support the idea of instability of marine ice sheets. However, permitting a jump in ice thickness at the transition zone means the ice sheet system can be in neutral equilibrium, with an infinite number of steady-state profiles.

Another kind of transition zone is a smooth one, where the basal traction varies gradually, for example, along an ice stream. The equilibrium dynamics have not been worked out for a full three-dimensional flow, but it seems that the presence of ice streams destroys the neutral equilibrium and helps to stabilize marine ice sheets. This emphasizes the importance of understanding the dynamics of ice streams and their role in the marine ice sheet system.

Attempts to include moving grounding lines in whole ice sheet models suffer from incomplete specification of the problem. Assumptions built into the models predispose the results to be either too stable or too unstable. Thus there are no reliable models that can analyze the glaciological history of the Antarctic Ice Sheet. Predictive models rely on linearizations to provide acceptable accuracy for the near future but become progressively less accurate the longer the timescale.

See also

Bottom Water Formation. Icebergs. Ice–Ocean Interaction. Sub Ice-shelf Circulation and Processes.

Further Reading

Doake CSM and Vaughan DG (1991) Rapid disintegration of Wordie Ice Shelf in response to atmospheric warming. *Nature* 350: 328–330.

Doake CSM, Corr HFJ, Rott H, Skvarca P and Young N (1998) Breakup and conditions for stability of the northern Larsen Ice Shelf, Antarctica. *Nature* 391: 778–780.

Hindmarsh RCA (1993) Qualitative dynamics of marine ice sheets. In: Peltier WR (ed.) *Ice in the Climate System*, NATO ASI Series, vol. 12, pp. 67–99. Berlin: Springer-Verlag.

Kellogg TB and Kellogg DE (1987) Recent glacial history and rapid ice stream retreat in the Amundsen Sea. *Journal of Geophysical Research* 92: 8859–8864.

Robin G de Q (1979) Formation, flow and disintegration of ice shelves. *Journal of Glaciology* 24(90): 259–271.

Scambos TA, Hulbe C, Fahnestock M and Bohlander J (2000) The link between climate warming and break-up of ice shelves in the Antarctic Peninsula. *Journal of Glaciology* 46(154): 516–530.

van der Veen CJ (ed.) (1997) *Calving Glaciers: Report of a Workshop 28 February–2 March 1997*. BPRC Report No. 15. Columbus, Ohio: Byrd Polar Research Center, The Ohio State University.

van der Veen CJ (1999) *Fundamentals of Glacier Dynamics*. Rotterdam: A.A. Balkema.

Vaughan DG and Doake CSM (1996) Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula. *Nature* 379: 328–331.

IGNEOUS PROVINCES

M. F. Coffin, University of Texas at Austin, Austin, TX, USA
O. Eldholm, University of Oslo, Oslo, Norway

Copyright © 2001 Academic Press

doi:10.1006/rwos.2001.0463

Introduction

Large igneous provinces (LIPs) are massive crustal emplacements of predominantly Fe- and Mg-rich (mafic) rock that form by processes other than normal seafloor spreading. LIP rocks are readily distinguishable from the products of the two other major types of magmatism, midocean ridge and arc, on the Earth's surface on the basis of petrologic, geochemical, geochronologic, geophysical, and physical volcanological data. LIPs occur on both the continents and oceans, and include continental flood basalts, volcanic passive margins, oceanic plateaus, submarine ridges, seamounts, and ocean basin flood basalts (**Figure 1**, **Table 1**). LIPs and their small-scale analogs, hot spots, are commonly attributed to decompression melting of hot, low density mantle material known as mantle plumes. This type of magmatism currently represent $\sim 10\%$ of the mass and energy flux from the Earth's deep interior to its crust. The flux may have been higher in the past, but is episodic over geological time, in contrast to the relatively steady-state activity at seafloor spreading centers. Such episodicity reveals dynamic, non-steady-state circulation within the Earth's mantle, and suggests a strong potential for LIP emplacements to contribute to, if not instigate, major environmental changes.

Composition, Physical Volcanology, Crustal Structure, and Mantle Roots

LIPs are defined by the characteristics of their dominantly Fe- and Mg-rich (mafic) extrusive rocks; these most typically consist of subhorizontal, subaerial basalt flows. Individual flows can extend

