of debate

relatively little attention has been paid to the question of how the debris gets evacuated from the pit. Three possible ways have been suggested (Smith 1941), namely solutional transport and washing out during excessive rainfalls, and deflation. Flotation is another possible mechanism, but no direct observations have been made. Moreover, the occurrence of deep closed pits devoid of any sediments (cf. Watson and Pye 1985) remains puzzling and no satisfactory explanation of their emptiness has been offered.

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A.S. GOUDIE

WETTING AND DRYING WEATHERING

Fluctuations in rock moisture – wetting and drying – can cause rock weathering. In many cases these fluctuations cause weathering as a result of the rock expanding during take up of water and its inability to return to the original dimensions upon losing moisture. Moreover, high moisture contents can diminish rock strength, and through time wetting and drying cycles can reduce the bonding strength of component minerals. This in turn leads to a decrease in rock strength and possibly even failure (Pissart and Lautridou 1984). Cycles of wetting and drying can be frequent in some environments (e.g. shore platforms) and experimental simulations have shown the effectiveness of the process (Hall and Hall 1996). Micro-erosion metre studies have demonstrated the importance of surface swelling on shore platforms developed on limestones and mudstones, though it is less easy to discriminate between the importance of the growth of salt crystals and the expansion from wetting and drying (Stephenson and Kirk 2001). Indeed, wetting and drying is frequently a sine qua non for salt weathering. Wetting and drying also operates on badland surfaces developed on mudstones in semi-arid areas (Cantón *et al.* 2001), though once again it is the combined effect of moisture cycling and salt dissolution that is crucial. Nonetheless, clay-rich rocks are susceptible to SLAKING with or without the presence of salts (Gökçeuğlu *et al.* 2000), and the process can cause severe disintegration, as with the tombs in the Valley of the Kings, Egypt, which have been excavated in Esna Shales (Wüst and McLane 2000).

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A.S. GOUDIE

WILSON CYCLE

The Wilson cycle is the hypothesis that oceans are born as rifts, grow by SEAFLOOR SPREADING, and finally close again (Wilson 1966). Six stages are identified:

1 Uplift and extension forming rift valleys (see RIFT VALLEY AND RIFTING) (like modern rift valleys in Africa).
2 Early seafloor spreading (Red Sea stage).
3 Mature ocean (Atlantic stage) with a broad ocean bounded by continental shelves and a spreading centre at the mid-ocean ridge.
4 Shrinking of the ocean, bounded by ISLAND ARCs. The Pacific is in this stage, though the Pacific is also spreading.
5 Further shrinking, compression, metamorphism and uplift of accretionary wedges to form mountains (Mediterranean stage).
6 All oceanic crust is subducted, and the continents converge on a collision-zone suture (e.g. Indus–Tsangpo suture in the Himalayas). The united plate breaks along a new rift, restarting the cycle.

Some think the cycle should be repeated by rupture along roughly the same old line of weakness, others think a new cycle can start along a new fracture. Van der Pluijm and Marshak (1997) believe that rifting may occur in warm and weak orogens. Since orogenic belts are regarded as the weaker portions of the crust, they tend to be sites of repeated rifting and collision. Rocks in a single orogen may record the effects of several phases of extension and contraction. In the eastern United States, for example, the Earth history involves two phases of rifting (Late Precambrian and Middle Mesozoic) and two phases of collision (Late Precambrian and Palaeozoic), with several subphases (van der Pluijm and Marshak 1997).

The apparent polar wander (APW) path of two continents experiencing a Wilson cycle will show parallel tracks that diverge at breakup and converge after closure to a second parallel track. Such palaeomagnetic support for the Wilson cycle has been claimed for the rifting and closure of the Atlantic (Piper 1987). Gondwanaland and Laurasia converged during Silurian times to form Pangaea. From then to the Mesozoic they had a common APW path, showing they formed a single continent. They then split up into the present continents along new lines, continuing the Wilson cycle.

The opening and closing of ocean basins up to 1,500 km wide takes about 500 Ma, but some are smaller and more short-lived. Because of the timescale, evidence for Wilson cycles comes mainly from plate tectonic interpretation of palaeomagnetic, structural and geological data. In general it has little direct effect on geomorphology, but Russo and Silver (1996) relate the formation of the Andes to the Wilson cycle.

