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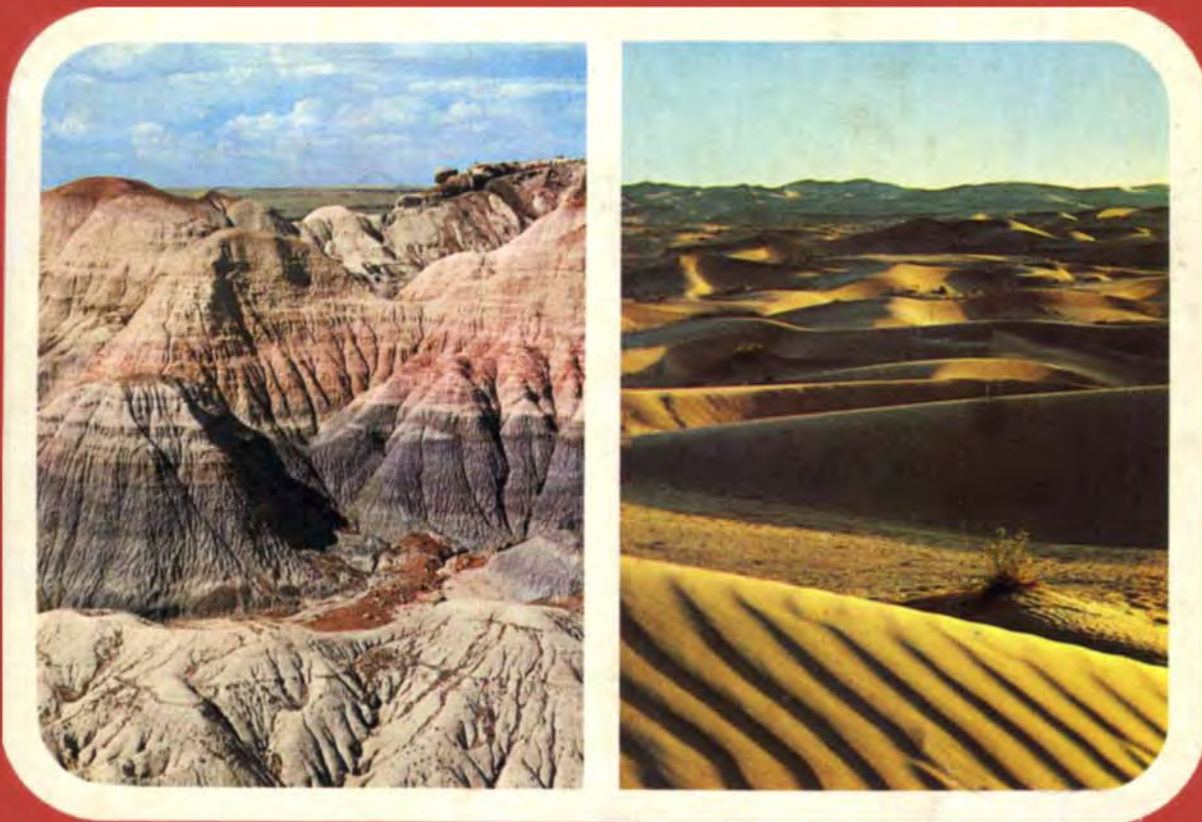
SOIL EROSION



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Developments in Soil Science 10

SOIL EROSION

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CONTENTS

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INTRODUCTION

Soil is the basis of production in agriculture and forestry, the nourisher of mankind and an important component of the human environment. Much more attention should therefore be given to the soil, and greater control with respect to its rational use, protection and improvement, needs to be exercised.

Challenging problems in soil conservation arise from the rapid growth of the world population and the ensuing requirement for increased food production on the one hand, and on the other, from rapid technical process giving rise to further industrialization and urbanization which involve the deterioration, destruction, intoxication, and contamination of the environment by industrial fumes and various chemical substances.

General consequences of these trends are the diminishing area of fertile land suitable for agriculture, the even more rapid decline of the per capita ratio of cultivated land, the rising proportion of deteriorated soil attributable to human activity, and the increase in soil levels of exogenous, often toxic substances which reduce on an ever increasing scale the soil's ability to perform its biotic functions in the *ecosphere,* i.e. in the *environment of biota.*

Sufficient food for a growing population may be obtained either by applying extensive farming techniques, i.e. by expansion of the area under cultivation, or by applying intensive methods which provide higher crop yields for a given area as a result of better utilization of the bioenergy potential of the soil.

Extensive farming methods which are possible mainly in sparsely populated regions and in marginal areas of the ecumene, involve the acquisition and cultivation of land by clearing the existing permanent vegetation cover, with little regard for the high soil-protective effect of the latter. The soils which become subjected to this process are usually deficient requiring high-cost improvement, are highly susceptible to erosion, or are situated in regions affected by highly destructive exogenous geomorphological processes attributable to the climate. Therefore the extensive expansion of cultivated agricultural land generally increases the danger of erosion.

Intensive farming usually involves large-scale production processes with a high level of mechanization, a preference for monoculture with a relatively lower protective effect on the soil, a rather narrow range of species, high doses of industrially manufactured fertilizers, various biocidal substances (mainly herbicides, insecticides and fungicides), and highly concentrated livestock production.

These intensive techniques on which most advanced countries depend have both positive and negative aspects from the standpoint of soil erosion and soil conservation. The positive aspects include decreases in the destruction normally caused by extensive cattle grazing, by unfavourable location of fields and boundaries, inappropriate agrotechnics, disadvantageous sitting, the building and maintenance of dirt roads, etc. Higher doses of fertilizers may also have a positive influence, especially natural fertilizers which increase the erosion resistance of the soil and soil permeability, and accelerate the growth of vegetation which is the most important factor in erosion control.

The negative aspects of intensive farming, when the latter is practised incorrectly, include increased surface runoff, reduced erosion resistance of the soil, increased amounts of eluates which increase the pollution of surface flows and result in eutrophication, and, ultimately, the general deterioration of the human environment. The same may be said of wind erosion.

A very special situation arises in those regions under direct attack from *industrial fumes* which, on the one hand, cause a deterioration of the site conditions, and therefore also of vegetation growth and, on the other, increase the level of pollutants and the intensity of erosion in general. In such cases intensive methods are of limited use and as a rule they tend only to increase erosion.

Intensive methods have similar effects, although on a smaller scale, in forestry.

Thus it seems that civilization, as it develops further, is going to face an increasing danger of accelerated man-made soil erosion, with increasingly harmful consequences for the human environment. If in the past erosion has been considered to be a soil disease, in the future it may become known rather as a *disease of the landscape,* to the extent of being a *pedophtoric* or *ecophtoric* phenomenon in some regions (Greek *pedon* – soil, *oikos* – husbandry, *ftora* – destruction, devastation).

Due attention should therefore be given to research in soil erosion. A characteristic of erosion is that it starts on cultivated farmland without any visible manifestations. When this happens, the danger is underestimated and erosion control measures are taken only in exceptional cases, or in those cases in which eroded soil has already lost its fertility.

Erosion control measures can be applied effectively only when the nature of the *erosion phenomena* and the effectiveness of measures under particular sets of conditions have been thoroughly studied. These questions are studied in the recently established *science of soil erosion,* or *soil erodology.* The aim of the theory is firstly to add to current knowledge the generalizing from the knowledge gained by observation of erosion phenomena, also including the principles of soil conservation, and secondly, to determine the best methods of improving the properties of eroded soil. In practice, erosion control is connected to a greater or lesser degree with improvement, and from the practical standpoint it is therefore possible to speak of erosion control soil *(amelioration)* (Kozmenko 1954, Żiemnicki 1968).

Soil erodology as a comprehensive assemblage of scientific information on erosion and erosion control is a young branch of science, yet the dangers of erosion and various methods of erosion control have been known to mankind since time immemorial. Up to the end of the 19th century this information was more or less empirical and confined locally. Only when a sufficient amount of practical and theoretical information had been collected, was it possible to develop a relatively comprehensive new theory in the form of a new scientific discipline.

The *development of erodology* as the theory of erosion in general has been complex, and was accelerated by specialists in many other fields. The broadest concept of *erosion* was developed by *geomorphologists, geographers* and *geologists,* who considered erosion mainly in terms of the development of the Earth's surface under the influence of exogenous forces (e.g. Penck 1894, Davis 1898, 1902, Lazarević 1973).

Pedologists began to study erosion in more concrete terms. The first to point out the dangers of erosion was Dokuchaev (1877, 1879), the founder of pedology. Wollny (1895) conducted the first interesting experiments on the effects of atmospheric precipitation on soil and soil wash. Specific research on rill and sheet erosion was first carried out by Kozmenko (1909, 1910), but the disastrous consequences of erosion for mankind were pointed out by the American soil conservationists Bennett and Chapline (1928). The subsequent years saw the beginnings of broadly-based, organized soil erosion research.

The third contribution of information on soil erosion towards the incipient erodological discipline came from the principles of *torrent and avalanche control.* These principles originated in the Alpine countries in the second half of the 19th century. The first specialists in this field were French. They, too, were the authors of guidelines on soil conservation in mountain regions and on the control of torrential floods (Reboisement des montagnes (1860), Gazonnement des montagnes, 1864). In addition to the works of Surell (1842), mention may be made of the classical work of Demontzey (1878, 1882) which became the basis for the rapid development of torrent control in many European countries. The promulgation of the Austro-Hungarian Act No. 117 in 1884 concerning the harmless deflection of water from the watershed is also linked with this work. Although the authors were mainly concerned with reducing load and protecting watercourses and watersheds against silt deposition, they gave a lot of attention also to the characteristics of erosion and to the erosion control.

Besides the three above-mentioned fields, specialists in other disciplines had also significantly contributed to the establishment of the new scientific discipline of erodology; these specialists were *water conservationists, glaciologists, geobotanists, agronomists, foresters,* and others. Hydrologists studied this phenomenon mainly from the standpoint of the development of rivers, lakes, and the silt pollution of water (Lopatin 1952, Makkaveev 1955), glaciologists investigated those aspects of erosion relating to surface formations and the destruction of soil by ice, snow, water, wind, and frost (Embleton and King 1968), geobotanists studied erosion in terms of the relationships between site conditions and vegetation, agronomists emphasized the importance of the conservation of agricultural land, and foresters studied erosion from the standpoint of the protection of forest soil, especially the improvement of forest management. Perhaps the broadest experience in soil conservation was derived from slope terracing which in many regions is a fundamental prerequisite of intensive agriculture.

Together with works dealing marginally with soil erosion within the bounds of another classical discipline, attempts were made at writing specialized erodological monographs, and later, text-books also. To date, there are tens of thousands of erodological publications of great diversity in subject and scope, dealing with different aspects of erosion and erosion control. Erosion phenomena and the conditions under which they originate, as well as the possibilities for erosion control in the various types of natural and cultivated area are so varied, that no individual, not even a most highly skilled group of specialists are any longer capable of producing an exhaustive work on this subject. The author for his part believes that such a work is not in any case needed, since only a part of the information that it would contain could be used for any one purpose. The work presented here should be viewed from this aspect, too.

In developing the concept of the science of *soil erosion,* the author considered that establishing a definition of soil erosion was of fundamental importance. The term soil erosion, first coined in English, was introduced by McGee in 1911. Later the first monographs on the subject appeared, among them *Soil Erosion and Its Control* by Ayres in 1936, and (in Russian) the extensive *Eroziya pochv* (Soil Erosion) published in 1937 by a number of co-authors. Studies in this field in other languages appeared much later, mostly after 1947.

H. H. Bennett, the American soil conservationist is generally considered to be the founder of the science of erosion which he introduced and established with his work *Soil Conservation* published in 1939, and later with his *Elements of Soil Conservation* published in 1955. In addition to these monographs a number of important studies on the subject have been published in English, among them works by Frewert et al. (1955), Archer (1956), Stallings (1957), Kohnke and Bertrand (1959), and Hudson (1971).

In the USSR, major studies in this field were published by Kozmenko (1948, 1954); an important contribution was the monograph by Sobolev (1948), and other works were published by Kocherga (1965), Zaslavskiĭ (1966), and Mirtskhulava (1970). In Italy, the valuable work by Oliva (1952) was published in three editions, and in Bulgaria, the work of Biolchev (1955) is of considerable importance. Other major contributors to the study of the subject are Moţoc (1956) in Romania, Żiemnicki and Józefaciuk (1965), Żiemnicki (1968) in Poland, Gavrilović (1972) in Yugoslavia, Schultze (1952), Glander (1956) in the German Democratic Republic, Kuron (1956), Richter (1965) in the German Federal Republic, and Furon (1947), Fournier (1960) in France.

In Czechoslovakia, where this work has been published, a number of other monographs and text-books on soil erosion have also been written (Spirhanzl 1952, Jůva and Cablík 1954, Cablík and Jůva 1963, Zachar 1960, 1970, Holý 1970, Riedl, Zachar et al. 1973, etc.).

In this work, the author faced the task of summing up in a brief survey all that is known with respect to terminology, classification, methodology, erosion factors, geography of soil erosion, and methods of erosion control in the pedosphere, being well aware of the fact that his work would be incomplete in every respect. He considers these questions to be fundamental in any comprehensive evaluation of erosion and erosion control, and believes that they will become the basis for the further development and improvement of theoretical and practical information provided by this new discipline.

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Chapter 1

BASIC TERMINOLOGY

1.1 The term erosion and its application

The word *erosion* is of Latin origin being derived from the verb *erodere* – to eat away (*rodere* – to gnaw), to excavate. The term erosion was first used in geology to describe the forming of hollows by water, the wearing away of solid material by the action of river water (Penck 1894), while surface wash and precipitation erosion was called *ablation* (Latin *ablatio* – to carry away). The problems of river erosion and its contribution towards the modelling of the Earth's surface were well understood by the end of the 19th century.

In addition to the terms erosion and ablation a number of other terms were used to express geomorphological processes caused by water and wind. These included the established terms *corrasion* (Latin *corradere* – to scrape together), *corrosion* (Latin *corrodere* – to gnaw to pieces), *abrasion* (Latin *abradere* – to scrape off), and *denudation* (Latin *denudere* – to strip), etc.

Whereas the term erosion was used to describe the process of washing away, usually in a vertical direction, the term corrasion meant mechanical, lateral washing away (by rivers); the term corrosion was used to refer to the chemical destruction of easily soluble rocks, the term abrasion referred to the process of scouring by sea water and wind, and finally the term denudation indicated the process of uncovering bare rocks, surface wash, and also sheet erosion.

Yet different authors have used these terms in different contexts and interchanged them thereby causing much ambiguity. Many authors now use the term erosion to encompass any form of destruction of soil or the Earth's surface by water, and recommend that the terms *deflation* and *abrasion* be used in cases of wind destruction.

This chapter attempts to present a more detailed account of the terms in use, and to classify them into comprehensive groups according to given criteria, the author is striving to distinguish between national and international terms in the case of each phenomenon. National terms help to maintain linguistic purity, whereas international terms promote international understanding, and make professional literature more intelligible. This task is, of course, a great challenge and has been accomplished here only as far as main categories of phenomena are concerned.

In this work the term *erosion* shall be used for the disruption of the soil mantle – the *pedosphere* (Greek *pedon* – soil), or the underlying rock base – the

Fig. 1. Grikes on nearly pure limestone in the Dinaric Kars caused by water corrosion and corrasion (Yugoslavia). (Photo D. Zachar.)

lithosphere (Greek *lithos* – stone, *sphaira* – ball) by the action of matter of exogenous origin, i.e. by *external geomorphic factors.*

In the broadest sense of the word these factors include *water, snow, ice, air (wind), weathered debris, organisms (plants* and *animals),* and *man.* These factors may be classified as biotic (Greek *bios* – life), i.e. relating to life, and *abiotic*

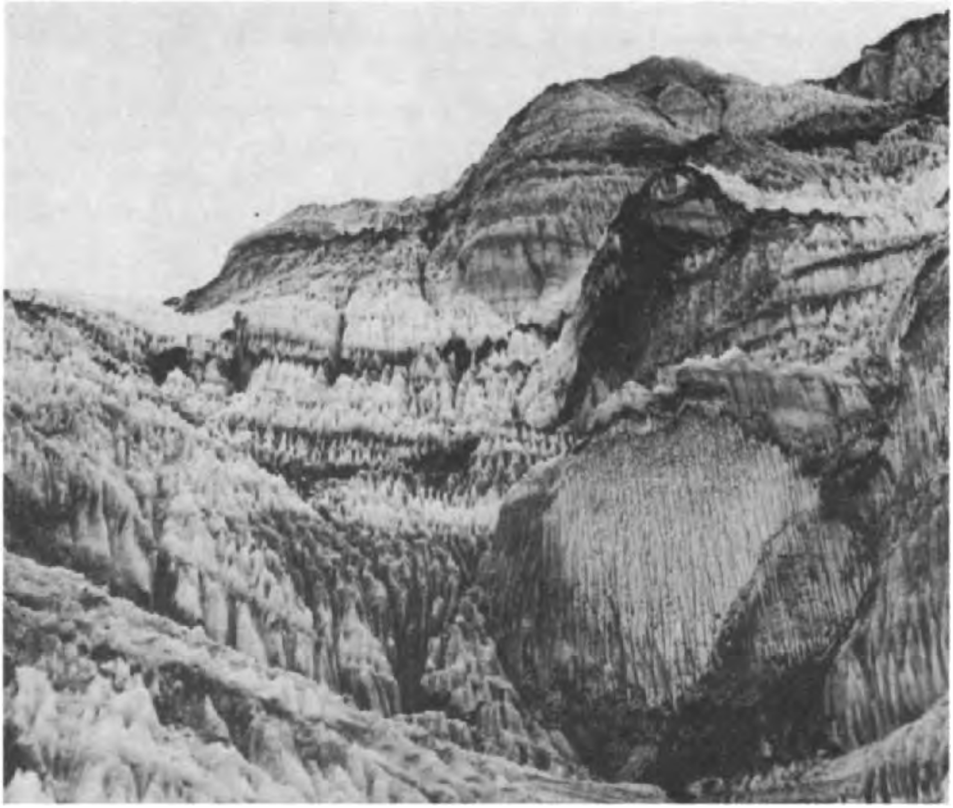


Fig. 2. Chemical erosion of salt strata (Salnik, Romania). (Photo D. Zachar.)

(inanimate). Most authors include within the term erosion only that destruction of the soil that is caused by abiotic factors, the activity of which is attributable to mechanical action, i.e. kinetic energy. Applying broader criteria, this type of erosion may be called *mechanical erosion,* i.e. *corrasion.* But erosion also includes chemical action which is connected also with the mechanical action of water (Figs. 1, 2). This type of erosion may be referred to as *chemical erosion,* i.e. *corrosion.*

In addition to these two main forms of erosion caused by abiotic factors *(abiotic erosion),* the *organogenic* aspects of *erosion* (Latin *organum,* Greek *organon* – organism, Greek *gennao* – I bear) may be distinguished and subdivided into *phytogenic* (Greek *fyton* – plant) (Lazarević 1975), *zoogenic* (Greek *zoon* – animal), and man-made, i.e. *anthropogenic* (Greek *anthropos* – man) *erosion.* The first two factors (plants and animals) do little more than nibble at the soil and substrate hence *arrosion* (Latin *arrodere* – to nibble). As far as erosion caused by man and animals is concerned, only that part of the activity caused by the natural factor is considered to constitute erosion.

In reality there are various intermediate types between these main types of erosion, so that *mechano-chemical erosion,* and *anthropo-zoogenic erosion,* etc. may be referred to.

In all situations several types of erosion always occur simultaneously or in some chronological sequence, forming patterns which are typical of a particular area. The decisive factors are the climate, the relief, the nature of the surface, and the activity of organism, especially the activity of man which has been responsible in recent years for an increasing specific influence on *erosion systems.*

1.2 The link between erosion and other landscape-modelling factors

Erosion is but one of a number of landscape-modelling factors. By its action the Earth's surface is either *being worn down* or *degraded* (Latin *degradatio* – degradation, reduction), or it is *being raised* or *aggraded* (Latin *aggradatio* – aggradation, raising) by deposition. The result of this process is a general *levelling* of the Earth's surface-planation (Latin *planare* – to level), and a condition of this levelling effect is the *disintegration* of matter in elevated regions of the Earth's crust. Disintegration (Latin *integer* – integral) involves the weathering of rocks – the first stage of the process of soil formation.

On steep slopes no external force is needed to set loosened *detritus* (Latin *detrahere* – to pull down) in motion. In many cases frost, high temperature, etc. separate pieces of weathered rock and the loose material moves downhill to form *piles of hill-side waste, debris cones, outwash fans,* and other formations. Because in most cases the rock is not completely weathered, the accumulated material is coarse-grained, gravelly, stony, or even bouldery. These phenomena are generally called *gravitational phenomena* (Latin *gravis* – heavy). They frequently occur in combination with erosion as *gravity erosion,* or *erosion gravity phenomena* (Fig. 3).

Similar phenomena are *slope crumbling* and *destruction* which result from tectonic processes, earthquakes, and other disturbances of a broader nature.

A different type of phenomenon arises in those situations in which large quantities of *deposits* and *hill-side wastes* have accumulated throughout geological periods. In addition to the rapid erosion of the surface layers in such cases, movement of matter caused by a loss of stability, reduced internal friction, increased weight, and weakening of the slope base may occur. Mass movement of weathered rock is generally referred to as soil slip, or *decerption* (Latin *decerpere* – to break off) (Murzaeva and Ryzhov 1962). The slipping or sliding of material disturbs the process of accretive erosion, and thus earthflows, in the same way as slips and creeps, enhance the effects of erosion.

Fig. 3. Combined gravitation and flow phenomena in the High Tatras (Czechoslovakia). (Photo D. Zachar.)

Fig. 4. Garland soils modelled by cryosolifluction and wash (Engadine National Park, Switzerland). (Photo J. Pelíšek.)

Fig. 5. Thermoerosion of the banks of the river Lena below Yakutsk (Peschannaya Gora, USSR). (Photo A. Jahn.)

A different set of phenomena occurs in areas with a periglacial climate, i.e. areas fashioned by *cryogenic modelling.* In such regions processes brought about by temperature changes are of chief importance. Among cryogenic phenomena some are closely connected with erosion, especially *regelation* (Latin *regelare* – to freeze again) which causes *soil flow, solifluction* (Latin *solum* – soil, *fluctus* – flow) in which soil destroyed by frost, saturated with water and turned slushy, flows down a frozen layer.

There is an extensive literature dealing with this problem and clarifying the pedoerosion process in periglacial regions in general. Among the most important works are those of Andersson (1906), Troll (1944a, b, 1948), Dylik (1951), Cailleux and Taylor (1954), Avenard (1961), Rapp (1960), Kaplina (1965), Hamelin and Cook (1967), and Embleton and King (1968). The most active factors in cryogenic erosion are water, snow, and wind, which together with frost give rise to a wide range of surface landforms. Among the best known *cryopedophenomena* (Greek *cryos* – ice, frost, cold, *pedon* – soil) are *crescent, garland* (Fig. 4), *terraced, hillocky, polygonal, girdled, circular,* and other soils.

This group of phenomena also includes various *niveo-aeolic* or *niveolic* (Hamelin and Clibbon 1962) *phenomena* caused by snow and wind (Latin *nivalis* – relating to snow, Greek *Áielos,* Latin *Aeolus* – the God of winds and storms), and the interesting process of *thermic erosion* (Greek *therme* – heat) which occurs mainly in rivers, and therefore could also be referred to as *cryofluvial* or *thermofluvial erosion* (Latin *fluvialis* – relating to the river) (Fig. 5). Besides cryofluvial erosion there are also *cryopluvial, thermopluvial* (Latin *pluvialis* – relating to rain), and *cryogenic erosion* processes.

There is no need to emphasize that the distinction between phenomena arising from *erosion, gravitation, land-slip, solifluction, cryogenic processes, nivation,* and other *phenomena* caused by natural forces is important not only from a theoretical point of view but also from a practical point of view. An even greater variety occurs when man interferes with natural processes, often changing their original character and setting them on a different course.

The contributions made by the various types and forms of erosion may differ widely in different cases, and therefore it is possible to distinguish between *prevalent* and *accessory erosion.* It may be said that accessory erosion occurs in almost all land-modelling processes, being a part of all polygenetic phenomena.

1.3 The term soil erosion and its basic components

Although the term *erosion* was in use in the 19th century, the term *soil erosion* was introduced later, at the beginning of the 20th century, and did not come into general use until the 1930s. The term was established and defined by Bennett, Fuller, Lowdermilk and Middleton in Anglo-American literature, Kozmenko, Pankov, Gussak, Sobolev, and Zaslavskiĭ in Russian literature, Kuron, Schultze, Glander, and Flegel in German, and Baulig in French literature.

The term *soil erosion* generally means the destruction of soil by the action of water and wind. Most authors dealing with problems of soil erosion include those phenomena related to the activity of man within the meaning of soil erosion. Some authors conceive soil erosion only as erosion caused by precipitation, while others include erosion caused by natural and man-made factors operating in conjunction. In this study soil erosion is taken to mean the destruction of soil by water, snow, ice, wind, animals, and man. *Soil erosion, eroziya pochvy, l'érosion du sol,* and *Bodenerosion* are considered to be equivalent terms.

For comparison, it may be mentioned that Bennett (1939) distinguishes between *normal,* i.e. *geological erosion* (sometimes referred to as *natural erosion*), and *accelerated erosion,* yet he considers only accelerated erosion as being soil erosion proper. Accelerated erosion is subdivided, according to this author, into *naturally accelerated* and *man-accelerated erosion.* Naturally accelerated erosion is caused by abnormal drought, avalanches, plant diseases, pests, etc.; the task of soil conservation schemes is to reduce man-accelerated erosion to the normal, or geological level, or, according to Lowdermilk (1935), to the *geological norm of erosion.*

Thus Bennett considers that soil erosion, in the true sense of the word, should only refer to that erosion which is more severe in intensity than the erosion that takes place in natural plant communities undisturbed by man or other factors. Using this interpretation, even a mild degree of erosion, if accelerated to a small extent by man (e.g. in humid regions) would be considered as erosion, while wind erosion in arid regions where the most severe types of erosion occur, would not be considered as erosion. It is, moreover, extremely difficult to assess in Bennett's classification what is normal and what is abnormal drought, and it is even more difficult to decide when this drought accelerates erosion so as to exceed the geological norm of erosion. These difficulties are pointed out by Schultze (1952).