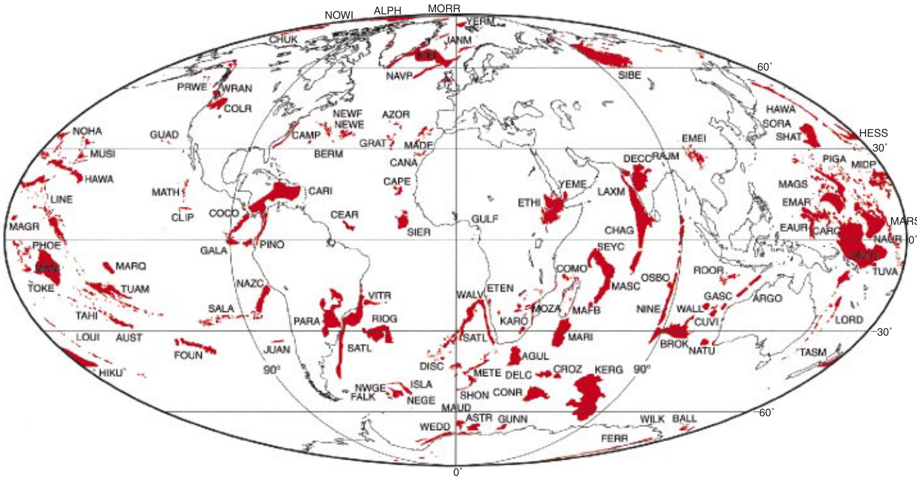

Figure 1 Phanerozoic global LIP distribution (red), with LIPs labeled (**Table 1**).

for hundreds of kilometers, be 10s to 100s of meters thick, and have volumes as great as 10^4–$10^5\,km^3$. Si-rich rocks also occur as lavas and intrusive rocks, and are mostly associated with the initial and late stages of LIP magmatic activity. Relative to midocean ridge basalts, LIPs include higher MgO lavas, basalts with more diverse major element compositions, rocks with more common fractionated components, both alkalic and tholeiitic differentiates, basalts with predominantly flat light rare earth element patterns, and lavas erupted in both subaerial and submarine settings.

As the extrusive component of LIPs is the most accessible for study, nearly all of our knowledge of LIPs is derived from lavas forming the uppermost 10% of LIP crust. The extrusive layer may exceed 10 km in thickness. On the basis of geophysical, predominantly seismic data from LIPs, and from comparisons with normal oceanic crust, LIP crust beneath the extrusive layer is believed to consist of an intrusive layer and a lower crustal body, characterized by compressional wave velocities of 7.0–7.6 $km\,s^{-1}$, at the base of the crust (**Figure 2**). Beneath continental crust this body may be considered as a magmatically underplated layer. Seismic wave velocities suggest an intrusive layer that is most likely gabbroic, and a lower crust that is ultramafic. If the LIP forms on pre-existing continental or oceanic crust or along a divergent plate boundary, dikes and sills are probably common in the middle and upper crust. The maximum crustal thickness, including extrusive, intrusive, and the lower crustal body, of an oceanic LIP is ~ 35 km, determined from seismic and gravity studies of the Ontong Java Plateau (**Figure 1**, **Table 1**).

Low-velocity zones have been observed recently in the mantle beneath the oceanic Ontong Java Plateau, as well as under the continental Deccan Traps and Paraná flood basalts (**Figure 1**, **Table 1**). Interpreted as lithospheric roots or keels, the zones can extend to at least 500–600 km into the mantle. In contrast to high-velocity roots beneath most continental areas, and the absence of lithospheric keels in most oceanic areas, the low-velocity zones beneath LIPs apparently reflect residual chemical and perhaps thermal effects of mantle plume activity. High-buoyancy roots extending well into the mantle beneath oceanic LIPs would suggest a significant role in continental growth via accretion of oceanic LIPs to the edges of continents.

Distribution, Tectonic Setting, and Types

LIPs occur worldwide, in both continental and oceanic crust in purely intraplate settings, and along present and former plate boundaries (**Figure 1**, **Table 1**), although the tectonic setting of formation is unknown for many features. If a LIP forms at a plate boundary, the entire crustal section is LIP crust (**Figure 2**). Conversely, if one forms in an intraplate setting, the pre-existing crust must be intruded and sandwiched by LIP magmas, albeit to an extent not resolvable by current geological or geophysical techniques.

Continental flood basalts, the most intensively studied LIPs due to their exposure, are erupted from fissures on continental crust (**Figure 1**, **Table 1**). Most continental flood basalts overlie sedimentary basins that formed via extension, but it is not clear what happened first, the magmatism or the extension. Volcanic passive margins form by excessive magmatism during continental breakup along the trailing, rifted edges of continents. In the deep ocean