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CLIFF OLLIER

WIND EROSION OF SOIL

Wind erosion (see AEOLATION) has for long been recognized as an important factor in the development of landforms in dryland regions, contributing to the development of such features as close depressions (PANs), streamlined hillocks (YARDANGs), deflated surfaces (STONE PAVEMENTs) and miscellaneous micro-forms (VENTIFACTs). Wind erosion may also modify the form of dunes, as, for example, by creating blowouts (see DUNE, COASTAL). The general context of wind erosion is discussed in various other entries (see AEOLIAN GEOMORPHOLOGY; AEOLIAN PROCESSES; DESERT GEOMORPHOLOGY, etc.). This entry concentrates on the erosion of soil by wind – one component of DESERTIFICATION.

Wind erosion is a natural erosion process. However, its intensity and magnitude is often increased by miscellaneous human activities such as cultivation, overgrazing, vehicular traffic, etc.

(Leys 1999). It has negative impacts on agricultural production and produces environmental pollution, including DUST STORMS. Attempts to understand and model wind erosion have been made for some decades, most notably by W.S. Chepil and colleagues from the United States Department of Agriculture (see Chepil and Woodruff 1963). Using empirical field studies and wind tunnels (see WIND TUNNELS IN GEOMORPHOLOGY), they developed a much used Wind Erosion Equation.

$$E = f(C, I, L, K, V)$$

where E is the potential loss, C is a local climatic index, I is a soil erodibility index, L is a factor relating to field shape in the prevailing wind direction, K is a ridge roughness factor for ploughed ground and V is a vegetation cover index. The equation was developed initially in the Midwest of the USA and drew attention to the factors which could be manipulated by farmers (namely, I, L, K and V).

The climatic factor (C) was a simple combination of two key climatic variables: annual wind speed and a moisture index. Plainly, dry, windy areas are likely to be most susceptible to wind erosion. The soil erodibility factor (I) is more complex and needs to be seen in terms of both individual grain size characteristics and aggregate characteristics. Fine sands and silts are likely to be most susceptible, partly because of the relatively low velocities required for their entrainment, but also because the presence of clay tends to produce wind-stable clods. The presence of large clods reduces the risk of wind erosion. The fetch distance over which the wind acts (L) is related to field size and the presence or absence of shelter belts of differing heights, spacing and permeability. The ridge roughness factor (K) is based on the experimental observation that the rougher the surface, up to about 6 cm, the lower the windspeed at the surface. Thus furrows at right angles to the wind will tend to dampen down rates of erosion. The vegetation factor (V) is absolutely fundamental, for a dense vegetation cover, especially if like grass it has short stalks and narrow leaves, does more than anything else to reduce erosion rates.

More recently predictive models of wind erosion have been developed (Shao 2000). One such model is the Wind Erosion Assessment Model (WEAM), which aims to account for the combined effect of climate, soil, vegetation and land use. The fundamental physical viewpoint of this model (Shao *et al.* 1996) is that wind erosion is a result of two opposing forces: the capacity of the wind to start and maintain erosion, and the ability of the soil to resist it. The wind's capacity to start and maintain erosion is the friction velocity u^* (the wind shear or drag on the soil), while the opposing quantity offered by the soil is the threshold velocity u^*_t (the minimum friction velocity that is required for erosion to occur). The former is determined by wind flow conditions and the surface roughness, while the latter is determined by such surface factors as soil texture, aggregate structure and moisture content.

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A.S. GOUDIE

WIND TUNNELS IN GEOMORPHOLOGY

Wind tunnels provide a means by which processes in AEOLIAN GEOMORPHOLOGY can be monitored with a fully controlled and regulated wind flow. This is particularly advantageous in the aeolian environment where natural fluctuations in wind velocity and direction make it difficult to perform repeatable experiments in the field.

Wind tunnels vary greatly in design characteristics. They usually consist of a fan sucking (or blowing) air through a contraction which flattens out streamlines and provides a non-fluctuating wind to a working section where experiments can be carried out. The size of the working section may vary from $0.1\,m^2$ and 1 m in length to more than $4\,m^2$ and over 10 m in length. Working velocities in wind tunnels are commonly between 1 and $20\,m\,s^{-1}$.

Geomorphological experiments in wind tunnels are normally concerned either with aeolian sand or dust transport processes (Butterfield 1998; see AEOLIAN PROCESSES) where sediment is placed in the working section, or reduced scale studies in 'clean' wind tunnels of flow over fixed models of dunes, hills or valleys (Walker and Nickling 2002). In the latter case not only must the landform be reduced in scale but also the structural characteristics of the windflow (the atmospheric boundary layer). This is often achieved through a series of upstream grids and spires which, together with a roughened working section floor, can provide scaled values of shear velocity, aerodynamic roughness and turbulence length scales which mimic full-scale values. Portable wind tunnels are commonly used to test wind erosion (SEE WIND EROSION OF SOIL) characteristics of agricultural fields. Here the floor of the tunnel working section is absent so that when placed at a test site the tunnel flow acts directly on the natural soil.