From among the other methods of classifying the components of soil erosion arising from man's interference, mention should be made of Schultze's scheme (1952) in which *denudation* by the *removal* (Abtrag) of soil is considered to be either 1. the result of *man-accelerated soil erosion* (durch Menschen beschleunigte, kulturlandschaftlich gegebene Bodenerosion), or 2. *normal,* or *natural denudation* (normale, naturlandschaftlich gegebene Denudation). In describing the total removal of silt by rivers the author speaks only of denudation. Consequently, in the broad concept of denudation as a removal process, he considers soil erosion, in the more narrow sense of the word, to be only that erosion (denudation) which is accelerated by man.

The classifications of Bennett and Schultze differ in that the broad use of the term erosion has been replaced in the latter system by the term denudation, and that the term soil erosion is used to include only phenomena of man-accelerated erosion to the exclusion of those caused by natural anomalies. Moreover, Schultze does not distinguish between erosion and denudation nor between erosion and removal, or, for that matter, between denudation and removal. These differences in the use of terms may also be found in other studies which may therefore be guilty of etymological error, since the term erosion as a process does not relate to the soil or soil particles which are removed, but is concerned rather with soil which is destroyed or corroded, while remaining in situ. It seems more appropriate therefore to distinguish between the meanings of the terms erosion and wash off, erosion and float off, wind erosion and blow off, etc. *Removal, wash off, blow off, displacement,* and other forms of movement of weathered material which becomes *loosened* and *translocated* by erosion, are among the many forms of transport that occur within the planation cycle, and these are followed by the third and last stage of the cycle, namely *sedimentation* and *accumulation.* Thus, removal is directly linked to erosion which may also be expressed in terms of removal (e.g. the intensity of erosion). Removal may be complementary to other phenomena besides erosion. This means that erosion without removal cannot occur while removal without erosion is possible.

Taking all considerations into account, the categorization of erosion firstly into natural, normal, or geological erosion, and secondly into accelerated, or soil erosion, seems not to be a satisfactory solution.

It should be noted at the outset that all phenomena occurring in nature without human interference are, according to the literature, natural phenomena, not solely the "normal" processes which result in the formation of a "normal" soil profile. The *"normal" soil profile,* i.e. any profile with developed genetic horizons, is formed in nature only where conditions with respect to climate, hydrology, soil, and orography are favourable. Natural soils, i.e. soils which have developed without human interference, may, however, occur in areas where a "normal" soil profile does not exist. Consequently, the term "normal" is much narrower than the term "natural", and conversely, the term "geological" is broader than the term "natural" since man also acts as a geological factor.

Nor is it permissible to restrict the use of the term soil erosion solely to the context of accelerated erosion (which many authors take as the meaning of soil erosion), for the very reason that man may not only accelerate erosion, but he may also slow it down, and if this is the case even decelerated erosion must be considered as soil erosion. In some situations, man may accelerate erosion, in others he may suppress it by establishing plantations. This happens, for instance, when natural vegetation which inadequately protects the soil is replaced by artificial plantations (e.g. forests) which provide a more effective means of land improvement. The intensity of soil erosion may be reduced by applying measures which slow down accelerated erosion caused initially by man. It is not entirely correct, therefore, to identify any form of erosion modified by man with accelerated erosion.

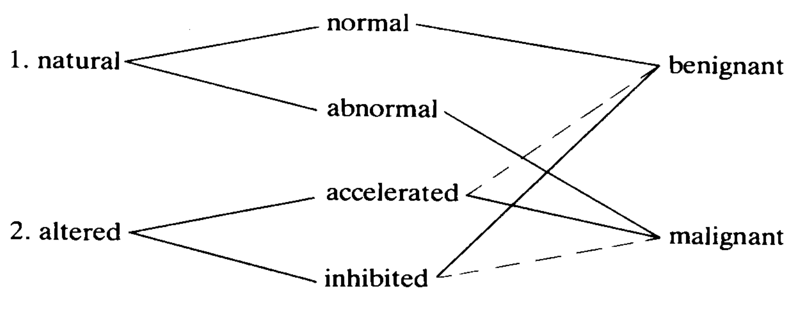
It is still less appropriate to identify the term accelerated erosion with soil erosion. Soil erosion, according to many authors, has occurred only since and where man has been engaged in agricultural activity. This concept of soil erosion is no longer acceptable either from a theoretical or a practical point of view. It is well known that many soils were affected by erosion long before artificial changes began to be made.

Although the author is well aware of the need to respect established terms which are preferably not changed except in extreme cases, he is of the opinion that persistent adherence to old concepts of erosion does not hold with reality and causes considerable difficulties of understanding. It seems to be more appropriate to classify erosion in the broadest sense of the word into either *natural erosion* and erosion influenced by man, i.e. *altered,* or *anthropogenic erosion.*

If it is desirable to preserve the established term *normal erosion* to refer to that erosion which occurs under "normal" conditions and which leads to the formation of a "normal" soil profile, then *natural erosion* may be subdivided into *normal* and *abnormal erosion.* In the latter case, removal is so great that no "normal" profile can develop. Similarly, altered erosion may further be subdivided into *accelerated* and *decelerated* (or *inhibited*) *erosion;* in decelerated erosion the intensity of erosion may be reduced only by the application of inhibiting control measures. In accelerated erosion man's interference increases the rate of natural erosion, and consequently this erosion is always *excessive.* In any case, the generally accepted term accelerated erosion should be maintained.

From the standpoint of soil formation and soil conservation the intensity of erosion is important; it may differ in natural erosion and altered erosion situations. The critical limit, in the author's opinion, is the point at which the rate of soil loss caused by erosion is equal to, or smaller than the rate of soil formation. Any erosion that occurs below this limit does not endanger the existence of the soil, and it is therefore suggested that such a level of erosion be referred to as *benignant;* when the critical point is exceeded *malignant erosion* occurs. Between these levels of erosion lies a balanced or *compensated erosion* situation in which the soil cover neither increases nor decreases.

Thus according to the author's recommendations, soil erosion may be classified, with respect to damage caused by man's interference, as follows



1.4 Erosion of other materials

The Earth's entire surface is subject to erosion, including areas covered by non-flowing water, glaciers, and snow. No solid matter, not even that of cliffs and rocks, escapes erosion. As is the case with soil, the surface of denuded rocks may develop rills engraved by the action of water. Rocks may be washed by raindrops and ground down by wind-carried crystals. These phenomena are termed *rock erosion,* or *lithoerosion* (Greek *lithos* – rock).

The term erosion is also used in technical literature to refer to the destruction of metals or other materials by flowing water, steam, gases, and other liquid or solid matter; e.g. the erosion of material in thermal power plants is well known. Ratner and Zelenskiĭ (1966) point out for example that in thermal power plants, water exceeding a speed of 50 m s^{-1}, causes the erosion of metals to a depth of several millimetres in 2–3 thousand hours and this results in a rapid deterioration of the equipment. The erosion of turbines due to water cavitation is a well-known phenomenon. In this connection the term *cavitation erosion* (Latin *cavitas* – hollow, cavity) is used (Noskievich 1959, Lichtmann 1962).

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Chapter 2

CLASSIFICATION OF SOIL EROSION

2.1 Classification of erosive agents

Erosive agents include water, ice (glaciers), snow, air (wind), detritus, plants, animals, and man. Adjectives describing the various types of erosion are formed by adding the suffix -ic, -ial, -ian, -genic to the root of the word. Some terms are well established and therefore a new, uniform terminology may meet with resistance. Based on these principal agents erosion may be classified as follows.

Water erosion, i.e. *aquatic* (Latin *aqua* – water), or *hydric* (Greek *hydor* – water) erosion is considered sometimes to be synonymous with fluvial erosion which, however, has a narrower meaning referring strictly to river erosion.

Glacial erosion comprises those phenomena which were named *glarosis* by Glock (1928), this being one aspect of *glaciation.*

Snow or *nival erosion* (Latin *nivalis* – relating to snow) is a part of the geomorphic action of snow which Matthes (1900) termed nivation.

Wind or *aeolian erosion* (orig. *aeolic,* incorrectly aeolitic) is better expressed, for the sake of consistency with other types of erosion, as *aerial erosion* (Latin *aer* – air); the term *aeolian* is, however, internationally established. Again it is a part of *aeolization* which includes the complex influence of wind on the development of the Earth's surface.

Ground or *soligenic erosion* (Latin *solum* – soil) has hitherto not been considered as erosion.

Finally, there is the separate category of erosion caused by *animals (zoogenic erosion), plants (phytogenic erosion),* and *man (anthropogenic erosion).*

2.1.1 Water erosion

Water erosion encompasses the destruction of the Earth's surface by raindrops, and by fluvial, subterranean and non-fluvial water, the most frequent destruction being that caused by non-fluvial water, mainly sea water. From this point of view erosion may be subdivided into two groups, namely *marine* or *maritime erosion* (Fig. 6), and continental or *terrestrial erosion* (Latin *terra* – earth). Marine erosion is sometimes also referred to as *subaquatic* or *submarine erosion.*

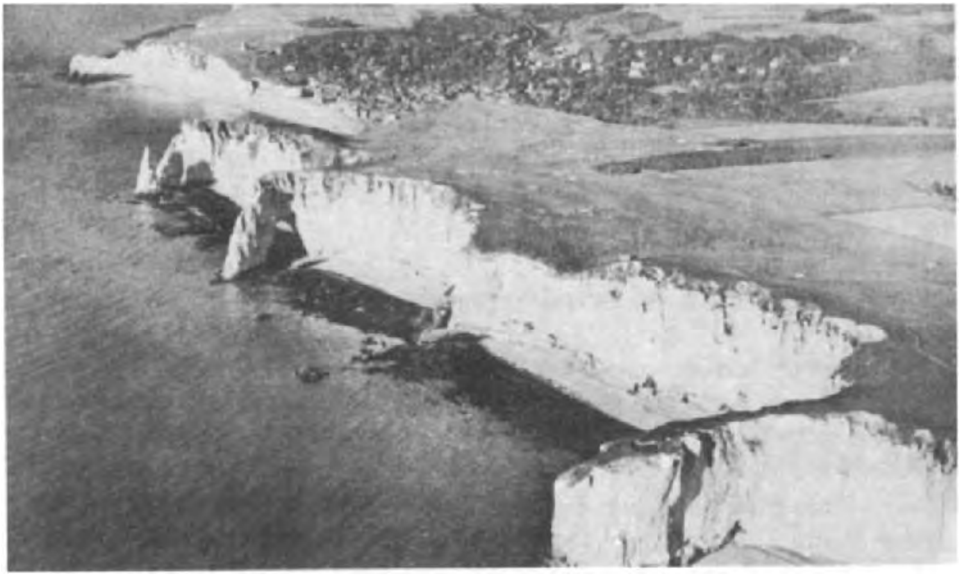


Fig. 6. Soil on the French coast damaged by marine erosion. (Aerial photo.)

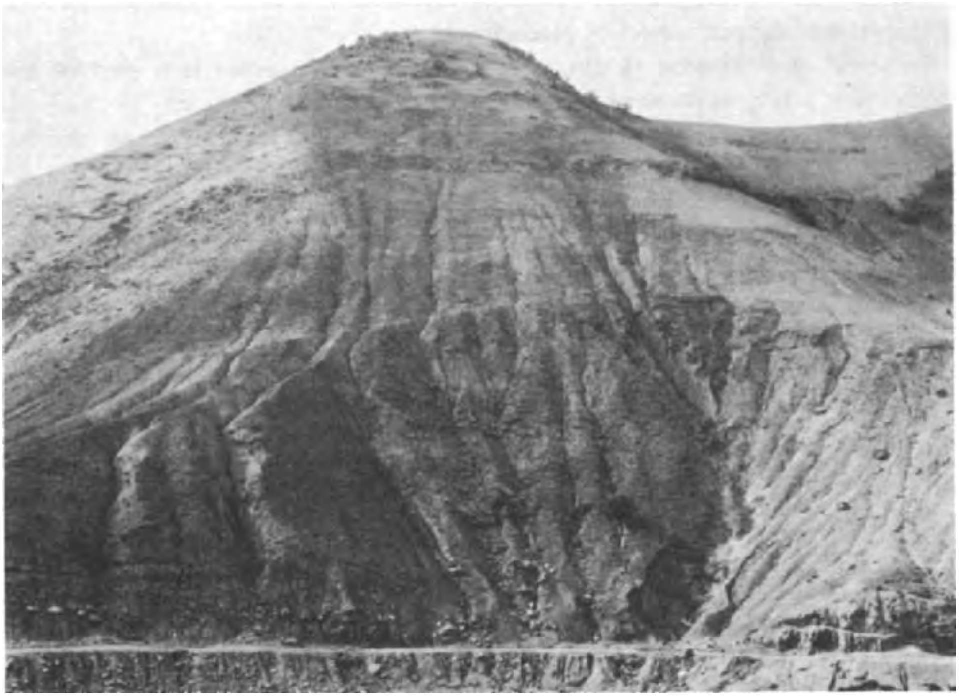


Fig. 7. Soil damaged by slope erosion under the arid conditions of Central Asia (Varzob river basin). (Photo D. Zachar.)

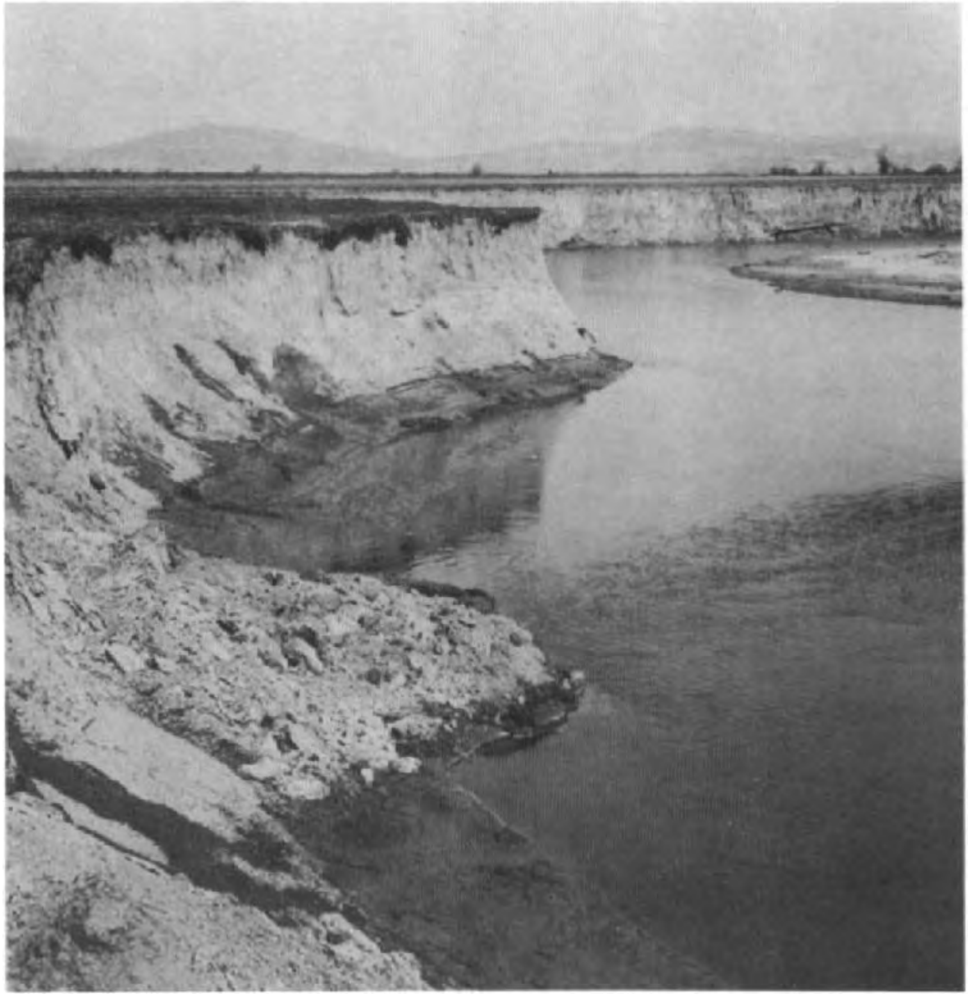


Fig. 8. Soil damaged by river erosion in the Ondava basin (Czechoslovakia). a – general view, b – detail. (Aerial photo.)

Both of these terms relate only to erosion of the sea bed, and consequently they have no bearing upon recently formed soil. Besides submarine erosion, erosion of the sea coast *(coastal* or *littoral erosion)* also occurs, however in this context the term marine or maritime abrasion has increasingly come into use in recent times. Coastal marine erosion is important as a factor involved in contemporary soil erosion.

Terrestrial erosion is also referred to by the older term *subaerial erosion.* It includes *precipitation* or *pluvial erosion* (Latin *pluvia* – precipitation), abbreviated: *plurosis, plurosion,* which is a part of pluviation and encompasses destruction

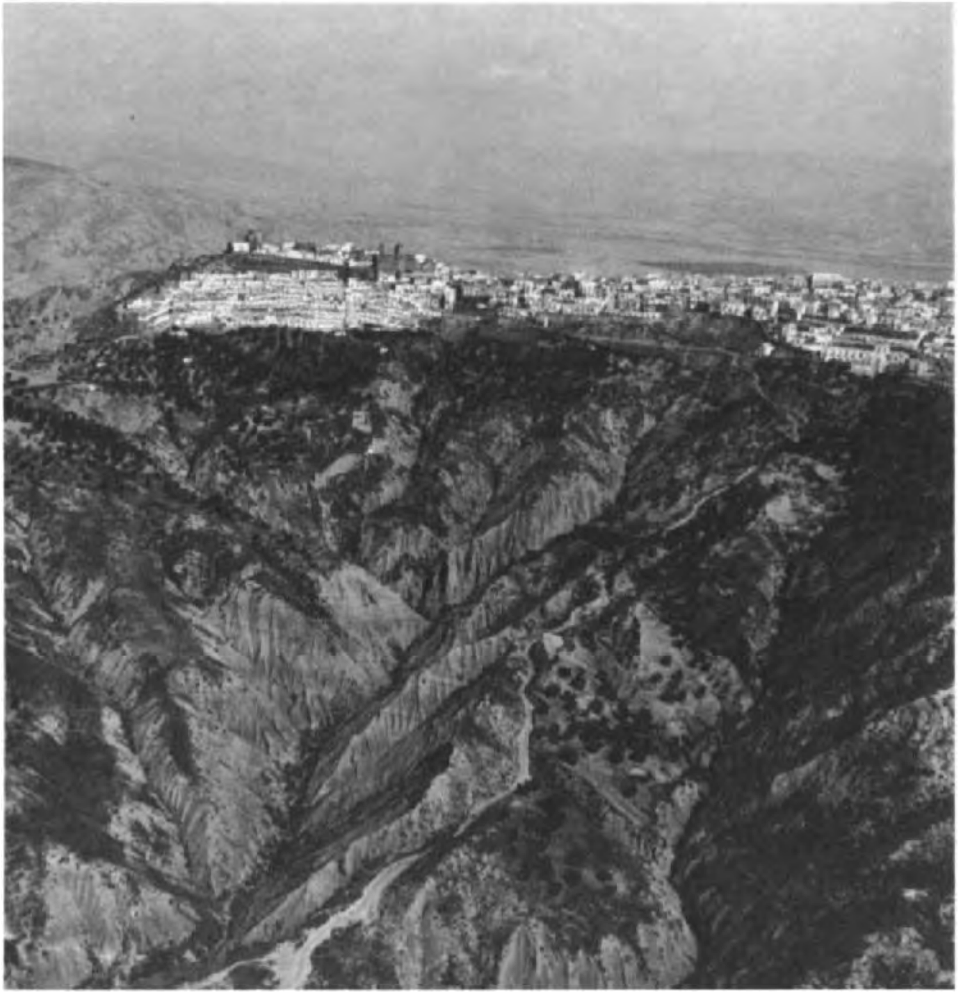


Fig. 9. Soil and entire terrain devastated by torrent erosion. The new part of the city was built after the old part had been destroyed in 1688 (Pistici in the Materna province of Italy). (By courtesy of Ende Riforma, Bari.)

by water in the form of rain, snow, and hail, respectively. For erosion caused by torrential rain (downpour) the author suggests the term *downpour* or *imbric erosion* (Latin *imber* – rain), and for *raindrop erosion* the term *guttation* (Latin *gutta* – drop), as the first stage of rainfall erosion. The term *eguttation* is used in speleology. In English the term *splash* is used to describe the displacement of soil by raindrops. The second stage of erosion caused by precipitation (snow and rain) is that resulting from surface runoff on slopes, and for this phenomenon the author recommends the term *slope* or *declival erosion* (Latin *declive* – slope) (Fig. 7).

Water erosion also refers to erosion caused by *river* or *fluvial erosion* (Fig. 8),

Fig. 10. Soil damaged by lacustrine erosion on the Orava Dam (Czechoslovakia). (Photo D. Zachar.)

contracted to *flurosion* (Latin *fluvius* – river) which is a part of *fluviation,* i.e. the shaping of a landscape by river water. This erosion is also termed *stream erosion* although this term is broader than river erosion since it includes dry river beds also. River erosion must also include *torrent erosion,* for which the author recommends the term torrential erosion (Latin *torrens* – torrent) (Fig. 9).

Lake erosion or *limnic erosion* (Greek *limne* – lake) is another form of water erosion. This international term is used mostly by hydrologists while geologists prefer the term *lacustrine erosion* (Latin *lacus* – lake) (Fig. 10). As in marine erosion, *coastal* (or tidal) *erosion* is recognized; the synonymous term *limnic abrasion* may also be used.

With respect to man-made constructions such as water reservoirs, the term dam erosion has been introduced; this kind of erosion is intensified by fluctuations of the water level.

A specific type of erosion strongly affected by man's interference is *irrigation erosion* which is included by some authors within the meaning of anthropogenic erosion, for the reason that the channels by which the water is conducted are artificial, and also that the water is driven into them by man-made devices. Irrigation erosion occurs mainly in arid regions in countries where there is widespread cultivation of rice or cotton (Japan, China, the USA, etc.).

The last type of water erosion is *channel erosion* (Latin *canalis* – channel) which acts essentially in the same way as does river erosion in natural beds. With respect to drainage systems in particular Henman (1966) speaks of *drain erosion.* This type of erosion also belongs to the category of anthropogenic erosion.

2.1.2 Glacial erosion

Glacial erosion is predominant in cold regions where the average temperature is below 0°C. A specific feature of glacial erosion is the action of a large mass of ice moving very slowly. Against this enormous force protective measures, including the encouragement of vegetation, have little effect. Another characteristic of glacial erosion is the fact that soil is damaged only at the edges of the ice, in new channels of ice and by melt water. The major part of the erosion energy is dissipated in the erosion of the bedrock.

The most pronounced form of glacial erosion is characterized by furrowing, cutting, ploughing, and scouring, and is given the scientific term *exaration* (Latin *aratio* – ploughing). Another form of glacial erosion is *griding* and *rubbing,* i.e. *glacial abrasion,* also referred to as *detersion* (Latin *detergere* – to cleanse) (Filip 1924, in Kettner 1954). In German the term for detersion is *schleifende Erosion,* and in Russian it is *shlifovanie.* The last type of glacial erosion to be considered is *severance,* i.e. *detraction* (Latin *detrahere* – to break off), which is a typical form of *disturbance* caused by glacial action. In German the equivalent term is *aushebende Erosion.*

Glacial erosion may be contracted to give the word *glarosion* which refers to those parts of the *glaciation process* affecting mainly soil genesis. Some authors restrict the term glaciation to the action of floating ice.

2.1.3 Snow erosion

The glacial zone is linked climatically with the zone of *snow,* or *nival erosion* which is pronounced in areas where there is a permanent snow cover (e.g. above the snow line). In contrast to glacial erosion, the different forms of snow erosion actively damage the soil, especially in avalanche channels where the great pressure and velocity of the snow cause erosion rills. It is recommended here that this type of snow erosion, as in glacial erosion, be referred to as *nival exaration.*

Fig. 11. Niveo-aeolian (nivaeolian) phenomena in aphytogenic terrain. Professor A. Cailleux on the picture (Poste de la Balcine – Great Whale River above Hudson Bay, Canada). (Photo A. Jahn.)

Soil is eroded also by the slow, creeping movement of snow, especially on leeward slopes where the soil is rubbed in a downhill direction. In addition to snow pressure, erosion is intensified by the movement of waterlogged soil and water runoff. In this case the term *subnival erosion* may be used.

Snow erosion is part of the process of *nivation* which, particularly in areas with a periglacial climate, enters into a variety of combinations with other modelling processes, especially *cryoplanation* and *gelivation* (the older term was *congelivation*), *glaciation, pluviation, fluviation,* and *aeolization.* In such cases one may speak of *nivaeolian* (Fig. 11), *nivoglacial, nivopluvial phenomena,* etc.

2.1.4 Wind erosion

Wind erosion is as important with respect to the soil as water erosion. Wind erosion occurs mainly in those areas where there is a lack of precipitation together with predominantly high temperatures, i.e. in arid regions. The decisive factor in wind erosion is vegetation, the importance of which for soil conservation increases with increasing aridness. Water erosion on the other hand, is a feature typical of

Fig. 12. Wind created dunes in the Sahara Desert (Great Eastern Ergs, Algeria). (Photo D. Zachar).

humid areas. Both types of erosion meet in semi-arid zones where they both occur in pronounced forms.

As with water erosion the terminology of wind erosion is etymologically ambiguous, even at the outset with regard to the basic word *aeolian*. Some authors consider this to be an unnecessary term synonymous with *deflation* (Kozmenko 1949, and others). The word deflation, derived from the Latin *deflare* – to blow away, is relevant to the removal of soil particles by wind erosion, and does not concern the soil which remains in situ. Consequently, deflation is not a substitute for aeolian erosion but rather represents a complement, a further stage of the process.

But wind also affects the soil, rocks, and minerals which remain in situ on the wind-blown surface. This material, which may also include the bedrock, is worn down by *wind-carried soil* particles or other *solid matter*. In this case *wind abrasion*, or alternatively *wind corrasion*, occurs. Recently it has been recommended that the term abrasion be used in the context of *littoral erosion* and that the term corrasion be used for wind abrasion.