Table 1 Large igneous provinces

Large igneous province	*Abbreviation* (**Figure 1**)	*Type*
Agulhas Ridge	AGUL	SR
Alpha-Mendeleyev Ridge	ALPH	SR/OP
Argo Basin	ARGO	VM
Astrid Ridge	ASTR	VM
Austral Seamounts	AUST	SMT
Azores	AZOR	SMT
Balleny Islands	BALL	SMT
Bermuda Rise	BERM	OP
Broken Ridge	BROK	OP
Canary Islands	CANA	SMT
Cape Verde Rise	CAPE	OP
Caribbean Flood Basalt	CARI	OBFB (partly accreted)
Caroline Seamounts	CARO	SMT
Ceara Rise	CEAR	OP
Central Atlantic Magmatic Province (VM only)	CAMP	CFB/VM
Chagos-Laccadive Ridge	CHAG	SR
Chukchi Plateau	CHUK	OP
Clipperton Seamounts	CLIP	SMT
Cocos Ridge	COCO	SR
Columbia River Basalt	COLR	CFB
Comores Archipelago	COMO	SMT
Conrad Rise	CONR	OP
Crozet Plateau	CROZ	OP
Cuvier (Wallaby) Plateau	CUVI	VM
Deccan Traps	DECC	CFB/VM
Del Caño Rise	DELC	OP
Discovery Seamounts	DISC	SMT
Eauripik Rise	EAUR	OP
East Mariana Basin	EMAR	OBFB
Emeishan Basalts	EMEI	CFB
Etendeka	ETEN	CFB
Ethiopian Flood Basalt	ETHI	CFB
Falkland Plateau	FALK	VM
Ferrar Basalts	FERR	CFB
Foundation Seamounts	FOUN	SMT
Galapagos/Carnegie Ridge	GALA	SMT/SR
Gascoyne Margin	GASC	VM
Great Meteor-Atlantis Seamounts	GRAT	SMT
Guadelupe Seamount Chain	GUAD	SMT
Gulf of Guinea	GULF	VM
Gunnerus Ridge	GUNN	VM
Hawaiian-Emperor Seamounts	HAWA	SMT
Hess Rise	HESS	OP
Hikurangi Plateau	HIKU	OP
Iceland/Greenland–Scotland Ridge	ICEL	OP/SR
Islas Orcadas Rise	ISLA	SR
Jan Mayen Ridge	JANM	VM
Juan Fernandez Archipelago	JUAN	SMT
Karoo	KARO	CFB
Kerguelen Plateau	KERG	OP/VM
Laxmi Ridge	LAXM	VM
Line Islands	LINE	SMT
Lord Howe Rise Seamounts	LORD	SMT
Louisville Ridge	LOUI	SMT

continued

Table 1 *continued*

Large igneous province	*Abbreviation* (**Figure 1**)	*Type*
Madagascar Flood Basalts	MAFB	CFB
Madagascar Ridge	MARI	SR/VM?
Madeira Rise	MADE	OP
Magellan Rise	MAGR	OP
Magellan Seamounts	MAGS	SMT
Manihiki Plateau	MANI	OP
Marquesas Islands	MARQ	SMT
Marshall Gilbert Seamounts	MARS	SMT
Mascarene Plateau	MASC	OP
Mathematicians Seamounts	MATH	SMT
Maud Rise	MAUD	OP
Meteor Rise	METE	SR
Mid-Pacific Mountains	MIDP	SMT
Morris Jesup Rise	MORR	VM
Mozambique Basin	MOZA	VM
Musicians Seamounts	MUSI	SMT
Naturaliste Plateau	NATU	VM
Nauru Basin	NAUR	OBFB
Nazca Ridge	NAZC	SR
New England Seamounts	NEWE	SMT
Newfoundland Ridge	NEWF	VM
Ninetyeast Ridge	NINE	SR
North Atlantic Volcanic Province	NAVP	CFB
Northeast Georgia Rise	NEGE	OP
Northwest Georgia Rise	NWGE	OP
Northwest Hawaiian Ridge	NOHA	SR/SMT
Northwind Ridge	NOWI	SR
Ontong Java Plateau	ONTO	OP (partly accreted)
Osborn Knoll	OSBO	OP
Paraná	PARA	CFB
Phoenix Seamounts	PHOE	SMT
Pigafetta Basin	PIGA	OBFB
Piñón Formation (Ecuador)	PINO	OP (accreted)
Pratt-Welker Seamounts	PRWE	SMT
Rajmahal Traps	RAJM	CFB
Rio Grande Rise	RIOG	OP
Roo Rise	ROOR	OP
Sala y Gomez Ridge	SALA	SR
Seychelles Bank	SEYC	VM
Shatsky Rise	SHAT	OP
Shona Ridge	SHON	SR
SiberianTraps	SIBE	CFB
Sierra Leone Rise	SIER	OP
Sorachi Plateau (Japan)	SORA	OP (accreted)
South Atlantic Margins	SATL	VM
Tahiti	TAHI	SMT
Tasmantid Seamounts	TASM	SMT
Tokelau Seamounts	TOKE	SMT
Tuamotu Archipelago	TUAM	SMT
Tuvalu Seamounts	TUVA	SMT
Vitória-Trindade Ridge	VITR	SR/SMT
Wallaby Plateau (Zenith Seamount)	WALL	OP
Walvis Ridge	WALV	SR
Weddell Sea	WEDD	VM
Wrangellia	WRAN	OP (accreted)
Yemen Plateau Basalts	YEME	CFB
Yermak Plateau	YERM	VM
Wilkes Land Margin	WILK	VM