Wind flow in wind tunnels can be determined with a pitot-tube which establishes velocity through the measurement of pressure. Where highly turbulent flows are encountered then electronic hot-wire anemometers are more commonly employed. Precise measurement of shear stresses on the surface of scaled models can now be accomplished using pulse-wire anemometers (Wiggs *et al.* 1996) or particle velocimetry whilst flow visualization over scale models is achievable by seeding the flow with smoke or covering the surface of the model with oil. In experiments involving sediment entrainment and transport sand traps can be erected at the downwind end of the working section to measure the rate of sand flux.

Whilst many advances have transpired as a result of the use of wind tunnels in research on aeolian processes and wind flow over aeolian bedforms, the biggest drawback of the approach is that there is commonly insufficient field-based empirical data with which results can be verified.

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GILES F.S. WIGGS

Y

YARDANG

A Turkmen word to describe wind abraded ridges of cohesive material (Hedin 1903). They have been described from a large number of arid regions (McCauley *et al*. 1977) where they have developed on a large range of materials. They have been likened in shape to the prows of upturned ships. They range in size from small centimetre-scale ridges (micro-yardangs) through to forms that are some metres in height and length (meso-yardangs), to features that may be tens of metres high and some kilometres long (mega-yardangs). The forms may go through a cycle of development and eventual obliteration (Halimov and Fezer 1989). Although they are dominantly aeolian erosion features there has been a considerable debate as to the relative importance of deflation, aeolian abrasion, fluvial incision and mass movements in moulding yardang morphology. That abrasion is important is indicated by polished, fluted and sand-blasted slopes, and the undercutting of the steep windward face and lateral slopes. It is probably the dominant process in hard bedrock yardangs whereas deflation may be important in the evolution of yardangs developed in soft sediments such as old lake beds. Fluvial erosion may provide an avenue along which wind erosion may occur but excessive fluvial erosion would tend to obliterate yardangs. Mass movements may also be significant when yardang slopes have been oversteepened by wind erosion.

One form of yardang, beloved of textbooks, but in reality not very common or significant, is the zeuge (plural zeugen). The term is derived from the German word for ‘witness’. Zeugen are tabular masses of resistant rock (2–50 m high) left standing on a pedestal of softer material as a result of differential erosion by sand-laden wind.

Plate 147 Characteristic aerodynamic yardangs developed by the wind erosion of Holocene pluvial lake beds in the Dhakla Oasis, Western Desert, Egypt

Plate 148 A series of wind eroded chalk outcrops in the White Desert, Farafra, Egypt. In addition to wind erosion, note the presence of case hardening and honeycomb weathering

At the mega-scale yardangs curve round in response to the trajectories of trade winds (as, for example, around Borkou in the Central Sahara; Mainguet 1972) and may show a remarkable parallelism over hundreds of kilometres (as with the *Kaluts* of the Lut Desert in Iran).

Study of the morphometry of yardangs has revealed relationships between different parameters. Ward and Greeley (1984) found a 1:4 width to length ratio; Halimov and Fezer (1989) found that the ratios of length, width and height were 10:2:1; while Goudie *et al*. (1999) found volume, length, width, height ratios of 18.7:9.9:2.7:1.

In soft materials yardangs can form quickly. In the Sahara, Mojave and Lop Nor deserts, Holocene lake and swamp deposits have been eroded at rates of *c*.2.5 to 5 m per 1,000 years.

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A.S. GOUDIE

YAZOO

A tributary stream/river that flows for a long distance parallel to the main channel before joining it. This usually occurs in mature river channels where natural levees have built up on the channel banks, above the mean height of the surrounding floodplain. The tributary channel is therefore unable to penetrate the trunk channel and lakes may form. However, more often the tributary channel is bounded by bluffs beyond the floodplain, thus forcing a Yazoo-type tributary downstream parallel to the main channel. The type example is the Yazoo River, which runs south/south-west parallel to the main Mississippi River from Memphis, Tennessee (USA) for almost 400 km. The Yazoo eventually infiltrates the Mississippi channel at Vicksburg, Mississippi. The elevated bed of the Mississippi River obstructs the Yazoo tributary and forces it to flow parallel along the floodplain (35–150 km wide) and with a Yazoo channel gradient of only 11.4 cm per km. Yazoo-type channels are also evident in deep sea environments (Hesse and Rakofsky 1992).