Finally, erosion also includes the turbulent scouring of honeycomb hollows in rock walls which are composed of soft minerals. This phenomenon may be called *aeroxystosis*, derived from the established term *aeroxysts* – *rock honeycombs*.

Thermal changes, rain, insects, and other factors participate as accessories in the formation of aeroxysts.

Wind erosion constitutes a part of *aeolization,* the latter being the complex of all those geomorphological processes that result from wind action. If wind erosion is the prevalent form of erosion, pronounced *aeolian territories* come into being, with aeolian erosion, deflation and accumulation overlapping one another (Fig. 12).

2.1.5 Earth erosion

Erosion of this type has not as yet been treated as an independent erosion phenomenon although phenomena connected with earth erosion are well known. A pronounced form of erosive destruction of the soil and rock substrata is caused by so-called *debris flows,* mud flows (Russian *sel,* German *Mure*). The effect is produced by the flowing motion of waterlogged earth, soil, gravel, debris, rock, etc. which gouges rills out of the substrate during its avalanche-like movement. The term earth flow, Mure, sel, refers to the moving mass and not to the eroded substrata. In a similar way soil and its substrata are eroded by landslides (decerption), especially when a flow occurs. Since during earth flows the combined mass of water and debris is *moving* as an *avalanche,* the word avalanche instead of flow is used in some languages. Because the earth flow is always associated with a water torrent (whether this be caused by rain, snow, a glacier, or a lake), the author recommends the term *earth torrent.*

In the process of erosion caused by the moving detritus of earth flows it must be taken into account that the detritus is the main erosion factor, while water functions as a lubricant which diminishes the friction of the moving mass. Although there are references in the literature to rills caused by debris flows (e.g. Plesník 1966), their origin as part of the erosion process has not been described so far, in spite of the fact that the action of earth flows is now widely regarded as an erosion process.

Earth flows occur episodically and the resulting rills, like those caused by avalanches, are then further acted upon by precipitation erosion. On soft rocks, earth flows may also take on the characteristics of landslides. Thus the concept of earth erosion as an independent phenomenon presents difficulties.

It should be stressed that the terms *earth flow, earth avalanche,* etc., refer only to the sporadic movement of material, although the creation of rills by such movement has the appearance of an erosion process. In a similar way, snow avalanches and the avalanche rills and channels, etc. are distinguished as separate phenomena.

Because rills caused by earth flows are fashioned by a torrent of water and mass of earth, and because soil (or earth) erosion is caused by the mass itself, the author proposes that this type of erosion be denoted by the term *earth erosion,* or *soligenic* *erosion* (Latin *solum* – earth, soil). Where soil erosion is combined with the erosive effects of water, the term *aquasoligenic erosion* (Latin *aqua* – water) is proposed. The flowing motion of earth is also closely associated with soil flow, referred to in the literature as *solifluction* (Latin *solum* – earth, soil, *fluctus* – flow), and a special chapter devoted to this phenomenon has been included in this work.

2.1.6 Soil flow

Soil flow or *solifluction* refers to the flow of soil (earth) under the influence of gravity – a common occurrence in regions with a nival or sub-nival climate. In Czechoslovakia solifluction processes have played an important role in the modelling of the Earth's surface in the pleistocene periglacial climate.

The term *solifluction* was introduced by Andersson (1906) who described it as the motion of a thawed, slushy mass down an inclined frozen bottom. It is in this sense that the term solifluction is used in the literature of Czechoslovakia.

This type of solifluction therefore involves a perturbation of the soil structure by regelation and a decrease in internal soil friction as the result of a greater water

Fig. 13. Solifluction clearly being influenced by slope gradient and the protective effect of vegetation (Engadine National Park, Switzerland). (Photo J. Pelíšek.)

Fig. 14. Aquasolifluction and subsequent pluvial erosion in Gisborne (New Zealand) on clay-type slate covered by volcanic ash. At the time of European colonization, the territory was covered by dense hardwood forests which were burned down between 1905 and 1910. After destruction of the forests and the introduction of grazing animals, earth flows developed in association with intense erosion in the regions of destruction, and there was a massive choking of watercourses and valleys in regions of deposition. (By courtesy of New Zealand Forest Service – photo J. H. Johns.)

Fig. 15. Another example of solifluction combined with soil erosion, this time on the Nunatak in the Craigieburn Range (New Zealand). In this case also, the forest, remnants of which can be seen, was destroyed by fire in 1910. In the lower part of the photograph there is a valley with crops invaded by mud sediments. (By courtesy of New Zealand Forest Service – photo J. H. Johns.)

content in the upper soil strata. Soil with an impaired structure and high moisture content loses cohesion, and consequently, under the influence of gravity, begins to flow. Since in this process the soil is not put in motion by the exogenous influences of erosion factors, such solifluction cannot be classified as an erosion phenomenon, and should rather be regarded as a cryogenic occurrence (Fig. 13).

But the process is not quite so simple, because surface water cannot enter the soil when the lower layers are frozen and therefore of limited permeability. If the soil is not protected sufficiently by vegetation, or if the slope is steeply inclined, surface water (mostly snow water) brings about erosion of the less cohesive surface layers and, moreover, soil flow of cryogenic origin may also occur. This is illustrated in Figs. 14 and 15, in which both erosion and cryogenic phenomena are represented. The relationship between cryogenic soil flow and erosion may vary. Depending on

Fig. 16. Flowing of finely weathered, loose, non-cohesive material after complete destruction of the vegetation at an elevation of 6,000 to 7,000 feet. The vegetation was destroyed during the last 100 years (Craigieburn Range, New Zealand). (By courtesy of New Zealand Forest Service – photo J. H. Johns.)

which of the two phenomena prevails, the author recommends the use of the terms *cryofluvial erosion* and *pluviocryogenic erosion*. It is proposed that this phenomenon be classified as a special case of slope erosion.

It should be added that the content of the term *solifluction* is usually defined within narrow limits. Soil movement or soil flow may also occur in other circumstances, as described by Andersson (1906); soil flow may result not only from cryogenesis, but also from precipitation soaking the upper layers of the soil. The earth flows mentioned earlier are well-known. In some cases flowing landslides take on the character of soil flow, as Záruba has described in several reports. The main observations are given in a book published in 1969 (Záruba and Mencl 1969). Heim (1924) and other geologists have also recognized the so-called *subaquatic solifluction* which takes place without the influence of frost.

Finally, soil (or earth) flows may occur in material which is not very cohesive, even without the influence of frost and water, owing to a decrease in slope stability or an increase in external pressure. This phenomenon is exemplified by the *flow of sand* (German *Schwimmsand*) which can be frequently observed in addition to others, on the leeward side of sand dunes. An interesting example of a dry soil flow of finely weathered, but not very cohesive material, is shown in Fig. 16. Coarse detritus moving without the influence of frost, water, or erosion factors, may also have the characteristics of a flow. In the German literature, flow of this type is referred to as *dry earth flow (trockene Muregänge)*. Within this phenomenon are included flows of coarse detritus on pedoeroded slopes, and dry flows of lava material in volcanic eruptions (Penck 1894, Rüger 1929).

Thus soil flow or solifluction is a broader phenomenon than is generally understood. This fact has been correctly pointed out by Dylik (1951) who recommends the term *congelifluction* (derived from the Latin words *congelere* – to freeze, and *fluxus* – flow), instead of solifluction. The author suggests that the term solifluction should be maintained to refer to soil flow occurring under different climatic conditions. Kacherin (1958) has proposed the use of the expression *cryosolifluction,* instead of the term congelifluction. Kaplina (1965) also considers that cryogenic solifluction is more correctly described by cryosolifluction than by the traditionally used solifluction.

The author recommends that the expression *solifluction should include all phenomena in which soil flow occurs.* According to present knowledge solifluction can be further divided into: 1. *frost-induced soil flow* (i.e. *cryosolifluction*),* as one factor in the cryogenic modelling of slopes; 2. *water-induced soil flow* (i.e. *aquasolifluction*), as a comprehensive term referring to all phenomena included within the author's term earth torrent (Russian *sel, selevie potoki;* German *Mure, Muregang*); 3. dry soil flow (i.e. *siccesolifluction:* Latin *siccum* – dryness), as a comprehensive term for the flow of loose, non-cohesive detritus.

2.1.7 Organic erosion

Soil erosion caused by living organisms is fairly common despite being little known; only seldom is it regarded as an erosion phenomenon, yet it forms a part of the total destructive geological activity in the broadest sense. Kettner (1954) made the distinction between *phytogenic* and *zoogenic erosion.*

Phytogenic erosion includes soil destruction caused by roots *(root erosion),* as described under the term *Wurzelerosion* by Glander (1956). This activity is, of course, positive from the pedological point of view (and therefore also in the context of soil erosion), because weathering replaces soil losses in the soil mantle caused by outside forces and the removal (harvesting) of organic plant material (Lazarević 1975).

Zoogenic erosion is different situation in which animals destroy the soil when searching for food, moving (Fig. 17), or excavating their hiding places on the

* Mizerov (1966) uses the term *moroznaya solifluktsiya.*

Fig. 17. Erosion caused by cattle (zoogenic erosion). "Dental" formations are produced by cloven hoofs (High Tatra region, Czechoslovakia). (Photo D. Zachar.)

surface and under the ground. The activity of *pedobionts* and *geobionts* is well-known; the latter dig out considerable amounts of earth which are subsequently carried away by water or wind action. Animal and plant activity is particularly dangerous around dams where it accelerates ***suffosive erosion*** and often causes flood disasters. Like domestic animals, wild animals living entirely above-ground, may cause erosion indirectly by destroying vegetation and impairing soil properties. Bennett (1939) describes cases in which locusts and other pests have induced intensive erosion by destroying vegetation.

2.1.8 Anthropogenic erosion

Man influences erosion indirectly mainly by accelerating soil erosion and increasing the consequent devastation. Man's indirect action mostly involves destruction of natural vegetation, the cultivation of crops with a small soil-protecting effect, exposing the bare soil, increasing and concentrating surface runoff, and changing the quality of the soil (e.g. by decreasing humus content, impairing soil structure, reducing levels of nutrients in the soil, diminishing fertility, polluting the soil with industrial fumes and dust, etc.). The grazing of domestic animals is another means by which man indirectly influences erosion.

Since erosion concerns only the destructive activity of natural factors, and since man indirectly increases the effects of these factors, anthropogenic erosion cannot be regarded as an independent type of erosion; to distinguish between the different causes of erosion, the latter should be named according to the basic natural factors. Because in many cases anthropogenic erosion has a specific form, it is possible to speak in terms of *agricultural erosion, silvicultural erosion,* as well as *grazing erosion* [Dzhunushbaev (1962) uses the term *pastbishtnaya éroziya*], *road erosion* [Armand (1956) *dorozhnaya éroziya;* Schultze (1952) *Wegerosion*], *logging* or *exploitation erosion,* etc. (Fig. 18).

A frequent cause of the acceleration of soil erosion, besides overgrazing, is fire, which diminishes the protective effect of vegetation and leads to a rapid degradation of fire-devastated sites; Fig. 19 shows an example of this in an Alpine region. In arid regions the pedoerosion process is connected with salinification (Fig. 20).

Fig. 18. Road erosion caused by the gouging of tracks and wash (surroundings of Banská Bystrica, Czechoslovakia). (Photo D. Zachar.)

Fig. 19. Overgrazing and fire bring about active erosion and the degradation of pastures. (By courtesy of Soil Conservation Service of N.S.W., Australia.)

Fig. 20. In arid regions erosion and salination, as well as a general decline in soil fertility are caused by overgrazing and the removal of vegetation. (By courtesy of Soil Conservation Authority, A.S.)

2.1.9 Conclusion

This is the complete classification of erosion by causative factor. In nature, however, different combinations of factors may occur and the various types of erosion rarely take place in isolation. Various phenomena frequently mix, alternate, or link up. Cholley (1950) speaks of erosion systems which under different conditions create outwardly diverse composite forms of soil destruction and surface modelling, although the systems themselves comprise uniform sets of factors.

Besides combinations of different erosion phenomena, combinations of erosion phenomena together with other phenomena of slope modelling may occur, for example, *gravitation erosion, decerption erosion, solifluction erosion,* and other phenomena. Żiemnicki and Józefaciuk (1965) mention *erosion land-slides,* Bury-Zaleska and Piotrowski (1964) speak of *erosion suffosive phenomena,* Kaplina (1965) describes *cryogenic erosion,* Plesník (1966) *earth flow gullies,* Yakutilov (1962) *erosion earthflow processes,* etc.

A survey of the main types of erosion classified by origin of erosion factor is given in Table 1. From the pedogenetic point of view precipitation and wind erosion are the most important ones. These types of erosion affect large territories and are important in terms of land economics. Other types of erosion are confined to smaller land areas.

Table 1. Classification of erosion by the active factor

Factor	Term	
	English	International
1 water	Water erosion	Aquatic erosion
1.1 precipitation, rain	Precipitation erosion,rain e.	Pluvial erosion
1.2 river	River erosion	Fluvial erosion
torrent	Torrent erosion	Torrential erosion
1.3 lake, reservoir	Lake erosion, reservoir e.	Limnic erosion, lacustrine erosion
1.4 sea	Sea erosion	Marine erosion
2 glacier	Glacier erosion	Glacial erosion
3 snow	Snow erosion	Nival erosion
4 wind	Wind erosion	Aeolian erosion
5 earth, debris	Earth erosion	Soligenic erosion
6 organisms	Biological erosion	Organogenic erosion
6.1 plants	Erosion caused by plants	Phytogenic erosion
6.2 animals	Erosion caused by animals	Zoogenic erosion
6.3 man	Erosion caused by man	Anthropogenic erosion

2.2 Classification of erosion by form

2.2.1 General

As a result of the action of exogenous factors in soil erosion, certain forms arise in the soil and on the earth surface which influence not only the development of the soil cover, but also the morphogenesis of the land. The classification of erosion phenomena by form meets with several obstacles which have so far made it impossible to create a uniform classification system. One of the main obstacles is the fact that no universal definition of soil erosion has hitherto been accepted; another defect is the absence of uniform criteria by which erosion forms may be assessed. Finally, consideration must be given to the fact that erosion is, in time and space, a very complicated and diverse phenomenon representing only one form of land modelling among many – a fact which complicates the situation, especially with respect to secular phenomena. Notwithstanding these obstacles, soil erosion can be classified by form. Such a classification is necessary because the form may give clues to important features of the erosion process. From the form of the erosion, the origin, intensity, development, and means of erosion control, etc. may be judged.

One aspect which needs to be clarified in the general assessment of form in erosion phenomena is the *scale of the erosion.* As has been shown, erosion includes phenomena ranging from soil ablation in which the soil is disturbed on an imperceptible scale, up to the formation of erosion rills and other forms. According to the scale on which erosion takes place, the author recommends the following classification: 1. *microerosion,* 2. *mesoerosion,* and 3. *macroerosion* (Greek *micros* – small, *mesos* – middle, *macros* – big). Under the influence of these a microrelief, mesorelief, or macrorelief develops, and therefore the dimensions of an erosion form should also correspond with the dimensions of the geomorphological relief.

With regard to present concepts of microrelief, mesorelief, and macrorelief microerosion may be identified with sheet erosion and small-scale rill erosion, mesoerosion with rill erosion and certain forms of lake or river erosion, and macroerosion with most of river and sea erosion in which recent forms give way to older, larger forms.

The question of *place* must also be considered. In most erosion forms it is assumed that the destructive activity of factors such as water and wind takes place on the surface of the pedosphere or lithosphere. Yet natural phenomena are usually complicated and erosion should also take into account those phenomena which are caused by exogenous factors (especially water and geobionts) beneath the soil surface – in the soil or in the substratum. These phenomena are well-known in the literature but are classified only reluctantly as erosion processes, although their erosion character is undisputed.

Since the activity of subterranean erosion factors results in specific forms distinct from surface forms, the author recommends the further classification of erosion into *surface erosion* and *underground erosion.* In this way the terms sheet erosion and surface erosion would not be confused because sheet erosion, as conceived by the author, is only one form of surface erosion, and is thus more restricted in meaning.

As the scientific equivalent of surface erosion, the author proposes the use of the term *exomorphic erosion* (Greek *exo-* – external, *morfé* – form) to describe erosion caused by external phenomena. Alternatively, *superficial erosion* (derived from the Latin *superficialis*) would also be a suitable term. For underground erosion the author suggests the term *cryptomorphic erosion* (Greek *kryptos* – hidden), or *subficial erosion.* The terms *exoerosion* and *cryptoerosion* could be used as abbreviations, and the equivalents in other languages would be: *poverkhnostnaya, podzemnaya éroziya, érosion superficielle, crypto-érosion, oberirdische, unterirdische Erosion.*

In the following sections attention will be given mainly to those erosion phenomena which are closely related to soil erosion. Forms of precipitation erosion are discussed first.

2.2.2 Forms of precipitation erosion

Precipitation erosion represents, together with wind erosion, the main area of interest in the discipline of soil erosion, and the greater part of all works published in the last 40 years in the world literature are concerned only with these two types of erosion. Under the prevailing conditions in Czechoslovakia precipitation erosion is more important than wind erosion and therefore more space is devoted to the former in this work.

Before starting an analysis of the various forms of precipitation erosion it should be noted that the precipitation erosion of soil is divided into two main groups according to the parts of the pedosphere and lithosphere, respectively, in which it takes place. The first group includes *surface phenomena* (i.e. exomorphic, or superficial phenomena), and the second group comprises *underground phenomena* (i.e. cryptomorphic, or subficial phenomena).

2.2.2.1 Surface erosion

Surface erosion caused by precipitation implies the destruction of soil by *raindrops* (hail) and *surface runoff* in the form of precipitation and snow water flowing down a sloping surface. In the literature, different systems are used for the classification of precipitation erosion, the more important ones being mentioned below.

In the first erodological monograph of Ayres (1936), *water erosion* is divided into 1. *sheet washing,* 2. *gullying,* and 3. *stream erosion,* or *river erosion.* Bennett (1939), discussing erosion caused by water, distinguishes: 1. *sheet erosion,* 2. *rill erosion, gully erosion* and *rock erosion.* Rock erosion is defined as erosion taking place in rocky areas in which *rock gorges* and *badland* develop. In the third American work, published by Kohnke and Bertrand (1959), the following classification is presented: 1. *sheet erosion,* 2. *internal erosion,* and 3. *channel erosion,* which is further divited into a) *rill erosion,* b) *gully erosion,* and c) *stream erosion.* Most American authors also include within the concept of erosion certain types of landslides and earth flows, etc.

In the Russian literature Kozmenko (1954) classifies currently continuing erosion as: 1. *washing* (smyv), and 2. *wearing* (razmyv). Sobolev (1948) divides water erosion into: 1. *sheet erosion* (ploskostnaya éroziya, smyv pochvy), and 2. *gully erosion* (ovrazhnaya éroziya). In sheet erosion the subtype 1a *rill erosion* (struĭchataya éroziya) is distinguished. Sus (1949), like some American authors, distinguishes between *normal* (normalnaya) *erosion* and *accelerated* (uskorennaya eroziya) *erosion,* the latter being subdivided into: 1. *surface* (poverkhnostnaya, ploskostnaya) *erosion,* 2. *rill erosion* (struĭchataya), and 3. linear, or *gully erosion* (lineĭnaya, ili ovrazhnaya eroziya).

For the sake of brevity, only the classification of Fournier (1956) will be cited from the French literature; this differs somewhat from the previously mentioned ones. Water erosion *(l'érosion hydraulique)* is divided into: A) washing of soil particles *(le détachement et l' entraînement des particles constitutives du sol),* and B) mass removal of soil *(le déplacement du sol en masse).* In both cases water may act on the soil surface or within the soil. Soil washing is subdivided into: 1. downpour erosion *(l'érosion par battage du sol),* 2. sheet erosion *(l'érosion en nappe),* 3. layer, or strip erosion *(l'érosion en nappes ravinantes),* 4. rill erosion *(l'érosion en rigoles),* and 5. wearing erosion *(l'érosion en ravins).* In forms 1 to 3, sculpturing by the erosion process is obliterated, while in forms 4 and 5 it is visible from the start. Mass removal of soil which, according to the author, has the character of erosion includes: 1. landslides *(éboulements),* 2. erosion by underground channels *(l'érosion par chenaux souterrains),* 3. mud flows *(coulées boueuses),* and 4. creeping of the soil *(reptation du sol).*

It would be possible to continue citing examples of classification systems, with very few identical systems emerging from the survey. Yet despite the variety of approaches to the classification of precipitation erosion by forms, sheet erosion is regarded in every system as a specific form, and only in a few cases is raindrop erosion placed first, although drop erosion also affects the surface. Another common feature in erosion classifications is the distinction given in all systems to gully erosion which involves a relatively small part of the ground surface and is very different from sheet erosion. Further classification of gullies as to size and general character differs widely, the main differences occurring in the assignment of small rills – rillets; some authors include these as an aspect of sheet erosion, others consider them to be a separate form, or include them in gully erosion. Other forms are seldom distinguished.

In this work *exomorphic* (surface) *precipitation erosion* is divided into: 1. *sheet erosion,* 2. *gully erosion,* 3. *multiform erosion,* and 4. *rock erosion.*

Sheet erosion

The main feature of sheet erosion is the more or less uniform erosion of the soil over the whole surface of the land of over a particular part of a slope. Erosion is caused by raindrops and surface runoff. As evenness of the surface of the slope increases, opportunities for the accumulation of water decrease and sheet erosion becomes more uniform. Accumulation of water may nevertheless occur even on the smoothest slope. It is therefore difficult to separate sheet erosion in which there are almost imperceptible outward signs of erosion from rill erosion, which in the final result, if the rills do not deepen, is a surface phenomenon. The intensity of accumulation of runoff water depends on the height of the water stream, the coarseness of the surface, and other factors.

The action of sheet erosion causes the soil mantle to become thinner and finally the *underlying rock* and *mineral substrata* is laid bare over a large area. Erosion relief occurs most often on agricultural land, where small unevennesses caused by rillets become effaced during the cultivation of the soil, so that the washing of small plates and rillets cannot be distinguished after a time.

Sheet erosion involves the removal of: 1. *particles loosened* by weathering (raindrops, frost, mechanical action of machines and animals, etc.), and 2. *easily dissoluble matter,* matter made soluble by weak acids in rain water.

Thus sheet erosion represents *microerosion* in the true sense of the word, i.e. the eroding and washing of the soil to produce small scale forms which may encompass raindrop erosion, laminar erosion, rillet erosion, and layer erosion.

The first phase of sheet erosion, specific with regard to form, is soil removal by raindrop action – *raindrop erosion.* In this way soil may be markedly disaggregated and eroded, but the displacement of soil particles remains relatively small (as long as no surface runoff occurs) though displacement being permanent. This is an important erosion factor on ridges, in furrows, and in erosion remnants, etc. In raindrop erosion the surface is acted upon selectively so that small holes, micropyramids and other forms occur, *raindrop erosion* thus becoming a part of *pedestal erosion, pinnacle erosion,* etc.

The second subtype in sheet erosion is *laminar erosion* (Latin *lamina* – thin layer). It occurs in any flow of water on an inclined soil surface where the kinetic energy of the water is small and only the finest soil particles are consequently washed away in a strongly selective manner. This erosion is sometimes called

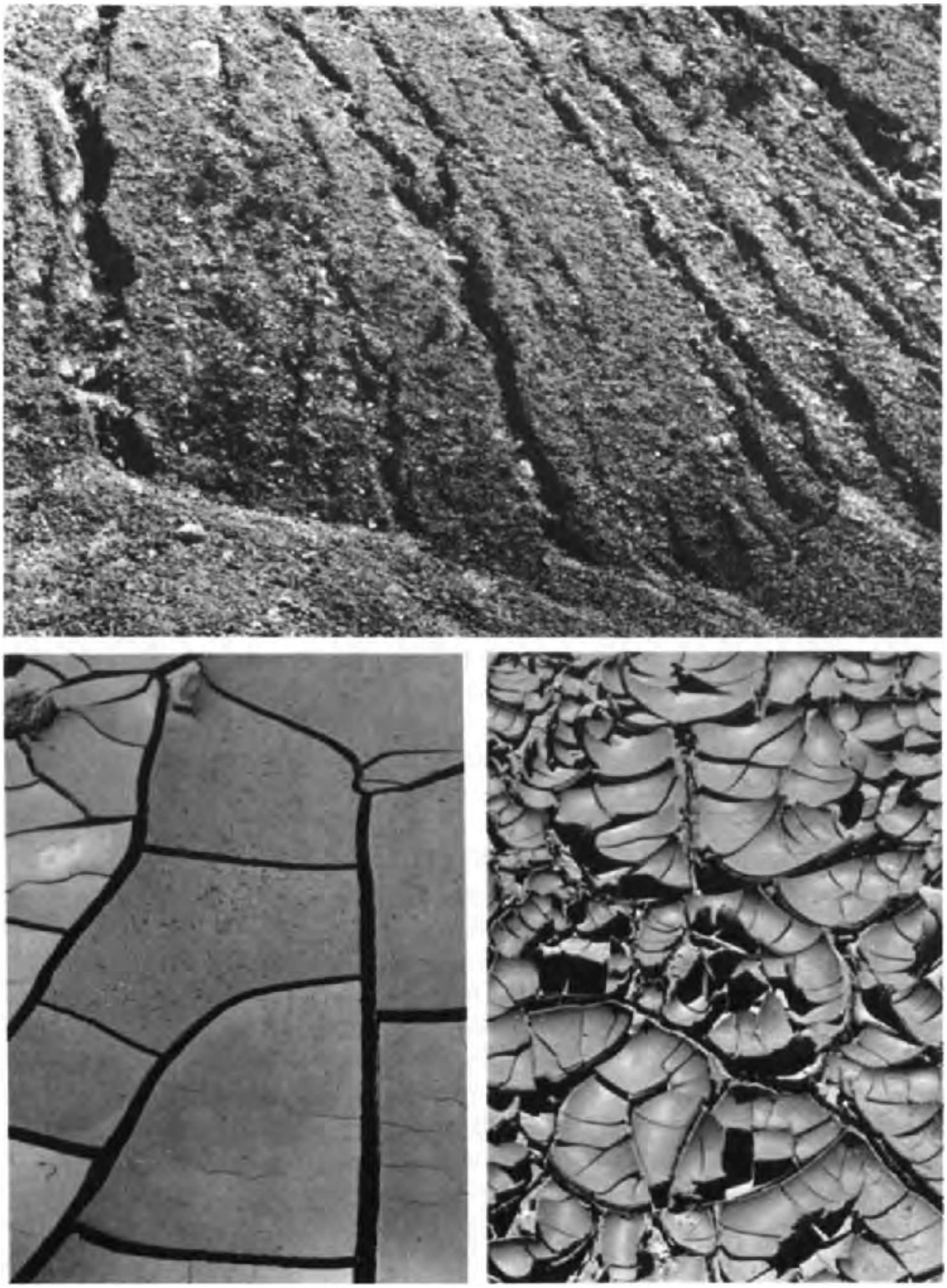


Fig. 21. Soil affected by rill erosion (skeletal content 43%). a – soil surface covered by coarse fractions at the foot of a slope, skeletal content up to 87%, b – loamy deposit with 5% skeletal content, c – fine deposit with 2% skeletal content and some scaling off in shells. (Photo D. Zachar.)