CFB, continental flood basalt; OBFB, ocean basin flood basalt; OP, oceanic plateau; SMT, seamount; SR, submarine ridge

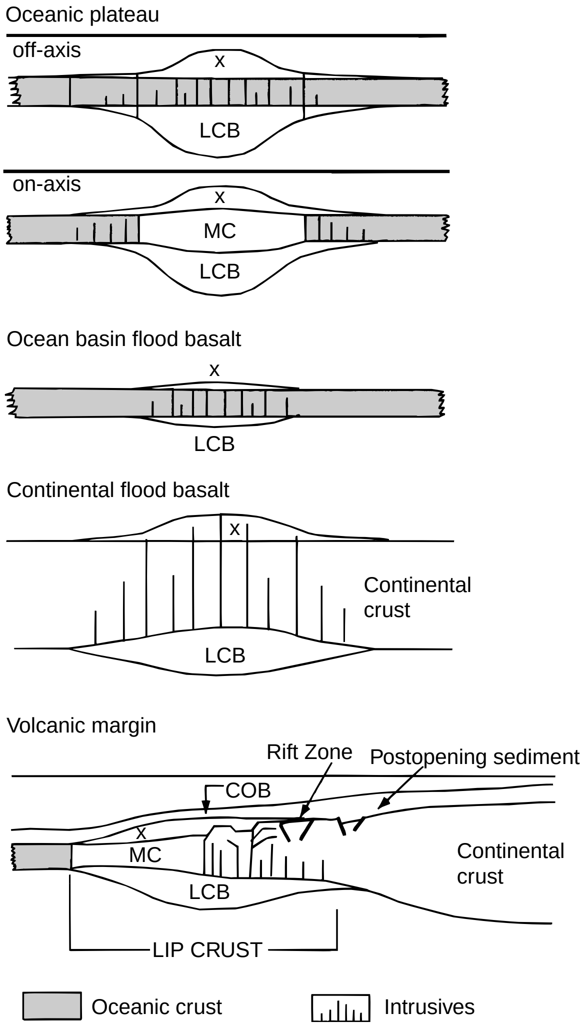

Figure 2 Schematic LIP plate tectonic settings and gross crustal structure. LIP crustal components are: extrusive cover (X), middle crust (MC), and lower crustal body (LCB), continent-ocean boundary (COB), Normal oceanic crust is gray.

basins, four types of LIPs are found. Oceanic plateaus, commonly isolated from major continents, are broad, typically flat-topped features generally lying 2000 m or more above the surrounding seafloor. They can form at triple junctions (e.g., Shatsky Rise), midocean ridges (e.g., Iceland), or in intraplate settings (e.g., northern Kerguelen Plateau). Submarine ridges are elongated, steep-sided elevations of the seafloor. Some form along transform plate boundaries, e.g., Ninetyeast Ridge. In the oceanic realm, oceanic plateaus and submarine ridges are the most enigmatic with respect to the tectonic setting in which they are formed. Seamounts, closely related to submarine ridges, are local elevations of the seafloor; they may be discrete, form a linear or random grouping, or be connected along their bases and aligned along a ridge or rise. They commonly form in intraplate regions, e.g., Hawaii. Ocean basin flood basalts, the least studied type of LIP, are extensive submarine flows and sills lying above and postdating normal oceanic crust.

Ages

Age control for all LIPs except continental flood basalts is sparse due to their relative inaccessibility, but the $^{40}Ar/^{39}Ar$ dating technique is having a particularly strong impact on studies of LIP volcanism. Geochronological studies of continental floor basalts (e.g., Siberian, Karoo/Ferrar, Deccan, Columbia River; **Figure 1**) suggest that most LIPs result from mantle plumes which initially transfer huge volumes ($\sim 10^5$–10^7 km^3) of mafic rock into localized regions of the crust over short intervals ($\sim 10^5$–10^6 years), but which subsequently transfer mass at a far lesser rate, albeit over significantly longer intervals (10^7–10^8 years). Transient magmatism during LIP formation is commonly attributed to mantle plume 'heads' reaching the crust following transit through all or part of the Earth's mantle, whereas persistent magmatism is considered to result from steady-state mantle plume 'tails' penetrating the lithosphere which is moving relative to the plume (**Figure 3**). However, not all LIPs have obvious connections to mantle plumes or hot spots, suggesting that more than one source model may be required to explain all LIPs.

LIPs are not distributed uniformly in time. During the past 150 million years for example, many LIPs formed between 50 and 150 million years ago, whereas few have formed during the past 50 million years (**Figure 4**). Such episodicity likely reflects variations in rates of mantle circulation, and this is supported by high rates of seafloor spreading during a portion of the 50–150 million year interval. Thus, although LIPs manifest types of mantle processes distinct from those resulting in seafloor spreading, waxing and waning rates of overall mantle circulation probably affect both sets of processes. A major question that emerges from the global LIP production rate is whether the mantle is circulating less vigorously as the Earth ages.