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STEVE WARD

Z

ZOOGEOMORPHOLOGY

Zoogeomorphology is the study of the geomorphic effects of animals (Butler 1995), and as such can be considered as a subset of BIOGEOMORPHOLOGY. Zoogeomorphology encompasses the geomorphic effects of both free-ranging, wild animal populations as well as the geomorphic effects of domesticated animals such as cattle. It considers the geomorphic role of animals regardless of their size, ranging from small invertebrates such as ants, worms and termites (see TERMITES AND TERMITARIA) to both terrestrial and aquatic vertebrates including fish, amphibians, reptiles, birds and mammals.

Animals geomorphologically alter the Earth's surface through both direct and indirect effects. Direct effects encompass surface erosion, transportation and deposition of rock, soil and/or unconsolidated sediments by animals. Animals accomplish these effects by digging for and/or catching of food, nest building, burrowing, mound-building, wallowing, eating soil or rocks, canal excavation and dam building (accomplished by beavers; *Castor canadensis* in North America, *C. fiber* in Eurasia). Indirect effects do not specifically remove sediment from a surface, but processes such as trampling associated with overgrazing may lead to a reduced infiltration capacity in the underlying soil, in turn resulting in greater surface wash, soil creep and rainsplash detachment. Trampling and associated overgrazing by domesticated animals also contribute to widespread bank erosion, riparian degradation and unstable channel morphologies (Trimble 1994; Magilligan and McDowell 1997; Naiman and Rogers 1997; Evans 1998).

The geomorphic impacts of animals may vary spatially and temporally through the course of a year, as wild animals migrate in search of changing food sources (Baer and Butler 2000). In the Rocky Mountains of western North America, for example, grizzly bears (*Ursus arctos horribilis*) create widespread diggings during the spring on lower floodplains in search of roots and tubers. As summer progresses, they migrate upslope into areas of snowmelt where they dig for burrowing mammals such as ground squirrels as well as for roots. Late summer sees the bears on high talus slopes, where they excavate large quantities of boulders, in search of insect larvae.

The seasonality of geomorphic impacts by animals is also a function of a region's climate. Excessive heat and dryness minimize the summertime geomorphic impacts of some desert dwellers. In colder climates, frozen ground limits the seasonal ability of animals to dig into the surface. During such periods, many burrowing animals may also hibernate, effectively cutting off geomorphic impacts until the subsequent spring (Butler 1995).

Beavers have some of the most profound geomorphic impacts of any animal in North America. Beavers are large (adult mass >15 kg), semi-aquatic rodents that instinctively attempt to build dams in response to the sound of running water (Butler 1995). Prior to European settlement of North America, as many as 400 million beavers may have occupied an area of approximately fifteen million square kilometres. By the year 1900, trapping of beavers in North America had nearly extinguished the species. During the twentieth century, conservation efforts re-established the beaver throughout their native range, but at population levels of only 10 per cent of the pre-European-contact level. As beavers continue to reoccupy their former range, their geomorphic influences also grow. For example, on the Roanoke River floodplain, on the Coastal Plain of the eastern United States, beaver ponds illustrated a tenfold increase in areal coverage between the years 1984 and 1993 (Townsend and Butler 1996).

Beaver dams impound water, creating pond environments into which streams flow and deposit sediment. Individual beaver ponds may accumulate as much as 20–30 cm of sediment across the floor of a pond annually (but only during the portions of the year when the inflowing stream is not frozen). Older ponds entrap significantly more sediment than do younger ponds in topographically similar sites (Butler and Malanson 1995). Beaver dams substantially reduce stream velocity and discharge, to the point in some cases of completely eliminating surface outflow on the downstream side of a dam (Meentemeyer and Butler 1999).

Animal burrows (such as wombats in Australia) and mounds (termites) may cover as much as several cubic metres, and are visible on aerial photographs and even satellite imagery (Löffler and Margules 1980). Some of the most widespread geomorphic impacts of animals are produced on the floor of the Bering Sea by the California gray whale (*Eschrichtius robustus*). Gray whales feed almost exclusively on bottom-dwelling organisms, especially amphipod crustaceans. The whales 'slurp' deep furrows and pits into the floor of the ocean, ingesting sediment and fauna. The whales filter the food from the sediment in their baleen, and subsequently expel plumes of muddy sediment near the surface (Butler 1995). These plumes, visible from low-flying airplanes, account for an enormous amount of sediment transport in the Bering Sea. Annual whale feeding there moves a minimum of 120 million cubic metres of sediment, nearly three times the annual load of suspended sediment discharged into the Bering Sea by the Yukon River, the fourth largest sediment source in North America.

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SEE ALSO: biogeomorphology; termites and termitaria

DAVID R. BUTLER

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