Fig. 22. Development of rill erosion into laminar erosion. Here and there the whole topsoil has been carried away by a downpour. (Photo D. Zachar.)

selective erosion but it should be noted that any type of erosion acts in a more or less selective way, and therefore this denomination is not appropriate in this instance. In arid regions where a layer of salt develops on the soil surface, so-called *pellicular erosion* (Latin *pellicula* – thin skin or film) occurs; this is mainly of a mechanico-chemical nature.

By virtue of the accumulation of sheet runoff water, *rill erosion* develops causing small rills with the dimensions of a few centimetres diameter in cross-section, and with a depth not exceeding that of the arable layer. The rillets sometimes develop in rows and furrows, etc., with the effect of increasing their dimensions and conspicuousness, but the traces of this erosion are removed during harvesting and cultivation. In this form of erosion, soil and particles displaced by water may be intensively separated and sorted (Fig. 21).

Layer erosion refers to a distinct type which the author has observed on several occasions on tilled land. In layer erosion the soil is washed away neither in laminae, nor in rillets or rills, but in a layer up to several metres wide and 10 to 25 cm deep, i.e. in apparent strips from which the topsoil has been entirely removed (Fig. 22).

Fig. 23. Laminar erosion in the transition belt between desert soil and irrigated land in Mesopotamia. Even on minimal gradients erosion occurs during heavy downpours (according to Buringh 1960).

A similar phenomenon was described by Fournier (1956). Layer erosion also occurs frequently in arid regions where the top layer is eroded first (Fig. 23).

The author recommends that all forms of sheet erosion be referred to as *area erosion,* to distinguish them from *linear erosion,* or *gully erosion.* It is proposed that the term *deluviation* (Latin *deluere* – to wash off) should be used to refer to the very important form of soil damage that is caused by area erosion. This term adequately covers those forms of soil wash by which the well-known deluvia originate. Against deluviation the vertical transfer of particles into the soil and substratum known as *eluviation* (Latin *eluere* – to wash out) is distinguished. Both processes are important constituents of soil pluviation.

Gully erosion

By the accumulation of larger quantities of water or by the gradual deepening of rills, erosion gullies of various size and form come into being (Fig. 24). The term gully is preferred because it has the broadest meaning. A number of forms may be distinguished in gully erosion.

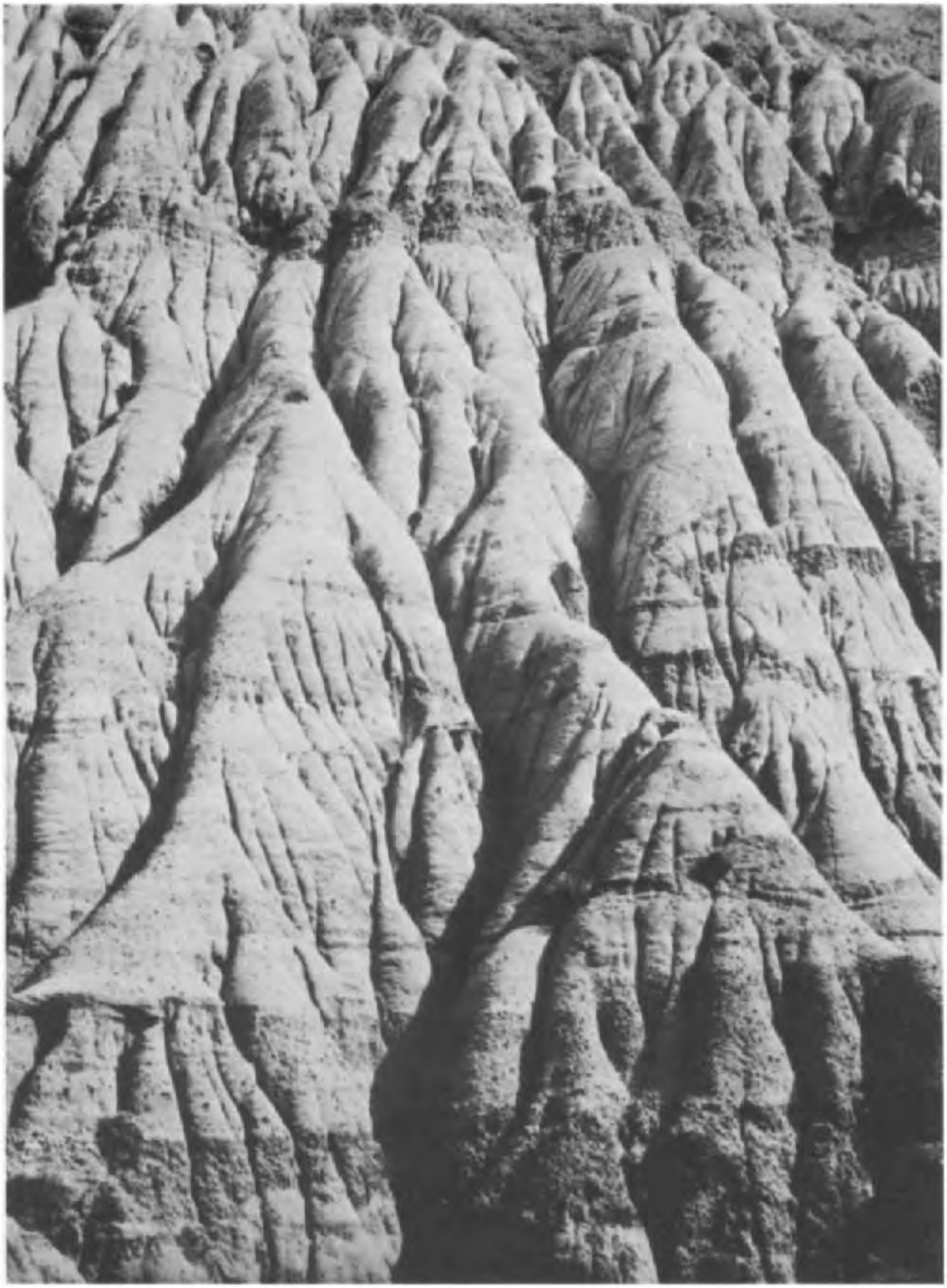


Fig. 24. Growth of small rills into deeper V-shaped gullies (Red Mountains, Romania). (Photo D. Zachar.)

Fig. 25. Retrograde gully erosion in Ordovician sediment, the steep slopes showing a hardsetting surface over a sodium enriched subsoil (central Victoria). (By courtesy of Soil Conservation Authority of Victoria, Australia.)

The first form includes any erosion gully with a depth of between 30 cm and 2–3 m. In this form, typical wash prevails with a marked *backward* or *retrograde* (Latin *retrogradare* – to move back) *erosion* and *vertical* (Latin *verticalis* – perpendicular) or *depth erosion,* the erosion curve being compensated by *waterfall erosion* (Fig. 25). Gullies have larger dimensions and their development is more complicated. Besides retrograde and vertical erosion, *lateral erosion* also appears here, together with accessory landslide, soil flow, and other phenomena (Fig. 26). Gullies may grow into *gorges* and *canyons* which usually belong, of course, to the hydrographic network and are modelled by river erosion.

According to the forms of erosion gullies when viewed in cross-section, *flat, narrow, broad* and *round gullies* are distinguished. Flat forms occur mostly on shallow soil, or in connection with a specific lithic structure of the slope. In this form, characterized by a broad V-section, lateral erosion prevails over vertical erosion. Narrow, acute forms are created with a narrow V-section, the breadth of the gully usually being equal to its depth, or smaller. These gullies are also referred to in the literature as *microcanyons* (Kayser 1961). *Broad gullies* have a wide bottom and

Fig. 26. Top of a gully formed in deep loess sediments (basin of Yellow River, China). (The author's collection of photographs.)

are U-shaped. Here lateral erosion prevails over depth erosion. Active gullies maintain steep or even perpendicular sides (Fig. 27).

From among permanent enduring forms, *dales, dells,* and *blind creeks* may be included as types of gully erosion. In Russian these forms are described by Kozmenko (1954) and other authors as *lozhbina, loshchina, balka, sukhie doliny* (sukhdoly); the corresponding German terms are *Dellen, Tilken,* etc. These differ from recent forms in that they have become stabilized and the bottoms have been built up by aggradation. Man's interference may bring about renewed erosion in these rills, this being superimposed upon whole systems of past and more recent forms. In the steppe and forest steppe regions of the USSR, so-called *ovrachnobalochnye sistemy* are well-known, and similar forms occur in other countries, too (Fig. 28). It happens frequently that recent forms replace older forms so that their origin and age cannot be assessed from superficial observation.

The main feature of rill erosion is the concentration of washing by water at the lowest level, the *erosion line.* This results in the development (on a slope, in a valley, or some depression generally) of a notch or rill in which typical washing,

Fig. 27. Erosion gullies: a – broad V form, b – broad U form. The gullies had developed in Permocarbon sandstones. Today the area is planted with trees (central Bohemia, Czechoslovakia). (The author's collection of photographs.)

Fig. 28. Bank and bottom gullies in an old ravine (Kanev dislocations, USSR). The author's collection of photographs.)

or erosion of soil by the mechanical force of water occurs. Another characteristic of gully erosion is the *fragmentation of the slope* – the opposite phenomenon with respect to sheet erosion in which there is a rounding off, lowering, or *degrading of the slope.*

Synonyms of sheet and gully erosion

Several synonyms may be listed for various forms, but they do not always represent identical concepts because they are used in different contexts and are frequently only of local significance.

The following expressions may be regarded, with a certain degree of etymological tolerance as being synonyms for *sheet erosion:* Russian – *smyv, smyvanie pochv, peremyvanie pochv, ploskostnoĭ smyv, ploskostnaya éroziya, poverkhnostnaya éroziya, ploshchadnoĭ smyv, ploshchadnaya éroziya, plastovaya éroziya, sloistnaya éroziya;* French – *ruisselement, érosion tangentĭelle, la dégradation superficielle;* German – *Flächenerosion, Flächenspülung, Flächenabtrag, Abtragung, Denudation,* etc. The following English expressions may be derived from foreign terms: *denudation* (in the broader sense), *tangential denudation,* and as proposed by the author, *areal erosion* and *deluviation* (which also includes rill erosion).

Fig. 29. Furrow erosion caused by heavy rain in a vineyard with the rows arranged incorrectly up and down the slope (Little Carpathians, Czechoslovakia). (Photo D. Zachar.)

Synonyms for rillet erosion are: English – *microchannel erosion, rill washing;* Russian – *struĭchataya éroziya, rucheĭkovaya éroziya, vodoroĭnaya éroziya, borozdkovaya éroziya;* French – *rigoles;* German *Rillerosion, Rillspülung, Rillabtrag, Rinnenerosion, Rinnenspülung, Rinnenabtrag.*

Still more difficult is the listing of synonyms for *gully erosion,* which the author uses as a higher order term equivalent to *linear erosion.* This expression is also frequently used in other languages, and is entirely consistent with the creation of an erosion form in the direction of the slope line or depression line.

Besides gully erosion, the superior term *furrow erosion* is sometimes used. The author suggests that the latter should be reserved for phenomena connected with the washing of furrows created by ploughing or other operations; the term is a literal translation of the German *Furchenerosion.* In the same sense as furrow erosion, it is possible to speak of *track erosion,* row erosion, road erosion, sunken *road erosion, ditch erosion, channel erosion,* or other forms of erosion caused by water in man-made gullies (Fig. 29).

The author does not recommend the use of gully erosion as a collective term for all rills. The English equivalents are *linear erosion* and *channel erosion;* in Russian *lineĭnaya eroziya,* and sometimes also *ovrazhnaya eroziya* and *razmyv* are used; synonymous French terms are *ravinement* and *incision,* and in German *Einschnitt-*

erosion, Furchenerosion, Furchenspülung, Furchenabtrag, Rissbildung, Erosionsrisse, Rinnsale, etc.

Forms created by gully erosion are referred to in English as *gully, wash;* in Russian the currently used term is *ovragi.* Many authors refer to these forms by the terms *vodoroïnaya, promoïna, ovrazhek, vrazhek, yar, baïrek,* etc. In German the most frequently used terms are *Grabenerosion, Grabenspülung, Schlottererosion, Schlotterabtrag.* The formation of large gullies is denoted by the terms *coules, barrancas;* in Russia *ovragoobrazovanie;* in German *Schluchtenerosion, zerschluchtende Erosion, Kerbschluchten, Runse;* in Italian *burone, fosco;* in French *ravin (ravins en V, ravins en U), ravins morts.* Other known terms are *wadis* (in North Africa), *nullach* (in India), and *donga* (in South Africa) (Hudson 1971).

The gully form of erosion caused by precipitation water is not the final form that is observed. The term gully erosion applies only to those cases in which the gullies are distinct and have a catchment area and a certain part of the *interlinear (interpluvial,* or *interfluvial)* space which represents the surface of the original slope. Gradually the gully ridges become *interconnected (anastomosis), piracy* occurs, and the *gullies merge* together causing *total destruction of both the soil and the slope.* This kind of forms arises from the most highly intensive erosion and the author recommends the use of the term *polymorphic erosion* in these cases (Greek and Latin *poly-* – many, Greek *morfé* – shape).

Polymorphic erosion

Polymorphic erosion refers to various forms of soil destruction which are the main modelling factor of so-called badlands, and therefore it would be possible to speak also of *badland erosion.*

The phrase *badland erosion* derives from the French *mauvaises terres.* The French used this expression to describe unmanageable terrain furrowed by erosion in the prairies of Dakota and Nebraska to the South of the Black Hills. This term also has equivalents in other languages, but always conveying a different sense, so that it has finally been deemed preferable to use the expression badland internationally. The German word *Ödland* chosen to replance the term badland has a broader meaning; within the concept of *Ödland type,* Weck (1952) includes heaths and salinas, etc. In Russian the term "oedlend" has been adopted, although it is sometimes replaced by the term *durnye zemli* (Lilienberg 1955).

In the geological literature badland refers to terrain composed of soft, easily erodible rocks in semiarid regions which, during torrential rains, are washed by deep rills and gorges separated by sharp ridges. The ridges are further eroded by a network of new, rapidly changing rills which quickly eat into the rock so that any soil is entirely destroyed and the slope becomes densely furrowed by a whole system of gorges, gullies, earth coulisses, ridges, and other forms (Fig. 30).

Fig. 30. Badland in Death Valley (Golden Canyon) with a relatively simple geosculpture (California, USA). (Photo V. Čermák.)

The dense interweaving of the terrain by gullies, coulisses, and other forms is variously described in the literature as *ravining, gullying, ovragoobrazovanie* (in Russian), *Zerschluchtung, Racheln* (in German), *ravinement* (in French) – processes which are part of badland erosion.

Badland types of erosion are perhaps most conspicuously developed in the loess regions of China where extremely aggressive and intensive erosion occurs in deep layers of easily erodible material, sculpturing the slopes into bizarre forms resembling cave draperies (Fig. 31). Similar forms also occur in North Africa, although on a smaller scale. In many instances the wind polishes the slope ridges while the central and lower parts are eroded by water. A fairly extensive literature has been

Fig. 31. Bizarre polymorphous erosion forms in China's loess region (basin of Yellow River). (The author's collection of photographs.)

published on these phenomena (Chuan Bin-Bej 1954, Messines 1958, and other authors).

In Europe erosion of the badland type has been described in detail in Italy where it is known as *calanco*. Its forms have been extensively studied by Kayser (1961), and methods of control have been discussed by Puglisi (1963).

Fig. 32. The most aggressive stage of calancos in the Pantone Largo valley in the Materna province of Italy. (By courtesy of Ente Riforma, Bari.)

According to Kayser (1961) calancos are steep slopes, sharply furrowed by narrow gullies separated by earth coulisses the ridges of which join together towards the watershed dovetail fashion forming a skeleton *(ossatura)* of erosion remnants. Calancos developing on slopes not protected by vegetation are permanently attacked by water which eats into the rock and rapidly erodes the mountain massif. Kayser refers to these calancos as ***denuding calancos (les calanchi denudes),*** or ***bare calancos (calanchi nus),*** and classifies them as the most progressive of calancos (Fig. 32).

Besides this most typical calanco which resembles the badland erosion forms described by American specialists, there are two further types. One of these is a less active form occurring on slopes partially covered by shrubs (remnants of forest stands); Kayser calls these ***shrub calancos (calanchi à maquis).*** The last

Fig. 33. Badland erosion to the south of the Paricutin volcano (Michvacán, Mexico). (By courtesy of U. S. Geological Survey.)

calanco form consists of the erosion remnants of massifs which Kayser calls *elephant backs (calanchi d'élephants)*. They represent the last erosion stage of so-called *calanco erosion*.

The study of calancos in Lucania (southern Italy) has shown that their occurrence is limited to very steep slopes (inclination ranging from 28 to 45°) where the soil has a high clay content and is mostly of pliocene age; precipitation is irregular with occasional torrential rain. The steep inclination of the slopes is caused by undermining and intensive depth erosion occurring in gullies.

As a matter of interest it may be added that Kayser includes within the scope of the sheet and linear erosion of this region, not only calancos, but also the so-called *frane* (Italian), which includes both *mud streams (coulées boueuses)* and *rock walls (les parois rocheuses)*. The first of these occur in the flysch zone as permanent erosion forms which become active at intervals of 5 to 10 years; they can be up to several kilometres long. In the intervening periods between flowing of the mass, the separation and flow regions are modelled by surface water. This latter erosion form occurs on steep hillsides where older geological strata are exposed (limestone and sandstone cliffs), or is caused by *accelerated erosion (érosion accelerée)* in unstable pliocene sandstones and conglomerates.

A particular type of precipitation erosion which is treated separately by some authors is rock erosion. This term was first used by Bennett (1939) who used it to describe erosion forms in the Rocky Mountains. In his opinion, rocks erosion occurs (in this case at least) without human interference. Its forms are diverse, and their analysis is beyond the scope of this work.

A special case of badland formation occurs in regions affected by "volcanic" rain, the origin of which is connected with violent turbothermal currents that develop during volcanic explosions (Fig. 33).

Special forms also develop in karst regions where the limestone rocks give rise to typical forms. Soil erosion also occurs here in specific forms, and it is therefore possible to speak in terms of a special erosion type – *karst erosion.* The specific features of karst erosion arise from the specific character of the limestone bedrock, its broken ground, permeability, weathering properties, etc. Grike erosion could perhaps be considered as a special type of erosion giving rise to various hollows, crevices and channels which are typical of limestone rocks. The forms of this erosion resemble neither the forms of area erosion, nor those of linear erosion, and share no likeness with badland erosion, either. Mostly it is the well-known phenomenon of corrosion that occurs, with the predominantly chemical characteristics of polymorphic erosion. But this type of erosion could also be classified as a subtype of rock erosion, and would therefore be linked to the kind of soil erosion which is observed in the typical karst surface and underground forms.

Specific types of rock erosion which may also occur in soils are *pressure erosion,* or *effodation* (Latin *effodere* – to dig out) (according to Bock), and *whirl erosion,* or *evorsion* (Latin *evorsere* – to whirl).

2.2.2.2 Underground erosion

Precipitation causes erosion through the effects of both surface runoff, and intrasoil or underground runoff. As mentioned earlier, the author proposes to refer to the group of erosion phenomena which arise in this way as *cryptoerosion.* Within this group *intrasoil erosion, tunnel erosion,* and partly also *karst erosion* are distinguished.

Intrasoil erosion

Water entering the soil plays a very important ecological role, and is also a soil-forming factor. Its various forms of motion bring about the vertical translocation of soil components which contribute to the formation of soil horizons and thus give the pedosphere its typical stratification. As a rule, the greater the porosity and the larger the proportion of non-capillary pores in which the motion of water is unimpeded, the faster is the flow of water under gravity, and the more readily are soil particles washed out by intrasoil water flows and underground water flows. This phenomenon is very pronounced on gravel and stony soils where, after the removal of vegetation, the soil is rapidly washed downwards into the coarse skeleton and is partially transported by the underground flow to the lower parts of the slope or into watercourses.

This form of erosion has often been described in the literature and interpreted in various ways. From among all the studies of the subject, the concepts of Sekera (1951) need to be mentioned first. He referred to intrasoil washout by the term *microerosion of the soil.* By this, he meant the mechanical washing of soil particles by gravitational water into the interaggregate space. In the author's opinion this is one form of interaggregate intrasoil erosion.

It may be assumed that intrasoil erosion can cause damage to shallow rendzina soil or to loess deposited on permeable limestone bedrock permeated by cavities through which soil may be washed into caves or other spaces. But the author considers the term *intrasoil erosion* to be more appropriate for this phenomenon, whereas the term microerosion should be reserved for small surface erosion phenomena occurring as a part of sheet erosion.

Another interpretation of intrasoil erosion is given by Kohnke and Bertrand (1959), who describe a similar phenomenon in so-called geological erosion. They include *leaching, surface erosion, landslides,* and *oxidation* within the meaning of geological erosion, and describe as leaching the dissolving of minerals and organic matter, and the translocation of these as solutes carried by vertical or lateral runoff with eventual transportation into the sea. They consider the washing out of calcium by the weak acids present in rain-water to be a typical example of geological erosion by dissolution. In an other instance, in the context of soil erosion, they distinguish *internal erosion,** a term which they use to describe the washing-in of soil particles into soil crevices. They point out that soil does not escape from the field in this process.

As a final example from among interpretations of intrasoil erosion, we may mention the work of Gorshenin (1959), who carried out some of his research in the Carpathians. He investigated steep slopes strewn with sandstone rocks and boulders. Detailed research revealed that after removal of the forest stand, the fine earth and humus were rapidly washed between the stones into deeper layers, and then were carried by underground waters into the river. He called this phenomenon *vnutripochvennaya éroziya* (intrasoil erosion). Its effects are not negligible since on slopes with an inclination of between 28 and 35° the average removal of eroded material was found to be 30 to 50 t ha^{-1} of humus and fine earth from the upper part of the slope, and 30 to 130 t ha^{-1} from the lower part. Where the soil is stony this represents a high loss.

Referring to the interpretations given in the above-mentioned works, intrasoil erosion may be defined, so far, as the washing of fine earth fractions by gravitation-

* It should be noted that Ilin has used the term *vnutrennaya denudatsiya (internal denudation)* for *(suffosis)* (Pochvovedenie 1935, No. 1).

al water through aggregates and coarse detritus is causing the *skeletonization of the soil* throughout the whole profile. The washing out of particles has both an ecological impact and a geomorphological importance because it diminishes the stability of the detritus and thus indirectly accelerates erosion of the slope. It goes without saying that water percolating through cavities *(transmittent water)* will also act chemically, and easily soluble matter will readily be washed away. What is important is the fact that in this process changes of form occur inside the soil, such as the broadening of pores, cavities, crevices, and the processing of weathered materials, etc.

The term intrasoil erosion could be taken as synonymous with the new phrase *intrasolum erosion* (Latin *intra* – inside, *solum* – soil), referring to one form of underground erosion. The expression internal erosion has a broader meaning and could be compared with the author's expression cryptoerosion, although the former term seems to be less explicit.

Tunnel erosion

Tunnel erosion is a specific form of underground erosion which was first described by Richthofen as *well erosion (Lössbrunnenerosion)* when it was discovered in China in 1872 (according to Schultze 1952). It occurs mostly in loess regions and involves the washing out of subsurface corridors by underground waters which accumulate over impermeable bedrock in the form of flows. Because of the widening and deepening of these channels or tunnels as erosion continues, and the consequent weakening of the ceiling, the stability of overlying layers is impaired. The final stage of tunnel erosion (after the collapse of the ceiling) is gully erosion, and therefore it is considered by Schultze (1952) and other authors to be a special case of gully erosion.

In the Russian and Polish literature tunnel erosion is referred to as *suffosis,* and has been described as such by Pavlov (1894). Apparently the word is derived from the Latin *suffodere* – to undermine. In the Geological Encyclopaedia the term suffosis is explained as the continuous dissolving and washing of the cement or soluble parts of rocks. In this sense the term *suffuziya* was also used by Rodionov in the classification of landslides; he distinguishes *suffosis landslides* as one particular form of landslide. Today the expression suffosis is understood in a broader sense and therefore the author recommends that *suffosis erosion* (as a synonym for tunnel erosion) should refer to that type of suffosis which causes *underground erosion.*

In the English literature, the terms *tunnelling erosion* (Downes 1946), *tunnel-gully erosion* (Gibbs 1945), *rodent erosion* (Bond 1941), *soil piping* (Carroll 1949), *piping erosion* (Brown 1962), *tunnelling,* and finally *piping* (Fletcher and Harris 1952) are used besides tunnel erosion.

Fig. 34. Erosion funnel – the beginning of tunnel erosion on the excavation site of a road (Cataula, Oregon, USA). (The author's collection of photographs.)

Fig. 35. Funnel development in loamy foothills (Tadzhikistan, USSR). (Photo D. Zachar.)

Fig. 36. Tunnel erosion in Ordovician sediments in the central region of Victoria. The funnels, which are connected by an underground corridor, increase in size and after the ceiling collapses they develop into gullies (see also Fig. 25). (By courtesy of Soil Conservation Authority of Victoria, Australia.)

Because the forms of suffosis erosion resemble the karst, territory modelled by this type of erosion is also called *pseudokarst,* or *sham karst.* According to the rock type in which erosion occurs, distinction is made in the literature between so-called *clastokarst* (referring to fine clastic material), *loess karst* (referring to loess layers), and *clay karst* (referring to clay rocks, etc.). Panov (1966) and others also distinguish *thermic karst* as a separate type of sham karst which occurs in regions with permafrost rocks and a periglacial climate. In this connection it is possible to speak of *pseudokarst,* and *thermokarst erosion,* respectively.