LIPs and Mantle Dynamics

The formation of various sizes of LIPs in a variety of tectonic settings on both continental and oceanic lithosphere suggests a variety of thermal anomalies in the mantle that give rise to LIPs as well as strong lithospheric control on their formation. Equivalent mantle plumes beneath continental and oceanic lithosphere should produce more magmatism in the latter scenario, as oceanic lithosphere is thinner,

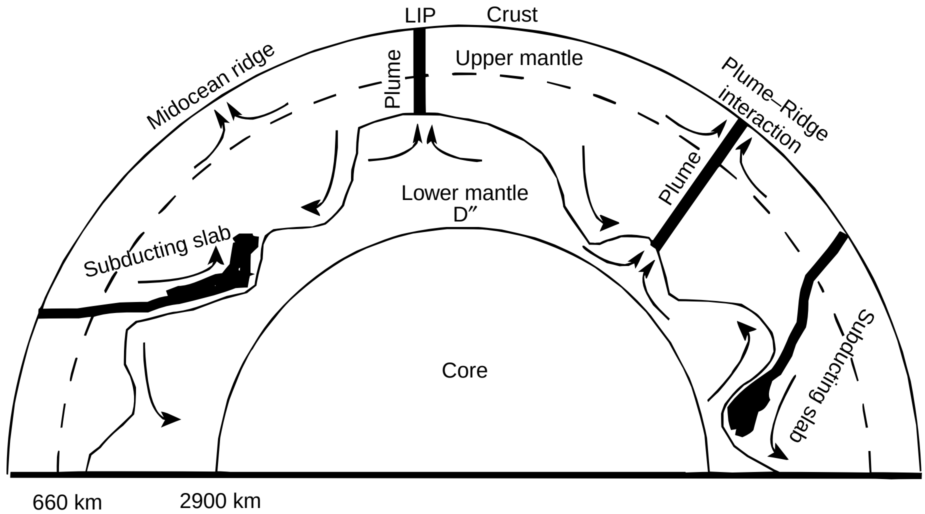

Figure 3 Model of the Earth's interior showing plumes, subducting slabs, and two mantle layers that move in complex patterns, but never mix.

allowing more decompression melting. Similarly, equivalent mantle plumes beneath an intraplate region (e.g., Hawaii) and a divergent plate boundary (e.g., Iceland) (**Figure 1**, **Table 1**) will produce more magmatism in the latter setting, again because decompression melting is enhanced. Recent seismic tomographic images of mantle plumes beneath Iceland and Hawaii show significant differences between the two.

Only recently, seismic tomography has revealed that slabs of subducting lithosphere can penetrate the entire Earth's mantle to the D″ layer at the boundary between the mantle and core at ~2900 km depth (**Figure 3**). If we assume that the volume of the Earth's mantle remains roughly constant through geological time, then the mass of crustal material fluxing into the mantle must be balanced by an equivalent mass of material fluxing from the mantle to the crust. Most, if not all of the magmatism associated with the plate tectonic processes of seafloor spreading and subduction is believed, on the basis of geochemistry and seismic tomography, to be derived from the upper mantle (above ~660 km depth). It is most reasonable to assume that the lithospheric material that enters the lower mantle is eventually recycled, in some part contributing to plume magmas.

LIPs and the Environment

The formation of LIPs has had documented environmental effects both locally and regionally. The global effects are less well understood, but the formation of some LIPs may have affected the global environment, particularly when conditions were at or near a threshold state. Eruption of enormous volumes of basaltic magma during LIP formation releases volatiles such as CO_2, S, Cl, and F (**Figure 5**).

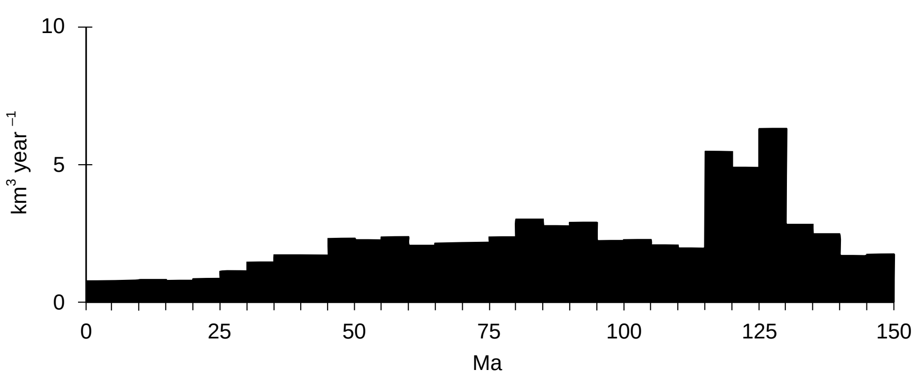

Figure 4 LIP production, corrected for subduction and averaged over a 15 million year running window, since 150 million years ago (Ma).

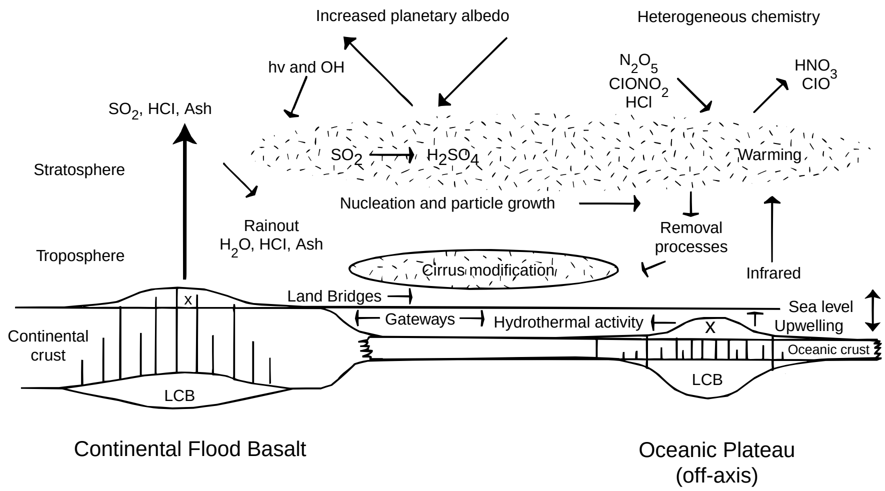

Figure 5 Environmental effects of LIP formation. LIP eruptions can perturb the Earth–ocean–atmosphere system significantly. Note that many oceanic plateaus form at least in part subaerially. LCB, lower crustal body; X, extrusive cover.

A key factor affecting the magnitude of volatile release is whether eruptions are subaerial or submarine; hydrostatic pressure inhibits vesiculation and degassing of relatively soluble volatile components (H_2O, S, Cl, F) during deep-water submarine eruptions, although low solubility components (CO_2, noble gases) are mostly degassed even at abyssal depths. Investigations of volcanic passive margins and oceanic plateaus have demonstrated widespread and voluminous subaerial basaltic eruptions.

Another important factor in the environmental impact of LIP volcanism is the latitude at which the LIP forms. In most basaltic eruptions, released volatiles remain in the troposphere. However, at high latitudes, the tropopause is relatively low, allowing large mass flux, basaltic fissure eruption plumes to transport SO_2 and other volatiles into the stratosphere. Sulfuric acid aerosol particles that form in the stratosphere after such eruptions have a longer residence time and greater global dispersal than if the SO_2 remains in the troposphere; therefore they have greater effects on climate and atmospheric chemistry. The large volume of subaerial basaltic volcanism, over relatively brief geological intervals, at high-latitude LIPs would contribute to potential global environmental effects.

Highly explosive felsic eruptions, such as those documented from volcanic passive margins, an oceanic plateau (Kerguelen) (**Figure 1**, **Table 1**), and continental flood basalt provinces, can also inject both particulate material and volatiles (SO_2, CO_2) directly into the stratosphere. The total volume of felsic volcanic rocks in LIPs is poorly constrained, but they may account for a small, but not negligible fraction of the volcanic deposits in LIPs. Significant volumes of explosive felsic volcanism would further contribute to the effects of plume volcanism on the global environment.

Between ~ 145 and ~ 50 million years ago, the global oceans were characterized by variations in chemistry, relatively high temperatures, high relative sea level, episodic deposition of black shales, high production of hydrocarbons, mass extinctions of marine organisms, and radiations of marine flora and fauna (**Figure 6**). Temporal correlations between the intense pulses of igneous activity associated with LIP formation and environmental changes suggest a causal relationship. Perhaps the most dramatic example is the eruption of the Siberian flood basalts (**Figure 1**, **Table 1**) ~ 250 million years ago, coinciding with the largest extinction of plants and animals in the geological record. Around 90% of all species became extinct at that time. On Iceland, the 1783–84 eruption of Laki provides the only human record of experience with the type of volcanism that constructs LIPs. Although Laki produced a basaltic lava flow representing $\sim 1\%$ of the volume of a typical ($10^3\,km^3$) LIP flow, the eruption's environmental impact resulted in the deaths of 75% of Iceland's livestock and 25% of its population from starvation.