The most detailed description in the literature is an account of *clay karst* (Gvozdetskiĭ 1954, Lilienberg 1955, 1962) which occurs predominantly in regions with a semiarid climate where the soil and geological substrata are deeply fissured. Rain-water entering the crevices after a drought period of 3 to 7 months, washes the crevices and creates vertical opening with different forms and dimensions (wells, funnels, chimneys, etc.). To the vertical forms are linked underground forms

Fig. 37. View of an erosion gully during the collapse of the ceiling (Uzbekistan, USSR). (Photo D. Zachar.)

which allow the water to pass over inclined layers to the local erosion base. Because in this type of erosion there is a broadening of vertical openings to resemble wells and a widening of underground tunnels, a more appropriate term would perhaps be *well-tunnel erosion*. The depth of the wells varies from 5 to 10 m and more, the lengths of tunnels reaching a few tens or hundreds of metres, before opening into gullies, gorges, rivers, etc. (Figs. 34–36).

Well-tunnel erosion may be considered as representing a transitional type between intrasoil and cave erosion on the one hand, and underground and surface erosion on the other. The labyrinth of vertical and underground channels becomes more complex and the corridors enlarge until the original mantle is reduced to a collection of funnels, pits, bridges, earth columns, and other forms which finally give way to a network of surface erosion gullies. Interesting patterns of this kind were observed by the author in the basin of the Zeravshan river near the city of Samarkand (Figs. 37–39), where all stages of *underground-surface erosion* were

Fig. 38. A maze of funnels, corridors, bridges and other forms produced by tunnel and suffosive erosion (Uzbekistan, USSR). (Photo D. Zachar.)

observed. Pseudokarst erosion phenomena are also commonly found in the USA, Australia (Fig. 36), in Asian loess regions and elsewhere. They are one of the causes of the origin of badlands (Fig. 40).

In some countries tunnel erosion is obviously a highly damaging factor in agriculture and its control is difficult, as has been proved for example in Tasmania (Colclough 1965).

A survey of forms of precipitation erosion is given in Table 2.

Table 2. Classification of precipitation erosion by form

Term	
English	International
1 Surface erosion	Exomorphous erosion
1.1 Sheet erosion	Areal erosion
1.1 Gully erosion	Linear erosion
1.3 Polymorphous erosion	Polymorphous — badland erosion
2 Subterranean erosion	Cryptomorphous erosion
2.1 Intrasoil erosion	Intrasolum erosion
2.2 Tunnel erosion	Suffosive erosion
2.3 Sham-karst erosion	Pseudokarst erosion

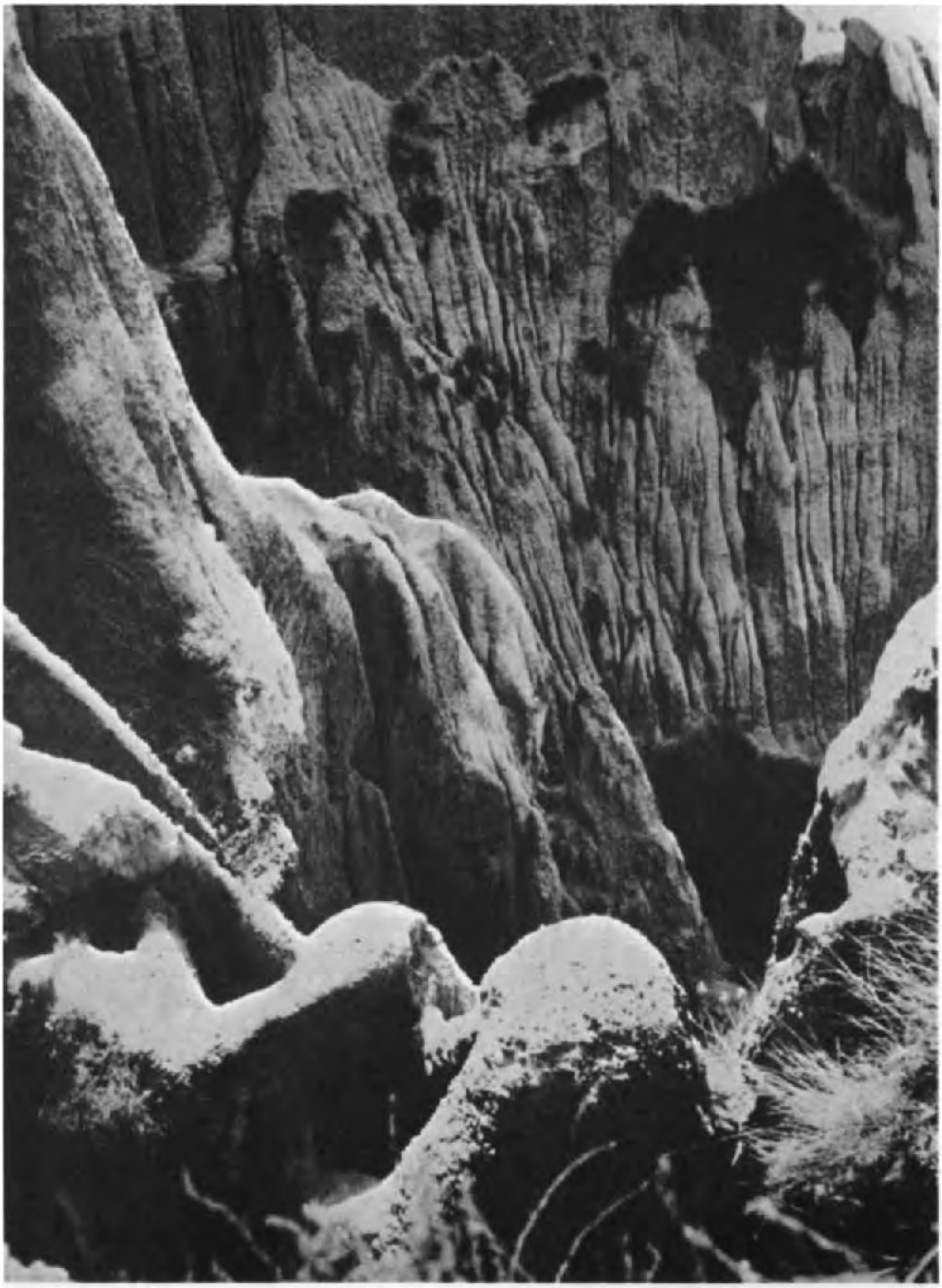


Fig. 39. View of bizarre forms of karst in a loam soil in central Romania (Repa Roşie, Romania). (Photo D. Zachar.)

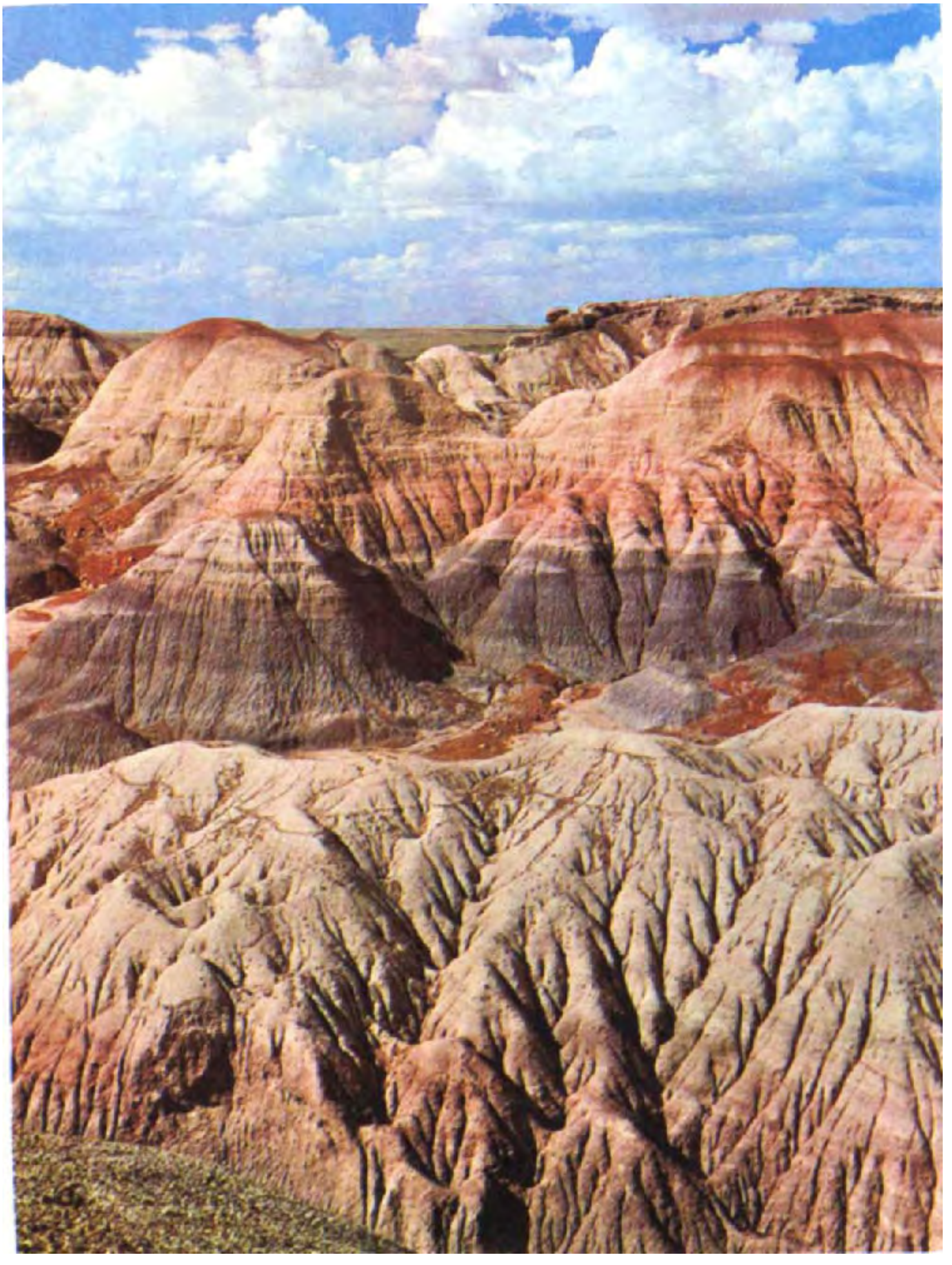


Fig. 40. Badland Blue Maze in Petrified Forest National Park (USA). Funnels are clearly visible and are a reliable indication of erosion underground forms. (Photo W. Davis.)

2.2.2.3 Forms of river soil erosion

River (fluvial) erosion occurs where there is a permanent water flow and usually shows a varying intensity as the flow of the water varies. The smaller the catchment area of the watercourse, and the less favourable are conditions for discharge, the greater is the fluctuation of erosion intensity. The uppermost branches resemble gullies and therefore constitute a transitional form between river and gullies. The boundary line between the hydrographic network and gullies remains arbitrary, especially in semiarid and arid regions. In particularly difficult cases the author suggests that the age and location of the gully could be used as criteria, taking recent and slope rills to indicate precipitation erosion, and older and bottom gullies to indicate river erosion.

According to the prevailing direction of influence, distinction can be made between *vertical* or *bottom erosion* which deepens the profile and compensates the erosion curve, lateral erosion which broadens the river bed and may cause a change

Fig. 41. Hollow formed on arable land during flooding of the Hron river (central Slovakia, Czechoslovakia). (Photo D. Zachar.)

in the direction of flow, and *regressive* or *retrograde erosion*. From this point of view gully and river erosion are very similar, but river erosion changes the total surface of the watercourse only to a small extent, and damage to soil only arises, in general, by lateral movement of the course of the river as it meands. The area covered by gullies may considerably increase at the expense of agricultural land. In gully erosion the typical action is regressive erosion, in river erosion it is lateral erosion.

In this connection, it is possible to speak of *river erosion of the soil,* occurring along banks and during flood conditions (Measnicov and Nitu 1965, Mizerov 1966). Under the influence of this process various kinds of undermining action may occur together with slips and rifts of banks and slopes; during floods surface wash, gullies, hollows and other forms may also occur (Fig. 41).

2.2.2.4 Lake and sea erosion

As far as soil is concerned, both of these forms of erosion are restricted to the *littoral* zone where *erosion,* if the water is non-tidal, is caused by surf water, and is therefore known as *surf erosion.* Besides the surf, the tides also cause erosion, the destructive action of which may be called *tidal erosion.* In a body of water confined by a dam, erosion occurs in several planes on account of the fluctuation of the water level, thus creating steps or stairs in the banks. The author recommends that this form of erosion be called *step erosion,* or alternatively *stage erosion.* As a matter of fact, it is a process of small-scale abrasion, and therefore this phenomenon could also be referred to as *microabrasion.*

Lake erosion, and especially *sea erosion* create a great wealth of forms and sizes which give littoral regions varied, but always specific appearances. In the context of soil erosion specific forms occur on shores composed of soft rocks which succumb to intensive abrasion. (In marine terminology, abrasion generally refers to the rubbing away of the coastline.) Water which is constantly in motion, although it may be non-tidal, undermines the littoral slope, wall, or cliff, which by gradual weathering and modelling has a tendency to assume a natural incline. In this way the coastline is constantly being pushed inland. For example, it has been observed that on some stretches of the West-German coast, land is receding at the rate of 2 to 4 m per annum (Glander 1956).

Because this type of erosion brings about the expansion of seas and lakes at the expense of land and soil, the author proposes to refer to it as ingressive, or intrograde erosion. The opposite of *ingressive,* or *intrograde erosion* is *regressive,* or *retrograde erosion* which erodes the soil in the uppermost reaches of the hydrographic network and in erosion gullies. A poor compensation for this process is the constructive action of water in river deltas where the land advances into stagnant expanses of water. This activity is nevertheless linked with soil destruction in the catchment area and therefore the balance of destructive and creative action weighs heavily against the land and leads to the reduction of the surface soil cover and a loss of soil fertility.

2.2.3 Forms of wind erosion

2.2.3.1 General

The velocity and direction of the wind relative to the relief of the terrain may vary greatly. In contrast to the action of water, *wind* or *aeolian erosion* affects usually the whole surface and only seldom surface strips, though even these strips have no constant air flow. Consequently the forms of wind erosion are governed by the characteristics of the air circulation, the configuration of the landscape and the structure of the substrata. It should, of course, be added that the surface of the terrain also affects the properties of the wind and thus influences its action.

In general, wind levels and rubs away protruding features of the landscape, and therefore wind erosion has the greatest effect in those places where the wind acts upon the landscape tangentially. Thus the wind damages the ridges of the terrain first of all, and no deposits are laid on such places. Consequently the soil on ridges is rapidly removed by the erosive action of the wind which thus denudes the bedrock. In addition, the air tends to be compressed above protruding features of the terrain, so that wind velocity is increased and the erosion effect is augmented.

Diagonal currents increase the type of air turbulence which is responsible for lifting particles and creating hollow forms, mainly in the less resistant places. Ascending currents represent a special case in which there is an upward action, unlike that of water. These currents arise from thermal turbulence and other convection currents, and are usually associated with tempest conditions; the resulting air turbulence produces a partial vacuum which is capable of lifting particles to a considerable height and transporting them over large distances before they fall, together with condensates, back to the ground.

Finally, wind may act perpendicularly to the substratum. This happens mainly in the case of cliffs and rock walls where the wind, acting in a direction more or less perpendicular to the rock face, produces various hollows and honeycomb forms. The character of these forms is, of course, much influenced by the internal structure of the rock.

Thus depending on the direction of the wind, with respect to the substratum on which the wind is acting, different forms may arise. On the windward side of ridges and cols, there is a predominant *attenuation* of the soil cover (Fig. 42), whereas on the plains the *levelling* of surface unevenness and the *drifting* of soil in the direction of the prevailing wind is predominant, and on slopes, cliffs, walls, protruding stones, etc., *abrasion* and *gnawing* are the main effects.

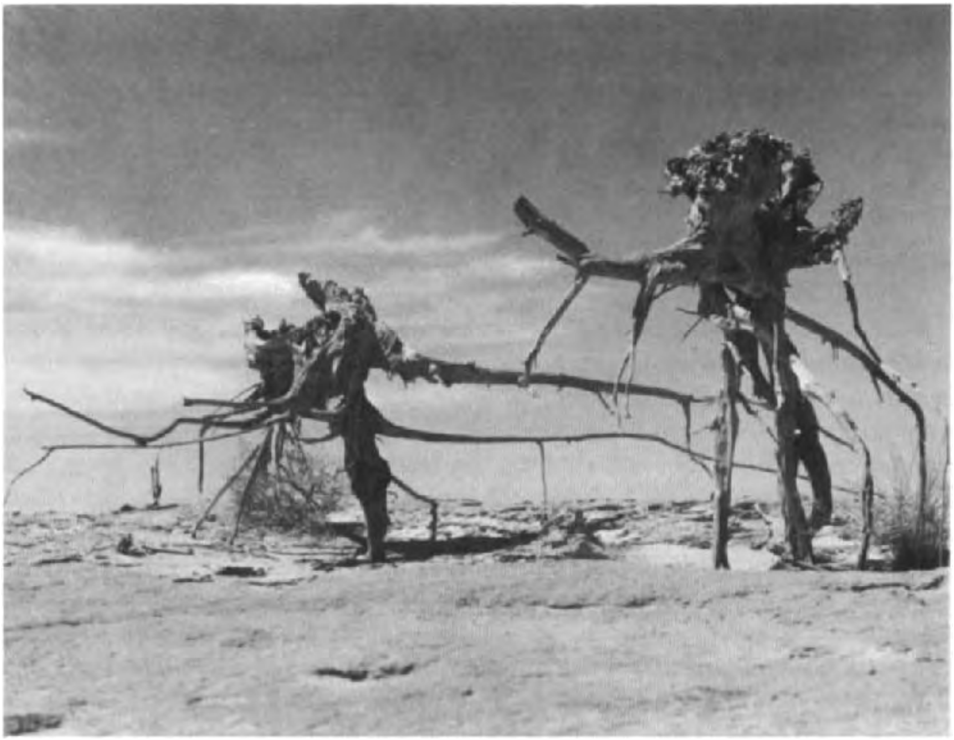


Fig. 42. Soil completely laid to waste by intense wind erosion. Erosion was accelerated by the removal of the vegetation cover and injudicious management (Victoria Malee, Australia). (By courtesy of Soil Conservation Authority of Victoria, Australia.)

Wind erosion causes a very distinct sorting of soil particles, together with skeletonization of the eroded soil and a granular homogenization of particles transported by the wind, all of which effects are included in the meaning of the term *aeolization,* representing a characteristic process of both soil destruction and soil formation.

All these forms depend not only on wind direction, but also on the wind force as a function of its velocity. At low velocity, only the smaller particles may be picked up and then perhaps carried only for a short distance. With growing wind force the kinetic energy of the wind increases and with it also the capacity for transportation. Finally, the wind also carries away sand particles which, especially if they are hard, serve as an implement of wind erosion in the abrasion of protruding soil or rock masses.

Two forms of erosion are generally distinguished, namely *erosion by deflation* (Latin *deflare* – to blow away) in which the wind carries away loose particles, and *erosion by abrasion,* or *aeolian corrasion* (Latin *corradere* – to scrape off) in which abrasion is caused by wind-carried particles. Deflation occurs mainly on loose rocks and soils, abrasion on hard rocks; however, corrasion also occurs as part of the process of deflation, and likewise during corrasion the abraded particles are, at the same time, carried away. Kettner (1955) notes that deflation without corrasion may occur, but not corrasion without deflation.

2.2.3.2 Deflation

The concept of wind erosion of the soil is usually associated in the mind with soil erosion in terms of the blowing away of loose particles, i.e. deflation. Because of this, the terms wind erosion and soil deflation are often wrongly taken to be interchangeable in the literature. As mentioned earlier, the effects of deflation are levelling of the soil surface, attenuation of the soil cover, and the creation of various erosion relics.

As the simplest form of deflation, the dislodging of soil particles from soil clods by the wind and their deposition in holes, furrows, and depressions may be considered. As a result of this process the unevenness of the ground is reduced and its surface levelled. The author proposed that this phenomenon – in agreement with Spirhanzl (1952) – be called *detrusion* (Latin *detrudere* – to push), i.e. the disintegration of soil clods. In detrusion particles are usually transported over short distances by the direct force of the wind. However the wind does not necessarily influence soil particles directly, but may act through the medium of other carried particles. The latter form the basis of sand drifts. During winter, pressure on the soil comes instead from flying ice or snow crystals. The bombardment of soil particles by others soil particles, and their removal by the current is called *extrusion*. This expression stems, obviously, from the Latin word *extrudere* – to thrust.

Both *detrusion* and *extrusion* occur when the wind direction is tangential or diagonal, the soil particles being moved by the wind's *pushing* force, and then *rolling* or *volvation* (Latin *volvere* – to roll), or leaping forward – *saltation* (Latin *saltus* – jump). A common feature of both phenomena is that particles start moving by air pressure or by impact from other carried particles. The various subsequent types of motion closely interact with one another. By detrusion the surface is first levelled, and when the force of the wind increases, a larger proportion of the particles start to move by rolling (volvation) and jumping (saltation). The quantity of material in motion increases as the mean velocity of motion increases and, consequently, the effect of the moving particles increases as they strike the soil surface and send other particles into motion. The broader the eroded expanse in the direction of the wind current, the larger is the number of transported particles. The phenomenon of mass particle movement was termed *avalanching* by Chepil.

Soil particles may also be blown away when air is drawn into the partial vacuum of an air funnel. This phenomenon can be seen to operate in the slipstream of a moving car, but also occurs as an aspect of the convection currents mentioned earlier. In this form of deflation, soil particles are lifted by the motion of the air, carried into the atmosphere and transported over long distances. This type of wind erosion is referred to in the literature as *efflation* (Latin *efflare* – to blow out).

It may be generally stated that in Czechoslovakia the pressure or dynamic effect of the wind mostly occurs in winter and spring, and the turbulent or vacuum effect occurs mostly in summer, mainly in anticyclonal weather during thunderstorms when there is increased vertical turbulence. In this way the well-known *dust storms, black storms,* or *sand storms* originate, known by different local names in different countries. Their occurrence is accompanied by *blood rains* and the *descent of dust* or *dust falls.*

2.2.3.3 Wind corrasion

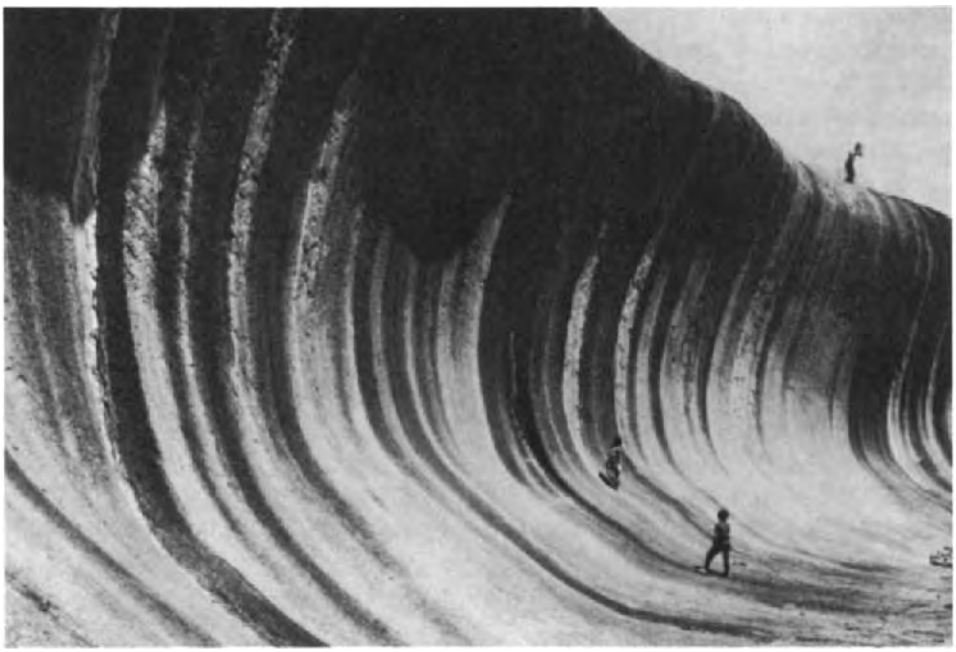
Easily eroded rocks such as sandstones are particularly vulnerable to wind, or *aeolian corrasion.* From among the best-known forms created by abrasion come the terms *wind-shaped, wind-cut, wind-worn, aeolian, faceted pebbles, glyptoliths, aeologlyptoliths* (Dylik 1951), *ventifats, honeycomb aeroxysts* (Latin *aer* – air, *xystus* – covered corridor) (Fig. 43), *rock columns, balanced rocks, rock windows, rock bridges, steep rock cliffs,* etc.

A special phenomenon is represented by so-called *jardangs* which arise on clay soil in arid regions. They consist of parallel rills separated by regular ridges which are similar to grikes and are caused by wind corrasion. Sven Hedin (1905 in Stejskal 1949), observing jardangs in the desert of Central Asia (Turkmenistan) calculated that the excavation of wind rills of 6 m took about 1,600 years (i.e. the growth rate was about 4 mm per annum).

The rubbing and polishing of rocks by the wind-borne grains is also referred to as *wind (aeolian) detersion* (Russian *vetrovoe shlifovanie*) (Danilevskaya and Yakovlevskaya 1962). This process proceeds much more slowly than the blowing away of loose soil particles, the intensity of abrasion depending mainly on the resistance of the material and the wind velocity. According to Cailleux (1942) the wind corrasion of a rock surface results in the removal of a layer of about 1 mm thickness per century (in Lukniš 1958). The long-term effect of aeolian corrasion is shown in Fig. 44, giving an indication of the work of wind driven particles throughout geological periods.

Fig. 44. The "Wawe Rock" granite wall (near Hyden, western Australia) polished by wind-carried sand. Coloured strips are caused by water flowing down the rock. (By courtesy of Western Australian Government Printing Office.) ▶

Fig. 43. Honeycomb aeroxysts in the Prachovské skály "sandstone rock town" (Czechoslovakia). (Photo D. Zachar.)