Conclusions

Oceanic plateaus, volcanic passive margins, submarine ridges, seamounts, ocean basin flood basalts, and continental flood basalts share geological and geophysical characteristics indicating an origin distinct from igneous rocks formed at midocean ridges and arcs. These characteristics include: (1) broad areal extent ($> 10^4\,km^2$) of Fe- and Mg-rich lavas; (2)

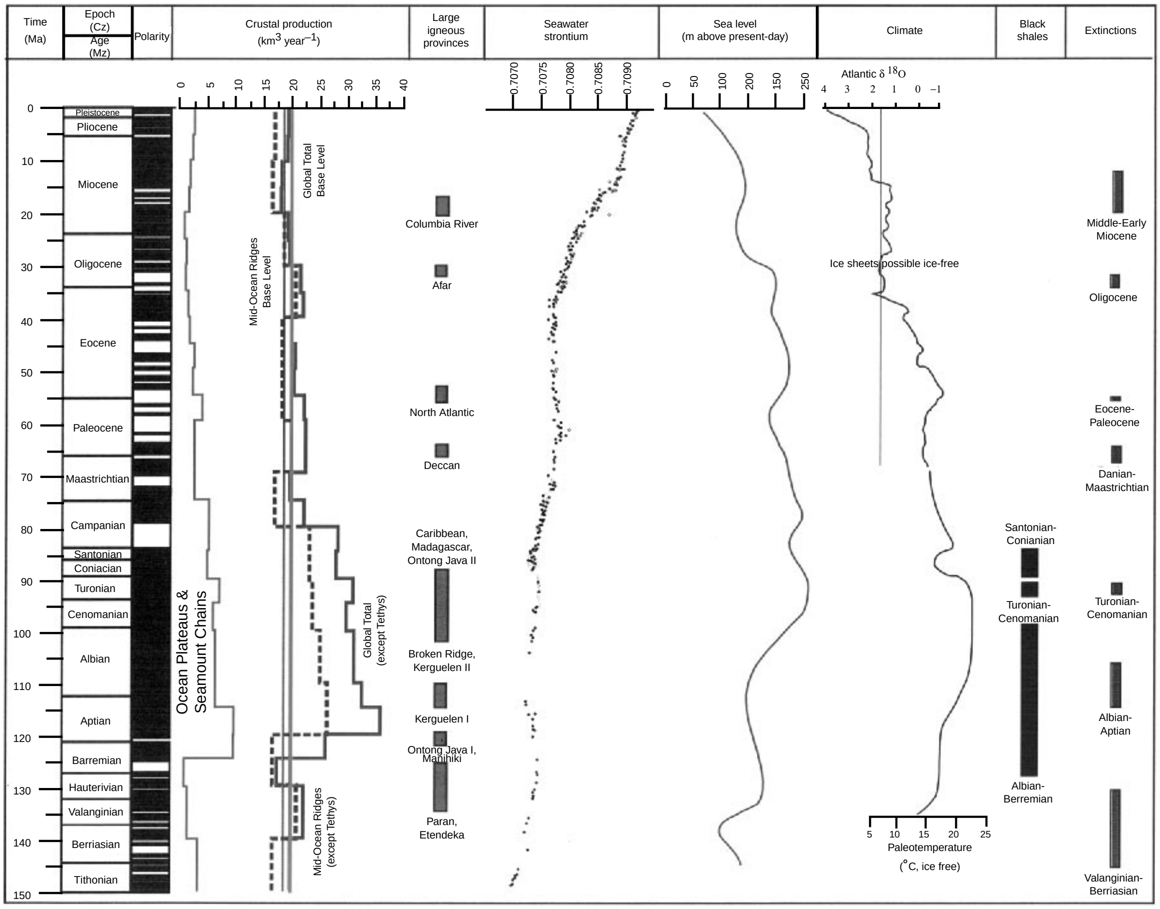

Figure 6 Temporal correlations among geomagnetic polarity, crustal production rates, LIPs, seawater strontium (Sr), sea level, climate, black shales, and extinctions.

massive transient basaltic volcanism occurring over 10^5–10^6 years; (3) persistent basaltic volcanism from the same source lasting 10^7–10^8 years; (4) lower crustal bodies characterized by compressional wave velocities of 7.0–7.6 km s^{-1}; (5) some component of more Si-rich volcanic rocks; (6) higher MgO lavas, basalts with more diverse major element compositions, rocks with more common fractionated components, both alkalic and tholeiitic differentiates, and basalts with predominantly flat light rare earth element patterns, all relative to midocean ridge basalts; (7) thick (10s–100s of meters) individual basalt flows; (8) long ($\leq$ 750 km) single basalt flows; and (9) lavas erupted in both subaerial and submarine settings.

There is strong evidence that LIPs both manifest a fundamental mode of mantle circulation commonly distinct from that which characterizes plate tectonics, and contribute episodically, at times catastrophically, to global environmental change. Nevertheless, it is important to bear in mind that we have literally only scratched the surface of oceanic, as well as continental LIPs.