2.3 Classification of soil erosion by intensity of removal

2.3.1 General

Erosion intensity can be expressed in different ways. In sheet and wind erosion the intensity is usually expressed in terms of *soil loss* or *soil removal* measured in m^3 per ha, or tons per ha. For small erosion intensities or isolated cases of *removal, values* are given in kg per ha, and for long-term erosion phenomena the annual average, or aggregate value is given. Besides volume and weight data, the intensity of linear erosion can be measured by the *length* or *density* of gullies expressed in km per square km, km per ha, or m per ha; it can also be measured in terms of the annual length increment of the gullies or in terms of the proportion of active gullies as a fraction of the total length (i.e. gully activity).

Overall decrease of soil depth by erosion and other factors can be expressed in two ways: either as the time taken for the removal of a certain depth of soil, or as the depression caused by soil removed in a given time. In the first case the expression *denudation metre* is used, meaning the time taken for a soil layer of 1 m thickness to be removed. If planation (denudation) factors are limited to erosion only, it is possible to speak of an *erosion metre,* indicating the time during which a 1 m thick soil layer is eroded away. Conversely, erosion intensity can be expressed in terms of the erosion height – the depth of soil removed by erosion in one year or other convenient measure of time. The erosion height is usually given in mm per annum. According to the time period in which erosion removal is expressed, *erosion height* may be stated in terms of the annual change, the average annual change for *n* years, the total change for *n* years, etc. In addition to these methods of expressing erosion intensity yet other methods are used which are essentially modifications of the basic expressions mentioned.

As to the proper classification of erosion according to its intensity, it has been mentioned already in the chapter on erosion that the most important index of soil erosion is the intensity of soil loss or soil removal. It has also been shown that as far as damage is concerned *erosion* may be either *harmless* (benignant)* or *harmful* (malignant). In harmless erosion the rate of removal is less than the rate of soil formation, whereas in harmful erosion removal is predominant, and this brings about reduction of the soil profile (mantle) and its final destruction. Erosion in which the rate of soil destruction equals the rate of soil formation is termed *compensative erosion* by the author.

The object of control measures is to reduce damaging erosion to the level of compensative erosion, i.e. to bring *inhibited erosion* (p. 24) as near as possible to

* Many authors take the view that normal erosion is useful because it regenerates the soil and maintains the continuous flow of nutrients and energy in the process of pedogenesis (Bennett et al. 1951).

compensative erosion, or perhaps even below this level. The reduction of inhibitive erosion to a harmless level is important mainly where control measures are applied to already eroded soil, and the aim is to encourage soil formation, which under natural conditions tends to occur only under a permanent vegetation cover.

As a rule, the mode of *expression* of the *erosion* process changes as the intensity of removal of material increases. Erosion of small intensity is usually indiscernible and specific changes occurring on the surface of the eroded soil cannot be seen. In this case we speak of *latent erosion;* it is represented by low-intensity forms of drop erosion (guttation), and pellicular and laminar erosion. Visible forms are caused by *expressive erosion,* e.g. by the rill, gully, and badland, etc. forms of erosion. The influence of latent erosion on the soil is manifested as a prolonged, chronic phenomenon, whereas expressive erosion is a sudden, acute phenomenon. Thus we may distinguish between prolonged or *chronic erosion* and sudden or *acute erosion.*

In order to classify erosion by its intensity it is first necessary to determine the *intensity of soil formation* due to weathering, since this is decisive for the determination of harmless, and compensative erosion, respectively. Therefore, information on soil weathering constitutes important theoretical erodological data. In the next chapter the author confines himself to discussing only that information which has a direct bearing on the intensity of weathering and the determination of the intensity of *compensative erosion.*

2.3.2 Compensative erosion

The subject of *soil formation* generally includes the origin and development of the soil mantle under the influence of soil-forming factors. The rate or intensity of formation depends, most of all, on the substratum and its properties. From the quantitative aspect the hardness, and the state of weathering of the substratum are important since they determine the rate of formation of weathered material over the surface of the unweathered bedrock. The latter process of soil formation is slow. More rapid is the situation in which weathered material accumulates on slopes as loose detritus or sediments which may build up into thick layers. In this case the rate of soil formation as a consequence of erosion is not so important, especially if the soft rock has favourable ecological properties.

A detailed study of this problem was made by Kohnke and Bertrand (1959). They found that a soil layer 90 cm thick had been formed in about 16,000 years, in a temperate climate, on moraine material; a 5 cm soil layer developed on material displaced by man after 100 years, and a 17.5 to 25.0 cm soil profile had developed on a sand dune 100 years after its fixation. Examples given in a general survey show that in various cases it has taken from 10 to 857 years for the formation of a 1 cm layer of soil on carbonate moraine, this being equivalent to an annual soil formation of 157 to 13,440 kg ha^{-1}. High rates of soil formation ranging from 6,272 to 13,440 kg ha^{-1} were found on artificially raised rocks, whereas on undisturbed rocks soil formation did not exceed 1,000 kg ha^{-1} $year^{-1}$. In general, the shallower the soil cover, the more rapid is the rate of soil formation, and vice versa; after a certain depth is attained (depending on natural conditions), the growth of the soil by weathering becomes stabilized. In most cases a depth of 20 to 30 cm inhibits soil formation because the influence of changes in microclimate is greatly attenuated at this depth.

Bennett (1955), referring to data given by Chamberlin (1909), also mentions that soil is formed relatively quickly at shallow depths. He suggests that a 2 to 3 cm layer of soil formed from the bedrock takes, under very favourable conditions with a good vegetation cover and soil protection, from 200 to 1,000 years to develop. This means that a soil layer 18 cm thick takes 1,400 to 7,000 years to form. With this rate of weathering soil would be formed at the rate of 0.026 to 0.13 mm $year^{-1}$, i.e. 324 to 1,620 kg ha^{-1}. Kukal (1964), analyzing data from the literature with respect to the intensity of weathering under different conditions, comes to the conclusion that the average rate of soil formation over the entire surface of the Earth is about 10 cm per 1,000 years, i.e. 0.1 mm $year^{-1}$, or 1 m^3 ha^{-1}.

From among the constituents of weathered detritus the most important is clay which, according to Barshard (1959), is created most rapidly in the uppermost layers (2 to 10 cm from the surface). Barshard estimates that for each 100 g of bedrock, 0.00001 to 0.002 g of clay is formed per annum. If weathering takes place within a 1 mm thick layer, then from 1.5 to 300 kg (average 150 kg) of clay are created per ha per annum. These figures are in good agreement with the data mentioned earlier. All this information is, of course, only of illustrative value, since the intensity and quality of weathering are very variable and change with respect to time and depth although other conditions may be constant. Gorbunov (1963) has established that the greatest weathering intensity occurs when vegetation of higher growth, especially forest growth which accelerates the process of weathering and soil formation, develop on the nascent soil. The same has been observed by the author in his research on the influence on soil formation of forest stands established on devastated land with a dolomitic bedrock (Zachar 1966).

The reliability of the values for natural soil formation under conditions of formal erosion may be checked by the complementary method of determining the intensity of soil removal from virgin soils with a "normal" soil profile and with protection from vegetation. Information to this end was obtained from American data by Smith and Stamey (1965). They evaluated soil losses due to erosion on 12 experimental plots located in different parts of the USA, each covered with close-growing vegetation and situated on slopes ranging from 1 : 8.3 to 1 : 1.6 inclination. They found that in natural plant associations the annual removal varied between 0.05 and 0.30 t $acre^{-1}$. Taking into consideration the fact that erosion removal under natural conditions could be twice as great, they concluded that normal erosion would vary between 0.1 and 0.6 t $acre^{-1}$, i.e. between 0.25 and 1.48 t ha^{-1} $year^{-1}$. As can be seen, these data correspond with the assumed rate of soil formation.

Sheet erosion

It may be supposed, following from these figures, that soil loss amounting to about 0.05 mm $year^{-1}$, or 0.5 m^3 (approximately 750 kg) ha^{-1} $year^{-1}$, will be damaging in so far as new weathering will not compensate the loss. In sheet erosion and deflation this value could be considered as the boundary between benignant and malignant erosion, i.e. the value of compensative erosion. The figure corresponds with the data of Bennett (1939, 1955, 1958), Kohnke and Bertrand (1959), Smith and Stamey (1965), and other authors. It may be expected that in temperate regions the compensative erosion value will be lower on harder rocks covered by naturally shallow soils, but the value will seldom fall below 250 kg ha^{-1}, compensative erosion will be higher on softer rocks, but probably will not exceed the limit of 1,500 kg ha^{-1} $year^{-1}$.

The author supposes that the values of 0.5 m^3 must not be exceeded, mainly because the fact that with growing depth the weathering intensity and soil formation changes, as well as for the reason that erosion affects the soil selectively and its fertility decreases even with low soil losses. The shallower the soil, the more dangerous is soil removal; the weaker the erosion, the more selective is its influence under a given set of conditions. Weak erosion removes the finest, lightest and most soluble soil components which are of great importance in the formation of soil with desirable properties, especially soil of high fertility. Therefore a loss exceeding 750 kg ha^{-1} $year^{-1}$ can be severe when it occurs continuously over a long period.

2.3.3 Permissible erosion

Compensative erosion indicates the ràte of soil removal which is permissible from the point of view of permanent soil conservation. In nature there are many cases in which compensative erosion is a theoretical goal to be attained by applying protective measures (erosion inhibitors). However short-term goals need not to be set at this value; if the soil has been seriously eroded already, a lower value than the rate of compensative erosion may be acceptable; also if the soil consists of large deposits of fertile sediments (e.g. loess) a more intensive rate of erosion may be tolerated without seriously impairing the fertility of soil.

Erosion which involves soil formation on the one hand, and conserves soil fertility at the same level on the other, is referred to by Smith and Stamey (1964, 1965) as *permissible,* or *tolerance erosion.*

These authors calculate permissible erosion by takig into account: (a) the actual soil stock, (b) the essential properties which are going to be required of the soil, (c) data on expected erosion losses, (d) data on future soil formation. It is implied in this that the more the soil is eroded, the lower is the permissible erosion; also the greater the depth of the soil and the higher the quality of the bedrock, the higher is the erosion value. Depending on various natural influences and economic factors, tolerance erosion may vary from zero up to values which should, in the author's opinion, exceed the limit of compensative erosion with rare exceptions.

Smith and Stamey (1965) report that in the USA various authors in different regions have calculated the levels of tolerance erosion which were recommended as a minimum requirement in the application of erosion control measures, and obtained values in the range 0.5 to 6 t acre^{-1}, i.e. 1.24 to 14.84 t ha^{-1} year^{-1}. It may be understood from the context that this represents the highest measure of permissible erosion in which the rate of soil removal is greater than the rate of soil formation.

A corresponding values for permissible erosion was given by Kohnke and Bertrand (1959), who consider that a rate of removal of 3 t acre^{-1}, i.e. 6.75 t ha^{-1} year^{-1} removal is dangerous for soils on glacial moraines in the Middle West of the USA. According to these authors, the intensity of permissible soil erosion loss depends on permeability, and the depth of the soil profile depends on how much the soil is affected by erosion.

It must be added that the actual, short-term view is not sufficient as a basis for making calculations of permissible soil loss. In many cases fertile sediments are deposited on rocks which possess undesirable ecological properties. In these cases it is preferable to reduce soil loss to a minimum, because the quality of the land will decline considerably after the removal of the upper soil layers, regardless of the rate of weathering in underlying strata. As well as this, the lowering of the soil surface will result in a shallower humus horizon, and nutrients that have accumulated in the upper soil layers will be washed away, thus impairing soil fertility and retarding any possible increase.

2.3.4 Classification of harmful erosion

We now come to the question of the criteria by which harmful, or malignant erosion should be classified. The most important criterion is the rate at which damage or destruction of the soil mantle is occurring. Verbally, degrees of erosion can be expressed as weak, medium, serious, severe, and catastrophic erosion. As to quantitative value, the author recommends the classification of erosion by intensity of removal, as follows:

Erosion causing an annual soil loss of between 0.05 and 0.5 mm, i.e. from 0.5 to 5 m^3 ha^{-1}, can be considered as representing weak erosion. Below the 0.05 mm limit erosion is harmless (benignant), and above 0.5 mm it is of intermediate intensity. If the soil is not shallow and the selection of material not intensive, erosion losses, though they may be harmfull do not involve major seasonal losses. Under the impact of erosion of this degree, a 20 cm thick layer of topsoil would take 400 years to be removed and losses of nutrients would represent only a minor fraction of the nutrients taken up by crops during this period and could easily be replaced by the application of fertilizers. The upper limit of weak erosion is basically equivalent to tolerance erosion, according to Smith and Stamey (1965).

Erosion which brings about an annual loss of 0.5 to 1.5 mm of soil, i.e. from 5 to 15 m^3 ha^{-1} $year^{-1}$ can be considered as representing medium erosion. At this level of erosion, approximately the same quantity of available nutrients is removed as that taken up by plants in a year. Maintaining fertility at the same level requires double the amount of fertilizer compared with cases of weak erosion, and inaccessible nutrients which may form the basis of future soil formation are lost permanently. At the rates of erosion mentioned, a 20 cm thick layer of topsoil would be removed in 133 to 400 years.

Serious erosion represents great danger to the soil because the topsoil or upper humus layer is carried away in a period representing one human generation. If serious erosion is taken as that causing a soil loss of 1.5 to 5 mm, i.e. from 15 to 50 m^3 ha^{-1} $year^{-1}$, this means that the topsoil will be removed in 40 to 133 years, and the rate of nutrient loss will be several times greater than the rate of uptake by vegetation. The losses due to erosion will only partly be replaced by the usual applications of fertilizers.

Severe erosion is an extreme danger to the soil since it can destroy soil in a relatively short time. Ranging from 5 to 20 mm annual soil loss, it removes the topsoil in 10 to 40 years, or even after a few heavy downpours or dust storms. Erosion of this intensity may also cause very heavy damage in a particular year when so-called seasonal damage occurs.

A further increase in the intensity of erosion usually has disastrous consequences for the soil because in heavy downpours or storms the entire topsoil as well as deeper layers are eroded. With an average rate of removal above 200 m^3 ha^{-1} $year^{-1}$, practically the entire topsoil is demolished by rills. In isolated instances the author has observed rates of removal exceeding 1,000 and even 2,000 m^3 ha^{-1} $year^{-1}$, the topsoil being destroyed almost entirely. The author refers to such erosion as *catastrophic* erosion.

As the intensity of erosion varies, so also do the erosion form, the rate of soil degradation, and the urgency of erosion control measures. At the outset the simplest control measures are sufficient, but later on more effective control measures are necessary. Schultze (1952) considers erosion in which the annual rate of removal is 4 m^3 ha^{-1} to be critical. The author's experience confirms that this rate of removal is harmful to plants (washing away seeds, laying bare the roots, etc.), and therefore in agreement with Schultze, the author considers erosion involving the removal of more than 0.5 mm (5 m^3 ha^{-1}) per year to be acute, and concludes by making reference to the strong selective effect of wind erosion, that the above-mentioned criteria for the classification of erosion by intensity of removal are also valid for this kind of soil destruction. A survey of the proposed classification is given in Table 3.

Table 3. Classification of sheet erosion and deflation by the intensity of soil removal

Grade	Intensity of soil removal [m^3 ha^{-1} $year^{-1}$]	Verbal assessment
1	< 0.5	No erosion, insignificant erosion
2	0.5–5	Slight erosion
3	5–15	Moderate erosion
4	15–50	Severe erosion
5	50–200	Very severe erosion
6	>200	Catastrophic erosion

Gully erosion

In gully erosion it would not be correct to express intensity in terms of quantity of soil removed from one hectare. A gully erosion usually represents a permanent loss of soil where agricultural production proceeds without appropriate protective measures and recultivation. Therefore any linear washing of the soil should be prevented. In this context the author classifies all active forms of gully erosion as malignant erosion. Further classification of gully erosion may be based on the density of gullies in km^2, the increase in overall gully length due to retrograde erosion in m per year, and the size of gullies.

A proposal for a classification of gully erosion by density was made by Bučko and Mazúrová (1958) for conditions in Czechoslovakia. Having standardized the data on gully density for the territory of Slovakia, they divided the distribution curve for gully length per km^2 into a six grade scale (Table 4). The second and higher grades of erosion are regarded as harmful erosion (Plesník 1958).

Table 4. Classification of gully erosion by total gully length (Bučko and Mazúrová 1958)

Grade	Total length of erosion gullies [km km^{-2}]	Verbal assessment
1	<0.1	Erosion nil or insignificant
2	0.1–0.5	Slight erosion
3	0.5–1.0	Moderate erosion
4	1.0–2.0	Severe erosion
5	2.0–3.0	Very severe erosion
6	>3.0	Catastrophic erosion

In the USSR Sobolev (1948) distinguished 11 levels of what he calls "ovrazh-nobalochnaya" erosion which, in the author's view, corresponds to past and recent gully erosion ranging from 0 to 1.1 km per km^2, each level representing a range of 0.1 km per km^2 within its limits. He added a twelfth level for mountain regions, where *gully density* is higher or the gully form changes into other forms.

In the author's opinion, the graded scale of Bučko and Mazúrová (1958) is generally preferable, since it has a broader scope and is better suited to the varied pattern of Czechoslovakia. It should be noted that Sobolev based his classification on the overall characteristics of large regions in which small areas with intensive erosion were ignored. When the investigated area is small, a more detailed classification of erosion becomes possible.

Although the density of erosion gullies is a good indicator of the intensity of linear soil erosion, it does not fully express its current activity. Therefore it is appropriate for the purposes of more detailed research work to take into account also the proportion of active gullies, or the activity of the latter in relation to stabilized gullies. A good index of gully erosion is also given by the rate of *gully growth* by retrograde erosion. The author recommends that gully erosion be classified in terms of six degrees of annual increment, the second and higher degrees in this case being harmful (Table 5).

Table 5. Classification of gully erosion rate of longitudinal gully growth

Grade	Growth rate of erosion gullies [m $year^{-1}$]	Verbal assessment
1	<0.5	Erosion nil or insignificant
2	0.5– 1.0	Slight erosion
3	1.0– 3.0	Moderate erosion
4	3.0– 5.0	Severe erosion
5	5.0–10.0	Very severe erosion
6	>10.0	Catastrophic erosion*

* High values for gully growth are quoted by Zanin (1962) who found that the mean annual gully growth in the Altai plains was 30 m. In 1955 some gullies grew by up to 70 m. The highest values for gully growth were obtained by Petrova (1962) in Novosibirsk, where a gully was observed to grow with an average annual increment of 90 m (over a period of four years), and the highest annual increment measured was 225 m. Similar values have also been recorded in the USA and other countries.

The intensity of linear erosion could be judged not only by the density and growth rate of erosion gullies, but also by other characteristics, particularly the *size of the gullies*. Of course, if gullies are assessed in isolation without regard for the land surface on which they occur, the classification will refer to the gullies and not to the eroded land. Setting up a generally valid scale is difficult in this case because the length of the gullies and the relationships between their various dimensions differ under different conditions.

The relationships between the various dimensions may vary widely according to the erosion conditions. Gully length depends mainly on slope length, gully depth depends on the thickness of the weathering mantle and the geometry of the slope, and gully width depends mainly on gully depth and the intensity of lateral erosion. Perhaps the most decisive factor in the classification of gullies is the gully volume, which gives a measure of the amount of erosion loss from a known surface. In the case of exceptionally large gullies further classification categories can be defined. Gullies are found which are several kilometres long, 50 m deep and 100 m wide or more from rim to rim.

Other forms

In other forms of precipitation erosion such as *badland* and *rock erosion,* the intensity of removal is always greater than the permissible level. These are the most destructive forms which end in the demolition of the pedosphere and lithosphere. By their action, no coherent soil mantle is allowed to develop, and the soil mantle is entirely destroyed. The author therefore recommends that both forms be marked separately on maps.

For subterranean forms classification criteria have not yet been developed and the study of these forms has not advanced very rapidly. In the author's opinion, it would be possible in the context of *intrasoil erosion* and the downward washing of soil particles into the subsoil, to adopt the criteria used for classifying of losses in shallow soils. In *tunnel erosion* the criteria are the same as in gully erosion. *Pseudokarst erosion* with visible forms belongs to the category of very dangerous erosion.

In *river erosion* sheet and gully erosion of the soil take place during flood conditions, and the undermining of the river banks follows as the result of lateral erosion. Whereas in the first (sheet and gully) form of river erosion the intensity of erosion may be judged by the rate of soil removal (Table 4), the criteria used to measure the degree of undermining should be different. The reason for this is, first, that soil lost by the undermining of one stretch of bank may contribute in part towards soil formation when the sediments are deposited on another part of the bank. Secondly, the volume of soil washed away by erosion of the bank depends both on the lateral shifting of the bank and on its height.

Taking into account that lateral erosion of 10 m year^{-1} in streams is not rare (Makkaveev 1955, Popov and Dekatov 1956), and in exceptional cases may approach several tens of metres over short stretches, it is possible, using as a basis the six category classification starting at 0.1 m year^{-1}, to draw up a provisional classification of lateral erosion in terms of the shift rate of the bank line (Table 6).

It should be noted that in larger streams rates of erosion are greater and the banks tend to be higher resulting in greater rates of soil removal. In the author's view soil loss is critical when occurs, on the author's scale, at the rate of 1 are to

Table 6. Grading of lateral river erosion by the rate of movement of the bank line and the rate of removal of soil

Grade	Verbal assessment	Lateral erosion [m year^{-1}]	Soil loss per 1 m bank height [m^3 km^{-1}]
1	Erosion nil or insignificant	< 0.1	< 100
2	Slight erosion	0.1– 0.3	100– 300
3	Moderate erosion	0.3– 1.0	300– 1,000
4	Severe erosion	1.0– 3.0	1,000– 3,000
5	Very severe erosion	3.0–10.0	3,000–10,000
6	Exceptionaly severe erosion	> 10.0	> 10,000

1 ha km^{-1} of watercourse, i.e. within a high range of values. It may be supposed that these losses are mostly compensated by soil formation following the deposit of sediments. More important are reductions in soil quality, because the newly formed land is usually of lower quality than that of the original land.

Solifluction erosion phenomena

As has been mentioned earlier, earth flows (aquasolifluction) represent a special case of denudation. Earth flows start from violent surface runoff where there is also an abundance of loose material in the river-bed or depression in which the water gathers. The surface runoff causes erosion both in the catchment area and in the hydrographic network as it carries away an accumulated mass of material. Also, the release of detritus and its displacement is a complicated process in which several denudation factors, including erosion, take part. Earth flows cause erosion by virtue of the movement of the slushy mass which deepens (vertical erosion) and rubs (detersion) the river-bed.

In any case, earth flows bear a narrow genetical relationship with erosion. Therefore, to include the classification of solifluction erosion phenomena in the chapter on erosion classification by the intensity of losses will not be a serious deviation from the central theme of this work. Using the same criteria as were used in the previously discussed forms of erosion, aquasolifluction phenomena may be classified by: 1. the amount of detritus removed in one flush, 2. the removal per unit surface area per year, 3. the active area as a proportion of the total surface of the catchment area.

The size of an earth flow depends, among other things, on the circumstances of its creation, the surface of the catchment area, and the frequency of occurrence, the rule being that the rarer the occurrence of a flow of a particular size, the larger is the flow. Ioganson (1962) refers to flows which recur every 1 to 3 years as very frequent, flows recurring every 10 to 15 years as very rare. As with floods, the most-feared flows are hundred-year flushes.

From data given in the literature and studied by Bogolyubova (1957) it may be inferred that in catchment areas ranging from 10 to 200 km^2 a one-time removal of detritus of volume up to 10^5 m^3 may be considered as low, whereas volumes above 10^6 m^3 may be considered as very high. According to Fleishman (1948), flushes of 4×10^5 to 8×10^5 m^3 can have disastrous effects. The largest one-time flow yet recorded in the literature, occurred in the large (223 km^2) catchment area of the Shinchay torrent where on 14th August 1955, about 10.5×10^6 m^3 were removed with a specific loss of 47×10^3 m^3 km^{-2}, or 470 m^3 ha^{-1}. In five selected small catchment areas of the Alazany and Araks rivers the loss varied from 47×10^3 to 87×10^3 m^3 km^{-2} (depth loss: 47 to 87 mm), calculated for the active area.

Data mentioned in the literature show that the rate of earth loss in earth flows varies between 10^3 and 50×10^3 m^3 km^{-2}. In sheet erosion and deflation the values vary between negligible amounts and values up to 20×10^3 m^3 km^{-2}; in an extreme case on a smaller plot over 10^5 m^3 km^{-2} was recorded.

It may be of interest to mention the classification of Kherkheulidze (1962) which is based on the active area of flows as a proportion of the total catchment area of the earthflows (Table 7).

Table 7. Grading of earth flow basins according to Kherkheulidze (1962)

Grade of damage caused in basin area	Severity of erosion	Active area of flows as a percentage of total basin area
1	Very slight	1– 3
2	Slight	3– 5
3	Moderate	5–10
4	Severe	10–20
5	Very severe	20–40

For the purposes of comparison it should be mentioned that the area covered by erosion gullies is usually lower than that covered by earth flows, and conversely, in the culminating stages of badland erosion, practically the entire surface of the catchment area becomes activated.

2.4 Classification of erosion phenomena by development

As mentioned earlier, erosion is part of the system of exogenous and endogenous forces influencing the changing relief of the Earth's surface. Together with other phenomena, erosion takes its place in the general development cycles which are typical for any one natural region, and which cannot, on the whole, be changed by any means of artificial intervention so far devised; however, the correct understanding of their nature can help in moderating their harmful effects and reducing the damage caused. Information on the development of erosion phenomena is the key to the correct application of erosion control measures.

In considering the development of erosion phenomena it should first be remembered that their character depends mainly on the *time* of occurrence and the magnitude of the *active* factor. When factors act continuously, erosion takes a regular course. There are few phenomena in nature involving *constant, unceasing erosion.* Often the erosion factor is confined to certain periods, appearing in seasons which recur according to a particular pattern each year. Such erosion is called *seasonal erosion.* Finally, erosion may occur sporadically, with long-time intervals in between events of extremely destructive effect. Such erosion is called *occasional,* or *episodic erosion.*

All influences (whether they be permanent, seasonal, periodic or episodic) manifest themselves by their effects on the soil and these effects may develop in the long-term, *annually,* in the short-term *(seasonally),* or *momentarily.* As a result of the long-term influence of erosion on the soil, *preterit* (Latin *praeteritus* – past) *phenomena* arise which have usually become stabilized. If these features are covered by younger deposits, they are also referred to as buried, or *fossil erosion phenomena.* At the opposite extreme from preterit phenomena are *contemporary* or *recent phenomena* which are usually active and still undergoing change.

The general rule is that the rarer the occurrence of particular erosion acting with a certain output of energy, the more apparent are both its destructive effect on the soil and the modelling of the slope. To illustrate the difference in the influence of erosion factors according to their frequency of occurrence and effect, the different effects of precipitation, firstly in regions with a temperate maritime climate and secondly in arid regions with occasional but violent downpours, is frequently cited. In the first instance rounded erosion forms arise and their development is continuous, while in the second situation the forms are sharp and their development is intermittent.

The second factor that determines the development of erosion forms is the *resistance of the soil* and *bedrock* against erosion, and this depends on *soil texture,* the *structure of the material,* the *solubility of its constituents,* the degree of *weathering* or *disintegration* of the rocks, the *stratigraphy,* and the degree of *soil protection* afforded by *vegetation,* etc.

All these varying properties cause erosion to act – not uniformly as might be expected in homogeneous material – but *selectively,* and this greatly influences the development of erosion phenomena. There are many factors which influence the development of erosion and they include practically all the factors and conditions which have some bearing on erosion in general. Next, some brief attention is given to some of the stages in the development of erosion forms.

In *sheet erosion* the selective activity of the slope runoff influences erosion development by removing from the soil surface initially the most easily washable particles, while larger and heavier material remains in situ, or is moved over short distances. In this way a coarse upper layer is created in the soil and this protects the soil from further washing; such a layer is called a *stone pavement,* or *cobble* *cover.** As the components of the primary skeleton of the soil become larger, the intensity of sheet erosion and its effect of gradually washing out the fine particles from the surface layers declines more rapidly. The same effect is also produced by the growing proportion of coarse detritus in the lower soil layers, where there is a transition from soil into bedrock and the skeleton and stones curb soil erosion, the latter showing a consequent gradual decline of intensity.

The intensity of sheet erosion also changes in the gradual erosion of *genetic soil horizons,* the rule being that the more differentiated the soil profile, the more pronounced are the changes in erosion development. The fastest erosion occurs with the sudden exposure of a humus horizon enriched by organic matter. Topsoil is also subject to rapid sheet erosion, being less resistant to erosion than the underlying soil. In lower horizons erosion takes a more varied course as is explained elsewhere in this work.

An important role is played by *vegetation* which may markedly affect the development of sheet erosion and erosion in general. Therefore all those conditions which influence the growth of vegetation also indirectly determine the development of erosion; vegetation provides the most effective instrument in slowing down erosion.

The same rules, as those governing precipitation sheet erosion apply similarly to *wind erosion* which also acts very selectively, so that removal from skeleton soils practically stops after the creation of a layer of stones.

These phenomena are well-known, particularly in deserts where *stone pavements* also *occur,* these being called *reg, serir* (stone pavement made of cobbles), and *hamada* (pavement with a preponderance of sharp-edged stones) in North African countries. In Australia stone deserts are referred to as *gibber plains,* in the USA as *scablands,* or generally as *boulder, boulder pavement,* or *rock plains.*

The author recommends the term *erosion flexibility* to refer collectively to the different forms of *adaptability of the soil to selective erosion;* erosion flexibility is high in skeleton soils, and small in fine-grain earth sorted by water and wind.

In *gully erosion* also the most intensive erosion occurs in the uppermost horizons and is greatly diminished on the bedrock which usually forms the main barrier to further *vertical erosion.* On the other hand, regressive erosion is usually limited by changing hydrological conditions in the catchment area. The surface of the gully catchment area above the line of activity diminishes as the growth of the erosion gullies proceeds upwards, and therefore the amount of affluent water decreases so that the *longitudinal growth* of the gullies gradually slows down together with elongation of the erosion curve on which further growth of the gullies in other directions depends. *Lateral erosion* depends on both vertical erosion and the

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* Another term appearing in the literature is *paved soil,* denoting a surface layer of stone lifted from the ground by frost.

natural angle of inclination of the slope. Finally, after the compensation of the erosion (declivity) curve, and the destruction of the erosion remains between the gullies, the latter disappear. Lower gullies sometimes fuse with the hydrographic network and become a part of it.

The development of *underground erosion* is somewhat different. In *intrasoil erosion* the loss is greater where the soil is more coarse-grained, and as a rule represents a further stage of sheet erosion on skeleton and shallow soils. Intrasoil erosion really ceases only after all the fine earth is washed away and the action of moving detritus is replaced by gravitation phenomena.

In *intrasoil wash* or *eluviation* (Latin *eluere* – to wash out), not only the soil texture but also the chemical composition of the soil is important, the accumulation of soluble matter being a key phenomenon. From this point of view mainly saline soils in arid regions are of interest where underground hollows and corridors develop under the influence of an accelerating wash process. The excavation is also accelerated by the creation of deep crevices and a diminishing permeability of peptized layers, and thus still larger quantities of water fill the crevices, and new corridors are created and existing ones expanded.

In a similar way *thermokarst* or *kryokarst* phenomena take place under the influence of the uneven freezing and thawing of the soil, their development being dependent on the water regime of the soil and the properties of the vegetation cover.

In *tunnel erosion* the destructive process starts with intrasoil erosion and underground washing of the binding elements. The second stage is the creation of corridors and tunnels, these becoming larger at the expense of upper layers (ceiling) which grow thinner. This phenomenon occurs mainly in places where water gathering takes place. The last stage begins after the collapse of the ceiling, and thereafter tunnel erosion proceeds according to the principles of gully, and river erosion, respectively. Sometimes suffosis erosion accelerates decerption and then merges with it.

Finally, in *karst erosion* (in this case *pseudokarst erosion*), the stages of intrasoil and tunnel erosion are combined together. A specific feature of underground erosion in a pseudokarst territory, especially in clay karst, is the almost regular appearance of corrasion, which together with mechanical washing creates typical forms.

In general, five stages of development can be distinguished in each form of erosion:

1. the *inception,* or *initial stage,* stage of early youth;
2. the *young,* or *juvenile stage,* stage of youth;
3. the *mature,* or *culmination stage,* stage of maturity;
4. the *late,* or *senile stage,* stage of old age;
5. the *extinction,* or *final stage.*

In some geomorphology papers the first stage is called the *embryonic stage,** and the third phase is often referred to as the *plenioerosive stage* (Latin *plene* – fully). The latter term could be applied to the mature phase in which the whole surface is under attack by erosion. The author recommends that growing, expanding erosion generally be referred to as *aggressive erosion* and diminishing, receding erosion as *degressive erosion* (Latin *degredi* – to descend).

These phases are best observed in gully erosion in which the gully dimensions increase with age and the activity also changes, so that the various phases of erosion are characteristically represented. In gully erosion, as in other forms of erosion, the identification of the erosion phase is of practical importance because correct assessment in this respect makes it possible to forecast further development and select the most effective and most economic control measures.

Perhaps the most detailed study of the various phases of rill development is found in the Russian literature. A lot of information has been collected in this field, beginning with the first decriptions of the phases of erosion by Bolotov (1781), to the most recent publications of a number of authors. Dokuchaev (1877), Kozmenko (1909–1954), and Sobolev (1941, 1948, 1960) and others have made the greatest contributions to our knowledge of gully development.

Of the many classifications in existence only Sobolev's system (1948) is worthy of mention.

Sobolev calls the *first stage* "promoǐny i rytviny" (the phase of *gully development*), which describes small rills from 30 to 50 cm deep originating during high volume surface runoff. The rill has a linear form viewed from above, and the profile is triangular at the outset, later becoming rounded at the bottom. In this stage the development of the gullies is rapid and if on action is taken to inhibit the process it quickly gives rise to large gullies.

The *second stage* is described by Sobolev as the stage of incising of the slope *gully* by retrograde erosion, giving rise to the so-called "vershina" gully head gradient. This is the stage of the maximum growth rate of the gully which then turns into a ravine by retrograd and vertical erosion. A characteristic feature of this process is the tiered bottom, the site of which progressively moves away from the original slope surface. The depth of the *slope* is 2 to 10 m, and the depth of the ravine is between 25 and 30 m or more.

The *third stage* is characterized by the straightening the *erosion curve* leading to a state of balance in which the stages on the bottom of the gully are levelled and the curve becomes typically concave with the *steepest slope* in the upper part and the *gradual slope* in the lower part. The straightening the erosion curve stops, of course, at the bedrock which thus determines its stage.

The *fourth* development *stage* is the phase of restraining (zatukhanie). It begins the moment depth erosion stops and the gullies banks cease to be renewed by

* E.g. the term embryonic cirque has been used.

Fig. 45. Gully which was formed in summer 1954 by a flush from a higher situated pasture (basin of the Hron river, central Slovakia, Czechoslovakia). (Photo D. Zachar.)

undermining. The ravine is widened by meandering and continued modelling of the banks which gradually become stabilized; gully sediments accumulate on the bottom and vegetation covers the soil on the banks. After the end of the fourth phase, the ravine is a so-called "balka" which signifies a stable state with the extinction of all erosion phenomena (the final phase according to the author's view).

Fig. 46. Another view of the same gully (Fig. 45) taken in winter 1955/56. (Photo D. Zachar.)

However, various deviations may occur in nature, particularly during very heavy downpours. Research observations made by the author (Zachar 1970) have shown that in one downpour in 1954, gullies and ravines were created which did not grow further and were soon stabilized; the slopes of the ravine were already breaking up and vanishing two years after their creation and gradually became covered by vegetation (Figs. 45, 46). In another instance, the erosion gully gradually came into being on the lower part of the slope, growing slowly upwards and becoming stabilized in the lower stretches, while the upper parts of the gully were still active.

Finally, gullies are known to be revived after lapses of several years. This revival, or rejuvenation may be connected with major climatic changes or impaired conditions of runoff in the collection area of the gully. Revival, and intermittent occurrence, are well-known phenomena in all kinds of erosion. The long-term course of erosion may best be studied by the analysis of sediments (deposits by water and wind).

In badland erosion the initial stage is gully erosion, the space between gullies (the interlinear or interpluvial space) becoming reduced to less than half of the original total surface area. Gradually the area occupied by gullies increases at the expense of non-eroded land, until finally the banks of the gullies close to form ridges. Gullies may also be widened by lateral and surface erosion so that the area is gradually corroded overall by the joint action of gully and surface erosion (juvenile stage). On slopes with a shallow mantle of detritus over rocky ground, vertical erosion usually stops at this stage and erosion remains are washed away by surface and intrasoil erosion (culmination stage). After the soil mantle or the less resistant rocks have been washed away, the intensity of removal of material diminishes rapidly and rock erosion begins, or the erosion process slows down (senile stage) and stabilizes (final stage).

On slopes consisting of softer rocks the deeper layers are corroded by the aggressive influence of water; gullies develop close to one another and are separated by sharp ridges (earth coulisses, goat backs, etc.). In this (culmination) stage gully piracy occurs, with gullies merging together and as such ramifying further to produce new branches. Lateral gullies develop and other forms of slope destruction occur, the original surface being so extensively corroded that it is no longer possible to discern the initial profile or level. In the third stage the earth mass is consumed by erosion so rapidly that erosion control becomes very difficult, and intervention can only be carried out at great cost. The author's third stage is referred to by Avenard (1965) as the *typical badland stage (bad-lands typiques)*, (Figs. 47, 48).

Only after reaching a state of equilibrium with respect to its activity does the erosion process begin to stabilize, the slopes becoming stable and unevennesses which arose during periods of intensive erosion being levelled out. In this fourth phase vegetation appears on the eroded remains, and in slowing down erosion still further, the plant growth hastens the arrival of the fifth phase of extinction, when the original massif of easily erodible rocks takes the form of rounded humps (elephant back formation) and flat slopes; the cover of vegetation shows a lusher growth at the foot than at the crest of the slope (Kayser 1961).

Because erosion of the badland type is very rapid and since it is mostly only the final stages that are observed, badland is sometimes considered to be the last stage

Fig. 47. Typical aggressive stage of badland erosion in fluvioglacial deposits. This type of erosion was termed destructive erosion by H. H. Bennett (region of Stob, Bulgaria). (Photo D. Zachar.)

in the erosion cycle. Yet the author does not believe that this interpretation has been properly investigated. There are examples from China, New Zealand, Central Asia, the USA and other countries, of badland erosion that has been caused by the recent disturbance of natural systems as a result of interference by man (deforestation, ploughing of the steppe, uncontrolled hydrological effects, etc.).

As to subterranean forms of erosion, a partial study has been made of the development phases of *tunnel erosion,* Hosking (1967) having investigated these forms of erosion in New Zealand. His classification is given in Table 8.

According to the author's scheme for the development stages of erosion processes, the first stage may be considered as being that in which the formation of crevices and vertical and horizontal hollows occurs. The damaged area is insignificant.

In the second stage a system of funnels, well-shaped openings and subterranean corridors develops, their bottoms coinciding with a more resistant stratum (layer). In this stage, surface water mostly enters the underground system and forms visible corridors which grow rapidly. The damaged area is about 5% up to a maximum of 10%.

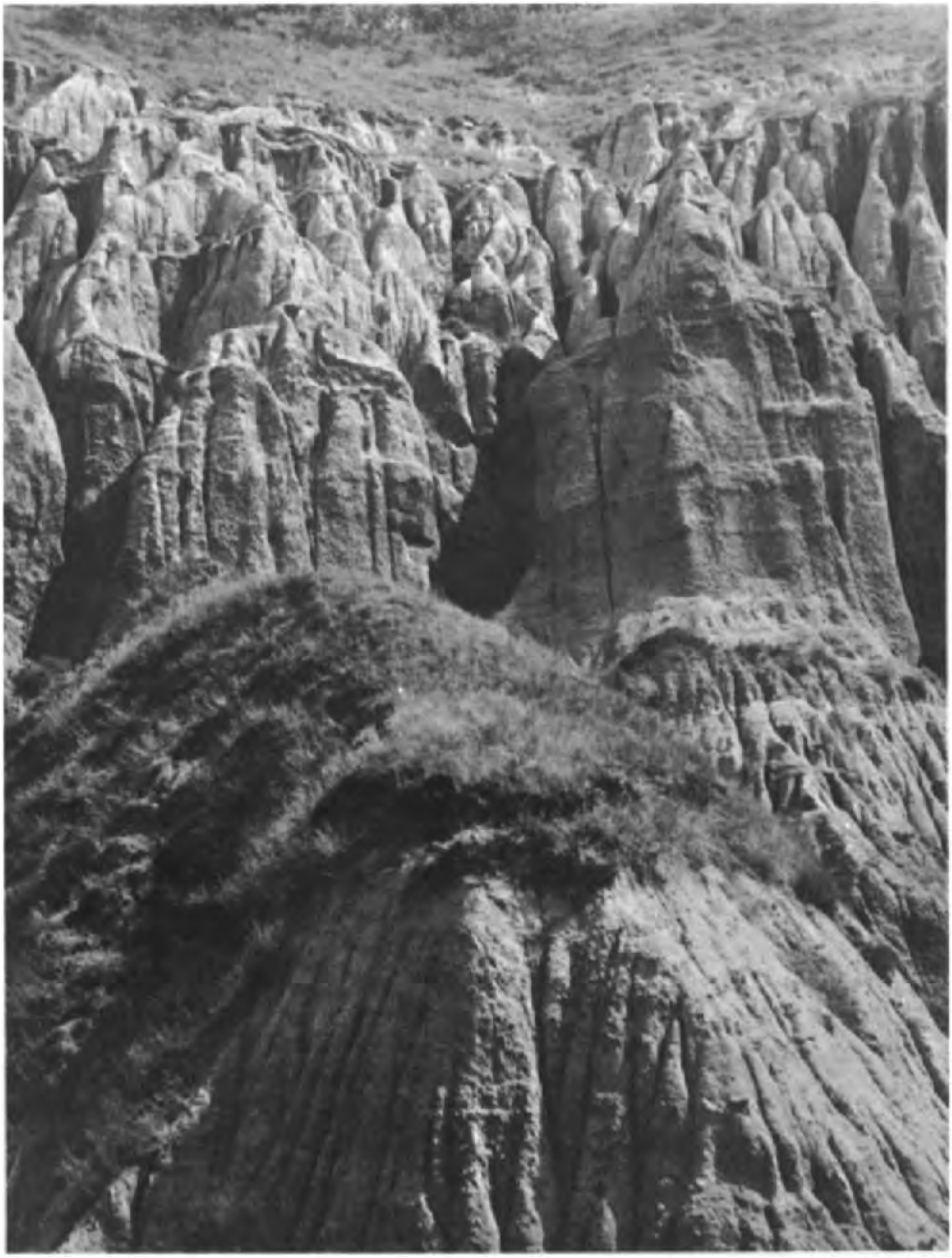


Fig. 48. Another type of badland erosion with commonly occurring pedestal erosion in the culmination stage of development (Red Mountains, Romania). (Photo D. Zachar.)

Table 8. Theoretical stages of tunnel erosion (Hosking 1967)*

Stage		Tunnel characteristics	Tunnel diameter or gully depth	Surface characteristics	Percentage of surface disrupted
Number	Name				
1	Youthful				
a)	Early	None formed		Cracks on surface	—
b)	Mid	Begin forming	Few inches	Cracks larger, silt to surface	—
c)	Late	Continuous, branching	Up to 1′	Unbroken, but tunnel lines visible	—
2	Advanced				
a)	Early	Continuous, uneven profile	1′–2′	Some potholes, 1′-diameter	5
b)	Mid	Enlarging	1′–3′	Potholes larger, more frequent	10–20
c)	Late	Enlarging, but filling from roof collapse	1′–4′	Gully with a few turf bridges	15–40
3	Gully				
a)	Early	Completely collapsed grass-covered floors	Up to 6′	Continuous gullies, unbroken interfluves	20–60
b)	Mid	Gullies enlarge (up to 12′), become shrub-filled or new tunnels form in interfluves			Up to 100
c)	Late	Gullies cover whole sloe, undergoing normal fluvial erosion			100

* Compiled from observations; terminology partly after Downes (3).

In the third stage the growth of subterranean corridors culminates in the collapse of the ceiling in the lower parts of the underground system and earth bridges are created. The area damaged is from 50 to 60%.

In the fourth stage the system of underground corridors is linked to the central gully and surface phenomena gradually prevail over subterranean phenomena which are active only in the upper regions and account for the retrograde growth of erosion gullies. The relief is very rough and difficult to negotiate in this phase.

In the fifth stage the subterranean forms of erosion completely disappear and erosion continues in surface forms. The relief is gradually levelled, the original soil having been completely destroyed.

The various stages can partially be seen in Figs. 34 to 40.

2.5 Classification of eroded soil

The long-term influence of erosion on the soil is to change its properties quantitatively or qualitatively. Very intensive forms of erosion may entirely destroy the soil. Quantitative changes include reduction of the depth of the soil profile and contraction of the soil surface area; qualitative changes involve soil properties and a reduction in fertility. The following part of this chapter deals with the classification of eroded arable land.

2.5.1 Classification of eroded soil on arable land

The classification of eroded soil is as complicated as the classification of soil itself. On the one hand it is desirable to make the classification as detailed as possible and to base it on the genetic features of the soil, while on the other, it is also desirable to keep the classification as simple as possible and to adapt it for practical use. The greatest difficulty is how to compare the eroded soil with the original soil profile, the so-called etalon. The comparability of soil profiles is also made difficult by, among other things, long-term soil cultivation and fertilization which considerably alter the properties of the soil, especially in the surface layers.

A number of diverse classification systems for eroded soil may be found in the literature. A survey of these systems shows that different authors have used different criteria for expressing the state of erosion. The main criteria most frequently used are the colour and contents of the humus, the degree of removal caused by erosion in the various genetic horizons, and the thickness of the remaining soil layer. Many authors (Shaposhnikov 1947, Kozmenko 1948, Lidov 1956, and others) take as the main criterion the percentage of humus removed, this being an important factor from the point of view of soil fertility. For the sake of comparison some of the classifications found in the literature are presented here.

The American literature centres around the classification of Bennett (1939) the basic form of which is widely used with various modifications (Table 9). The first three classes (1−3) relate to the upper layer of the soil which, according to Bennett, comprises the A horizon (topsoil), and the other two classes (4, 5) relate to the B horizon (subsoil).

Sobolev (1939−1960), who devoted several papers to the classification of eroded soil, distinguishes five classes of *washed soil* in his monograph of 1948 (see Table 10). A still more detailed classification is given by Sobolev in his work of 1954, in which he distinguishes only four classes for both groups of soil types: I. weakly washed soil, II. moderately washed soil, III. heavily washed soil, IV. maximally washed soil. Since in the fourth class practically all the soil is washed away, Sobolev does not name it as a soil type. He omits from the classification the

Table 9. Classification of soils eroded by sheet erosion (Bennett 1939)

Grade	Category	Proportion of upper soil layer removed [%]
0	Insignificant erosion	None
1	Slight erosion	0— 25
2	Moderate erosion	25— 75
3	Severe erosion	75—100
4	Very severe erosion	Whole upper layer removed
5	Exceptionally severe erosion	Over 75 % of lower soil layer removed
6	Other erosion phenomena	Landslides, solifluction, etc.

Table 10. Nomenclature of eroded soils. Initial stages of erosion development on arable land (Sobolev 1948)

Grade of soil erodedness	Chernozem and similar soil types (leached chernozem, etc.)	Podzol and similar soil types (gray forest soil, solide, etc.)
1	Soil surface with wash marks in the form of 5—10 cm deep rills; not more than half of the humus horizon washed away	Soil surface with wash marks in the form of 5—10 cm deep rills; humus horizon totally or partially washed away; ploughing disturbs the podzol horizon
2	At least half and up to the whole of the humus horizon washed away; transition horizon cut through by plough	The podzol or saline horizon is partially or totally washed away; part of illuvial horizon disturbed by ploughing
3	Transition horizon partially washed away	The illuvial horizon is partially washed away
4	All soil horizons totally washed away	
5	All loose parts of the weathered mantle (fine earth) washed away down to the bedrock	

Table 11. Grading of soil erosion according to Kohnke and Bertrand (1959)

Grade	Nomenclature and characteristics
+	Recent sediments
1	Nil or slight erosion; 0—25 % of upper soil layer washed away
2	Moderate erosion; 25—75 % of upper soil layer washed away
3	Severe erosion; over 75 % of upper soil layer washed away
4	Very severe erosion; most of the upper layer and part of the lower soil layer washed away; gully development starting

Table 12. Grading of medium deep and undeveloped soils by degree of erodedness arising from sheet erosion

Grade	Soil erodedness	Proportion of soil removed (% of the original thickness of the soil profile)
1	Slight	0— 20
2	Moderate	20— 40
3	Severe	40— 60
4	Very severe	60— 80
5	Soil completely eroded	80—100

fifth category, in which the topsoil is deposited in the loose, lowest layers of soil. In the author's opinion, this class is justified in some cases.

Of the many other systems, the classification of Kohnke and Bertrand (1959) from the American literature is worthy of mention (Table 11). As can be seen, the classification is based largely on the topsoil layer and resembles Bennett's classification. Other soil damage is omitted, but sediments are included so that the subsoil is not actually classified in terms of erosion. American methods of classifying eroded soil differ from Soviet methods in that they attach greatest importance to the topsoil A horizon, which contains the largest stock of nutrients and organic matter.

Taking into account the advantages and disadvantages of the various classification systems, the author considers it practical to divide eroded soil into five classes, giving due regard to the stratigraphy of the genetic horizons of various soil types. Basically the entire soil profile, including the C horizon, is divided into five parts for the purposes of classification, the thickness of the layers of the various strata being given by the thickness of the respective horizon. Such a classification is simple and also corresponds to the qualitative differences between the various soil horizons.

On moderately deep soil and mixed slope soil the relative thickness of the layer removed from the soil profile, including the loose part of the C horizon, is critical for making an assessment of the state of erosion. The mechanical participation of the soil lost in the various classes is estimated according to the principles of the previous classification (Table 12).

It is supposed that such a classification also reflects the reduction of soil fertility that occurs in the various classes. Initially, fertility decreases largely as a result of the removal of the richer layers, and later as a result of the soil profile becoming shallower. In the fifth class local denudation of the bedrock occurs so that only a residue of the original soil mantle remains (Fig. 49).

Fig. 49. Rendzina soil, originally tilled land, now completely eroded in the upper part and severely eroded in the lower part of the slope (Slovak Kars, Czechoslovakia). (Photo D. Zachar.)

2.5.2 Classification of eroded soil on pastures

Erosion on pastures develops differently from erosion on arable land where the erosion features are obliterated by cultivation. This difference in development arises from two factors, the first being the low vegetation protecting the soil on pastures, the second being the physical effect of animals on the soil. Erosion does not appear in distinct forms if overgrazing is avoided and if the grass cover is well maintained with orderly grazing management. With heavier grazing, the variety of species in the vegetation is reduced, the stand becomes thinner and lower, and its soil-protecting effect vanishes.

In places where cattle are frequently on the move, *trodden down tracks* or *paths* come into being. In this way a slope may become terraced, giving rise on the one hand to an unprotected level area with heavily disturbed soil, and on the other, to an abrupt bank where soil does not remain in place because of the steepness, the damage caused to the plant cover, and mechanical disturbance by passing animals. The steeper the slope, the more frequently the livestock is driven across, and the larger the grazing stock and the weight of individual animals, the more intensive is the erosion process. The pathways, which initially are few with a width of 40 to 50 cm, eventually increase in number, the strips between them gradually disintegrating into small raised areas of low growth. At a further stage of development these

Table 13. Grading of eroded three-phase soils on pastures

Grade	Soil erodedness	Description of the eroded profile of three-phase soil
1	Slight	Paths become conspicuous over the ground, vegetation becomes sparce, soil is washed, and part of the A horizon is removed (up to 50% of its thickness and 25% of its area)
2	Moderate	The proportion of land from which soil has been removed is the predominant part, more than 50% of the A horizon (20 to 40% of the total soil thickness removed), and up to 50% of the land surface is denuded
3	Severe	Denuded soil prevails (up to 75%), the whole A horizon (40 to 60% of the total thickness of the soil profile) is removed
4	Very severe	Denuded soil prevails (up to 75%), the A + B horizons (60 to 80% of the total thickness of the soil profile), are removed, and loose substrata are eroded
5	Soil completely eroded	The soil is denuded over 75% of the area, almost all loose soil is removed, the bedrock is becoming exposed

Fig. 50. Completely eroded soil of what was a pasture. Erosion was originally started by deforestation; later the protective effect of the vegetation was reduced by over-grazing and the mechanical compaction of the soil by cattle (Tematín Hills, Czechoslovakia). (Photo D. Zachar.)

areas are also eroded; as soon as the vegetation is destroyed and the soil denuded, erosion progresses rapidly, breaking up and dissecting the microrelief still more.