See also

Deep Sea Drilling Results. Geophysical Heat Flow. Gravity. Magnetics. Mid-ocean Ridge Geochemistry and Petrology. Mid-Ocean Ridge Seismic Structure. Mid-Ocean Ridge Tectonics, Volcanism and Geomorphology. Propagating Rifts and Microplates. Seamounts and Off-ridge Volcanism. Seismic Structure.

Further Reading

Carlson RW (1991) Physical and chemical evidence on the cause and source characteristics of flood basalt volcanism. *Australian Journal of Earth Sciences* 38: 525–544.

Campbell IH and Griffiths RW (1990) Implications of mantle plume structure for the evolution of flood basalts. *Earth and Planetary Science Letters* 99: 79–93.

Coffin MF and Eldholm O (1994) Large igneous provinces: crustal structure, dimensions, and external consequences. *Reviews of Geophysics* 32: 1–36.
Cox KG (1980) A model for flood basalt vulcanism. *Journal of Petrology* 21: 629–650.
Davies GF (2000) *Dynamic Earth: Plates, Plumes and Mantle Convection*. New York: Cambridge University Press.
Duncan RA and Richards MA (1991) Hotspots, mantle plumes, flood basalts, and true polar wander. *Reviews of Geophysics* 29: 31–50.
Hinz K (1981) A hypothesis on terrestrial catastrophes: wedges of very thick oceanward dipping layers beneath passive continental margins – their origin and paleoenvironmental significance. *Geologisches Jahrbuch* E22: 3–28.
Macdougall JD (ed.) (1989) *Continental Flood Basalts*. Dordrecht: Kluwer Academic Publishers.
Mahoney JJ and Coffin MF (eds) (1997) *Large Igneous Provinces: Continental, Oceanic, and Planetary Flood Volcanism*. Washington: American Geophysical Union Geophysical Monograph 100.
Morgan (1981) Hotspot tracks and the opening of the Atlantic and Indian oceans. In: Emiliani C (ed.) *The Oceanic Lithosphere, The Sea*, Vol. 7, pp. 443–487. New York: John Wiley.
Richards MA, Duncan RA and Courtillot VE (1989) Flood basalts and hot-spot tracks: plume heads and tails. *Science* 246: 103–107.
Saunders AD, Tarney J, Kerr AC and Kent RW (1996) The formation and fate of large oceanic igneous provinces. *Lithos* 37: 81–95.
Sigurdsson H, Houghton BF, McNutt SR, Rymer H and Stix J (eds) (2000) *Encyclopedia of Volcanoes* San Diego: Academic Press: 1147pp.
Sleep NH (1992) Hotspot volcanism and mantle plumes. *Annual Review of Earth and Planetary Science* 20: 19–43.
White R and McKenzie D (1989) Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research* 94: 7685–7729.
White RS and McKenzie D (1995) Mantle plumes and flood basalts. *Journal of Geophysical Research* 100: 17543–17585.

INDIAN OCEAN EQUATORIAL CURRENTS

M. Fieux, Université Pierre et Marie Curie, Paris Cedex, France

Copyright © 2001 Academic Press

doi:10.1006/rwos.2001.0367

Introduction

Dynamically the equatorial area is a singular region on the earth because the Coriolis force is small, vanishing exactly at the equator. This results in a current structure that differs from that at other latitudes. Moreover, the equatorial current system of the Indian Ocean is entirely different from the current system found near the equator in the Pacific and Atlantic. This is principally due to its different wind forcing, which is described in the first section below. The systems of strictly equatorial currents at surface and at depth are reviewed in the second section. The third and fourth sections describe the North-East and South-West Monsoon Currents, north of the Equator, and the South Equatorial Countercurrent and the South Equatorial Current, south of the Equator.

The Atmospheric Circulation over the Equatorial Indian Ocean

The winds over the tropical Indian Ocean are quite different from the winds over the Atlantic and the Pacific tropical oceans, where the NE and SE trade winds blow always in the same direction. Instead, the Indian Ocean (**Figure 1**), north of 10°S, is under the influence of a monsoonal circulation, with complete reversal of the winds twice a year. The winter monsoon (December–March) blows from the NE in the Northern Hemisphere and from the NW south of the Equator toward the intertropical convergence zone (ITCZ) located near 10°S. The change in direction at the Equator comes from the change of sign of the Coriolis force. The summer monsoon (June–September) blows from the SW in the Northern Hemisphere in continuity with the SE trade winds of the Southern Hemisphere, particularly in the western part of the equatorial ocean. The winds are stronger during the summer monsoon season. At the equator, during the monsoons, the winds have a preponderant meridional component, southward during the winter monsoon and northward during the summer monsoon, particularly near the western boundary. The result is a strong annual cycle in the meridional component of the winds corresponding to the reversals between the NE and the SW monsoons, particularly in the western region along the Somali coast where the winds are the strongest.

Between the monsoons, during the two transition periods, at the Equator, moderate eastward winds blow in spring (April–May) and in fall (October–November), with maxima between 70°E and 90°E