Little attention has been paid so far to the classification of eroded pastures. The only classification of soil erosion on pastures that has been proposed is that of Sobolev (1948), in which by analogy with arable land eroded grazing land is classified into five categories with the first of these divided into four subclasses. The detailed classification of the first category is based on the degree of surface damage to the subsoil horizont. Following from Sobolev's classification and the author's own experience, grazing land damaged by erosion may be classified according to the recommendations, given in Table 13.

In assessing the degree to which the soil has been affected by erosion, the reduction in the thickness of the soil horizon is a key factor, whereas the percentage denudation indicates the fraction of the surface from which soil has been removed. The proportion of the surface covered by vegetation may change very quickly in different phases of erosion. A permanent, or long-lasting reduction of the vegetation cover occurs only in the last phase, when plant life cannot find favourable conditions for growth (Fig. 50).

2.5.3 Classification of eroded forest soil

There is a general opinion that forestation constitutes the most perfect means of soil control, being an effective stabilizing agent, and bringing success where other measures have failed. Despite this, erosion phenomena causing damage to the soil may occur on forested land, such damage being caused by logging, skidding and hauling of timber, by grazing, and also by water (Fig. 51) and wind. In some cases the erosion may be linked with solifluction and landslides, etc. Finally, soil may be damaged by wind-blows. The literature does not contain any account of the various kinds and forms of damage that occur in forest soils, nor unfortunately does it offer the necessary information for building a classification of eroded forest land, and therefore only a few remarks on this subject will be made in this chapter.

From among the erosion phenomena already mentioned, *logging erosion (exploitation erosion)* is considered to be the most typical occurring on forest soil, and is one form of man-made (anthropogenic) erosion. It involves the gouging and scraping of the soil or bedrock when timber is hauled over the ground and when heavy vehicles are in use (Figs. 52, 53).

Information on logging erosion (in Russian the term *expluatatsionnaya éroziya* is used) from the Transcarpathian beech forests, is given by Polyakov (1962), who established that after clear-felling 15 to 75% of the soil surface was eroded. Damage from logging erosion over the investigated area (which had an inclination of 25 to 28°) amounted to 128 m^3 ha^{-1} caused by gravitational skidding, 104 m^3 ha^{-1} caused by tractor skidding, 92 m^3 ha^{-1} by animal skidding and 21 m^3 ha^{-1} as

Fig. 51. Forest soil severely damaged by erosion after clearfelling (West Carpathians, Czechoslovakia). (Photo D. Zachar.)

a result of cable-way skidding. As the size of the damaged surface area and the degree of mechanical disturbance of the soil increased, so the extent of precipitation erosion also increased, attaining 238 m^3 ha^{-1} in areas damaged by gravitational skidding, 180 m^3 ha^{-1} in areas damaged by tractor skidding, and 24 m^3 ha^{-1} resulting from cable-way skidding. In total, from 43 to 336 m^3 of soil per ha were removed from the felled area.

A similar study by Popov and Dekatov (1956) showed that timber felling and skidding on a 17–19° inclination resulted in soil disturbance over 62–96% of the surface area and the amount of eroded soil amounted to 139–596 m^3 ha^{-1}. This research refers to several regions in the Northern Caucasus.

Although these losses associated with logging occur once in the rotation period, improperly managed operations can have severe consequences and make afforestation more difficult. Therefore the author recommends that soils of clear-felled areas damaged by this kind of erosion be classified according to the criteria given in Table 12. If rills are caused by skidding and hauling, the author would suggest

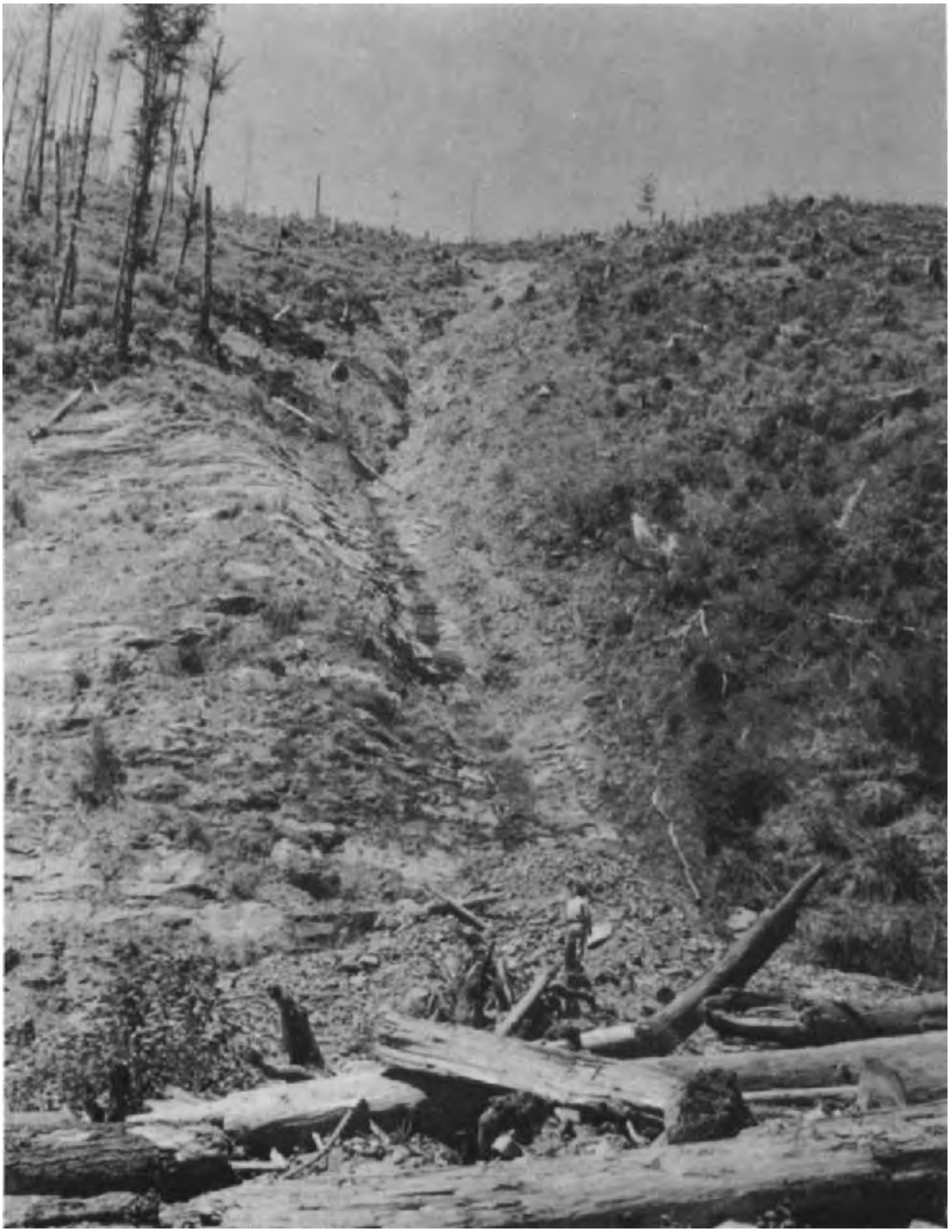


Fig. 52. Erosion resulting from the dragging of logs (Coast Range, Oregon, USA). (Collection of photographs of the College of Forestry and Forest Industries in Zvolen, Czechoslovakia.)

Fig. 53. Erosion resulting from the clearfelling of a Douglas Fir stand and from road construction (Alsea, Oregon, USA). (Collection of photographs of the College of Forestry and Forest Industries in Zvolen, Czechoslovakia.)

making use of the criteria in Table 4. The same recommendation is made where damage caused by landslides and solifluction occurs in forested areas.

A special phenomenon is that of *wind throw erosion* which is caused by trees blown down by the wind. Mizerov (1966) describes this type of erosion under the name of *vetrovatel'naya éroziya.* The effect is to create hollows with a diameter of 1.5 to 2 m and a depth of 1.5 m, the area of individual hollows varying from 2 to 8 m^2. This creates a special kind of microrelief making for more varied ecological conditions on forest land. On the one hand the burial and inversion of the soil profile takes place, and on the other, the less valuable substratum, or even the bedrock, is denuded. Often the hillocks have a tendency to slide laterally, thus dissecting the microrelief still more. On wet sites there tends to be a high water level in the wind throw hollows, particularly in spring or during rainy periods, and this makes afforestation difficult or even impossible. In localities exposed to the wind, the wind throw of trees occurs frequently, so that whole soil mantle disintegrates into a series of protuberances and hollows, and the soil looses its most valauble properties.

According to the proportion of soil affected by wind throw erosion, the first category in the scale of damage may be considered as being up to 20%, the second from 20 to 40%, the third from 40 to 60%, and the fourth from 60 to 80% of the cleared area. The highest category can be considered as that in which over 80% of the surface area is damaged, with other harmful effects also being in evidence.

2.5.4 Classification of soil damaged by wind erosion

As in precipitation-mediated surface erosion, the soil may also be damaged by wind erosion involving the deflation of finer soil particles and the corrasion of coarser material. Of these two forms only deflation is of a malignant character, and since it acts very selectively, a pronounced depression of the soil profile occurs only on fine-grained soil. On a skeleton soil, the finer particles having been completely removed, a protective coarse-grained layer consisting of drift sediments develops on the surface, thus preventing further blow off. Deflation is, therefore, most destructive on sandy soil with a high proportion of third and second fractions of fine earth.

From among the many classifications of eroded soil which exist, again those of Bennett (1939), Sobolev (1948), and also that of Gael' and Smirnova (1965) are most worthy of mention.

Table 14. Classification of soils eroded by wind and soils buried by wind deposits, according to Bennett (1939)

Grade	Erodedness of soil	Percentage of upper soil layer removed
1	No visible erosion	—
2	Slight erosion	0— 25
3	Moderate erosion	25— 75
4	Severe erosion	75—100
5	Very severe erosion	The whole layer including 25—75% of the B horizon
6	Exceptionally severe erosion	The whole layer as well as 75% or more of the B horizon

Grade	Accumulation	Thickness of layers [cm]
1	Slight	0— 15
2	Moderate (uniform)	15— 30
3	Moderate (undulating)	15— 30
4	Deep	30— 90
5	Sand drifts (small dunes)	90—180
6	Sand drifts (large dunes)	> 180

Bennett classifies soil eroded by wind into various groups according to the degree of blow off and the degree of deposit or soil burial (see Table 14).

Sobolev (1961) classifies soil damaged by deflation into four groups, and buried or covered soil into three groups (Table 15).

Table 15. Classification of soil eroded by wind and soils buried by wind deposits, according to Sobolev (1961)

Grade	Erodedness of soil	Wind removal of soil
1	Slight	Up to half of the A_1 horizon
2	Moderate	Up to the B_1 horizon
3	Severe	Up to half of the B_1 horizon
4	Very severe	Up to the C horizon
Grade	**Soil burial**	**Thickness of aeolian deposit [cm]**
1	Slight	<20
2	Mderate	20—40
3	Deep	>40

Table 16. Classification of aeolian soils according to Gael' and Smirnova (1965)

I Sand content of topsoil	Clay content lost [%]
1 Low (moderately light colouring)	<25
2 High (light colouring, sand drifts)	35—50
II Erodedness of soil	**Thickness of genetic horizons blown away**
1 Slight	Up to half of the A horizon
2 Moderate	Whole A horizon
3 Severe	Up to half of the B_1 horizon
4 Very severe	Up to the B_2 horizon
5 Exceptionally severe	Up to the C horizon
III Degree of burial	**Thickness of aeolian deposit [cm]**
1 Slight	<12
2 Shallow	12— 25
3 Medium-shallow	25— 50
4 Medium-deep (small dunes)	50—100
5 Deep (small-medium dunes)	100—300
6 Very deep (medium dunes)	> 300

Gael' and Smirnova (1965), having made a study of eroded soil in dry regions of the Soviet Union, recognized the following classes: I. Topsoil with an increasing sand content (opeschanennost' pakhotnogo sloya), II. Soil damaged by erosion (erodirovannost'), and III. Soil buried (pogrebennost') by drift material (Table 16).

In the case of wind-damaged soils in temperate regions the author recommends using the same classification as that used for soil eroded by sheet erosion. In terms of the three proposed soil horizons, the first category includes situations in which up to 50% of the A horizon has been blown away, in the second category the entire A horizon is missing, in the third category 50% of the B horizon is eroded, and in the fourth category the bedrock is attacked. The classification is similar to that of Gael' and Smirnova (1965) with the exception of the third category in which these authors include only half of the B_1 horizon. The burying of soil by deposits is discussed in the chapter dealing with sediment classification.

2.6 Classification of erosion remains

In the last stages of soil erosion, detritus occurs only in the form of relics and remains which may differ in form and character. The remains are interesting both with regard to form and the genesis of erosion phenomena, the study of which may be based on erosion remains. The author recommends that erosion remains be classified as *relics, rudiments, selective remains,* and *inhibited remains.* Although it may sometimes be difficult to determine the boundaries between these categories, the author is nevertheless of the opinion that this classification, at least partially, expresses their character and origin.

The first group – *relics* (Latin *relinquere* – to leave behind) – occurs on sites where the soil, because of favourable natural conditions such as a gentle incline, natural vegetation, etc., has not so far been affected by erosion and has therefore kept its original properties. Soil relics are particularly important for comparing the properties of the eroded soil with those of the original soil (i.e. for selecting the so-called etalon by which erosion on neighbouring land is judged), and therefore the soil profile is intact in this kind of remnant. Erosion relics are preserved in the form of tables, strips, ridges, columns, and other forms (Fig. 54).

By progressive corrosion of the soil the surface area and volume of erosion remains decrease and their shape becomes constant. Finally, only *rudiments* (Latin *rudimentum* – beginning) of the original soil cover remain and these bear little resemblance to the original state of the soil, either in a horizontal or a vertical direction. In sheet erosion and deflation, soil in the last stages of the erosion process may be considered as a soil rudiment, the upper horizons which are charasteristic of the soil profile already having been removed, with the loose bottom horizon currently being attacked by erosion (Figs. 42, 49). In gully erosion

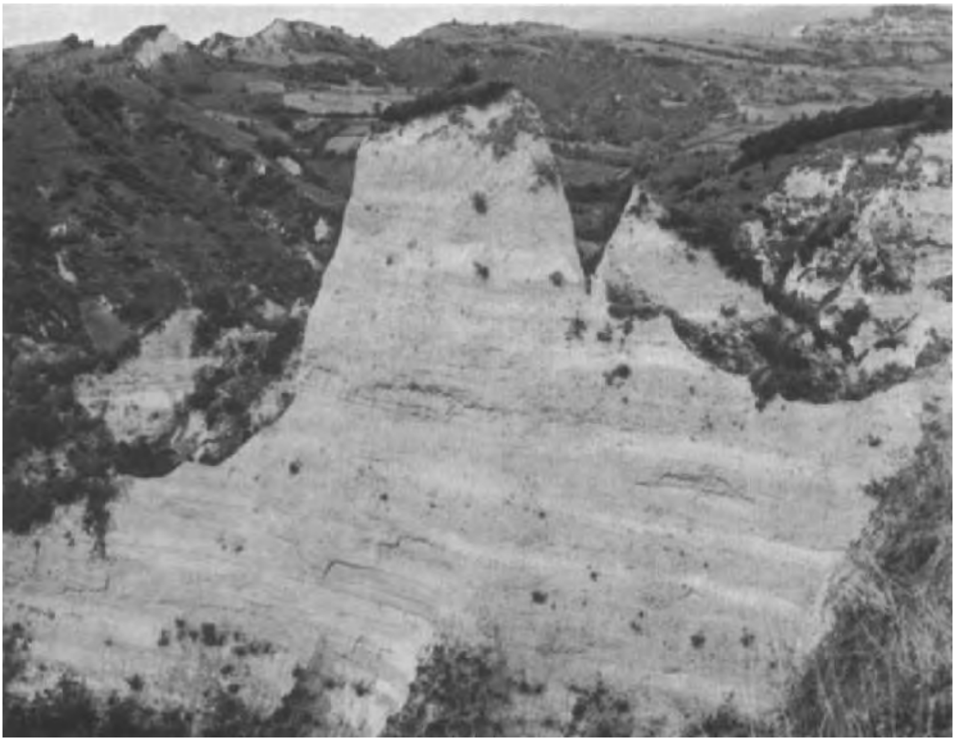


Fig. 54. Towering erosion remnants indicating the height of the original relief, the intensity of erosion, and the properties of the original soils (sandstone rocks near Melnik, Bulgaria). (Photo D. Zachar.)

the soil is preserved in the form of *ridges, ribs, coulisses,* etc. (Fig. 55). In badland erosion these forms develop into *earth coulisses,* usually joined into a system of erosion remains, the main ridge, coulisses, and lateral ribs forming the skeleton (Figs. 31, 32, 47). Under the influence of sheet and gully erosion the rudiments of the soil are modelled, levelled and rounded into various *hump backs* (such as the *elephant backs* already mentioned), *tongues, hollows, loaves,* etc.

Sometimes the summit of an erosion remnant may be protected by vegetation, a stone, or some less erodible component, and thus a plate, or table, etc. may be formed. In such cases the lower parts of the formation are usually eroded uniformly on all sides so that the remnant takes the form of a truncated cone. Interesting forms arise where erosion has a stronger effect near the ground, thus producing shapes which are narrower at the foot. Of these formations, *earth pyramids* (Fig. 56), *pillars, columns, needles, trunks, mushrooms, perched boulders,* etc., are best known. In rain erosion and deflation so-called *micropyramids* are created. Elongated forms with remnants of grass turf may be referred to as camel backs; these forms arise both from water and wind erosion.

Fig. 55. Soil remains after erosion. a – in a neovolcanic region, and b – in a limestone region (Štiavnica Mountains, Slovak Kars, Czechoslovakia). (Photo D. Zachar.)

Fig. 56. Earth pyramids formed in fluvioglacial sediments (South Tyrol, Italy). (Photo D. Zachar.)

Thus erosion takes a course that is not uniform, but rather selective, according to the resistance of the various components. The author therefore recommends that the type of erosion remnants which have arisen from predominantly selective erosion be referred to as *selective remains,* these being found to consist more frequently of rock material than soil. In a broader sense a *coarse-grained* or *stony layer* of drift sediment on the soil surface is also a selective erosion remnant. Such layers take the form of *rock pavements, stone fields,* or *rock debris (talus),* etc. A common feature of selective remains is the creation of a coarse-grained layer on the soil surface, or in the topsoil.

These erosion remains are the origin of *residual erosion remains* which arise mostly as a result of underground erosion, the erodible components being washed away and the resistant components remaining in the soil to form the skeleton of an existing remnant. The erosion processes that are involved here are *intrasoil lixiviation, leaching, decalcification, desalination,* etc. – processes by which *residual soils* and *sedentary soils* come into being. In the pedological literature such soils are also referred to as *impoverished soils,* or *eluvial soils.* Besides *pluvial residua* there are *aeolian residua* such as the *rock pavements* and other formations already described. By the action of tunnel erosion, *hollow soils, excavated soils,* or *undermined soils* arise. This process may be generally referred to as the *tunelling,* or *suffosis* of *soil.*

The last to be considered are inhibited erosion remains which have survived thanks to artificial protection measures, while erosion continues on unprotected soil.

2.7 Classification of sediments

2.7.1 General

Any erosion activity or disturbance is followed by the *transport* of eroded particles and *sedimentation* and *accumulation* of the transported material. Because these phenomena are complementary, a few remarks will be in order on the third and last phase of the planation process which is considered by geologists to be a *constructive activity,* since the territory receives new material, is upgraded, and new soil is formed as a result. It should be added that from the pedological standpoint sedimentation is a creative activity provided that it proceeds slowly with sediments being deposited of higher quality than the original soil of the area concerned. Otherwise sedimentation activity has a negative influence.

Sedimentation results from the transport of bed load and its deposition at another site, there usually being intensive selection of material during this process. Just as the revival of erosion and new soil erosion may take place (Figs. 14, 15, 57), so also new deposits of sediments may occur, usually in connection with a change of the erosion factor and the sedimentation environment.

With regard to sedimentation caused by human intervention, we can distinguish (as in the case of erosion) between *natural* and *man-made (anthropogenic) sedimentation.* In the same way, *harmless (benignant)* and *harmful (malignant) sedimentation* may be distinguished. Damage caused by sedimentation includes the rapid burial of crops and the deposition of qualitatively poor material on good land (Fig. 58).

An important characteristic of all sediments is their stratification, which is based on the *grading of sediments.*

After colonization of virgin territory by vegetation a mantle of detritus with fine-grained fractions gradually develops, and in the "normal" process of erosion

Fig. 57. Erosion forms on the surface of aeolian sands on the shores of the Baltic Sea (Poland). (Photo D. Zachar.)

on soil covered by natural vegetation (natural erosion), this results in a *positive gradation of sediments,* the material being increasingly finer towards the surface. The finer the eroded material and the weaker the kinetic force involved in the erosion process, the higher is the quality of the sediments (in terms of size of particles), and vice versa.

As a result of human acceleration of the erosion process, by changes in the climate or other conditions, erosion may be revived and a *negative gradation of sediments* may appear, as is found in the textural structure of eroded soils. If the supposition is made that the gradation of detritus in uneroded eluvial soil is initially positive, we find that by selective erosion from above downwards, increasingly coarser material is uncovered and eroded, thus producing a negative gradation of sediments. Harmful erosion is usually associated with a negative gradation, and harmless erosion with a positive gradation of sediments (Figs. 58, 59).

This analysis of the origins of gradation is far from complete. By the selective action of the erosion process negative strata of sediments come into existence on the soil surface, and this may impose limitations on the erosion process and bring

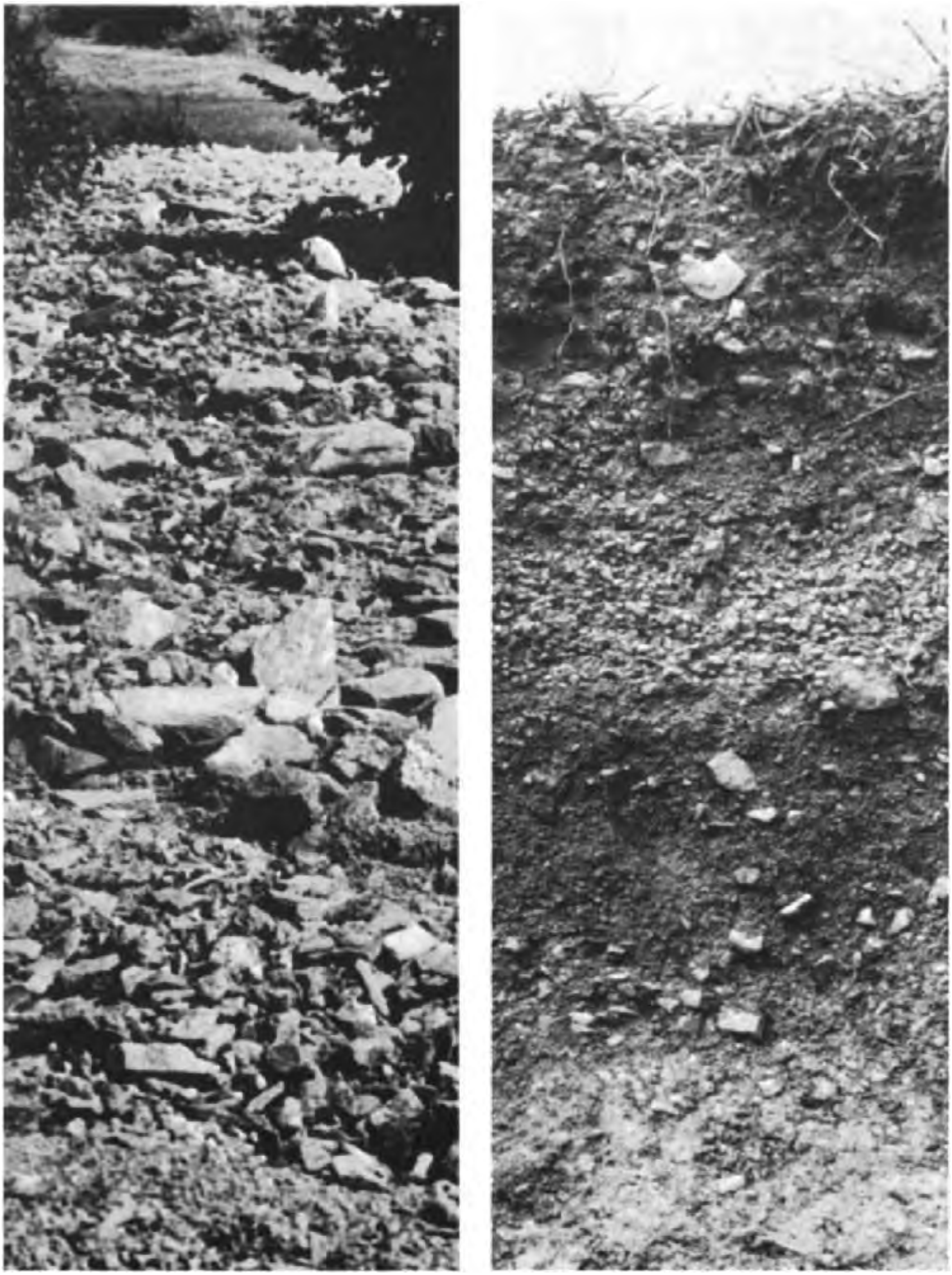


Fig. 58. Harmful sediments resulting from accelerated pluvial erosion. a – deposit cone of the erosion gully, b – horizon buried by coarse-grained deposits (Low Tatras, Slovak Kars, Czechoslovakia). (Photo D. Zachar.)

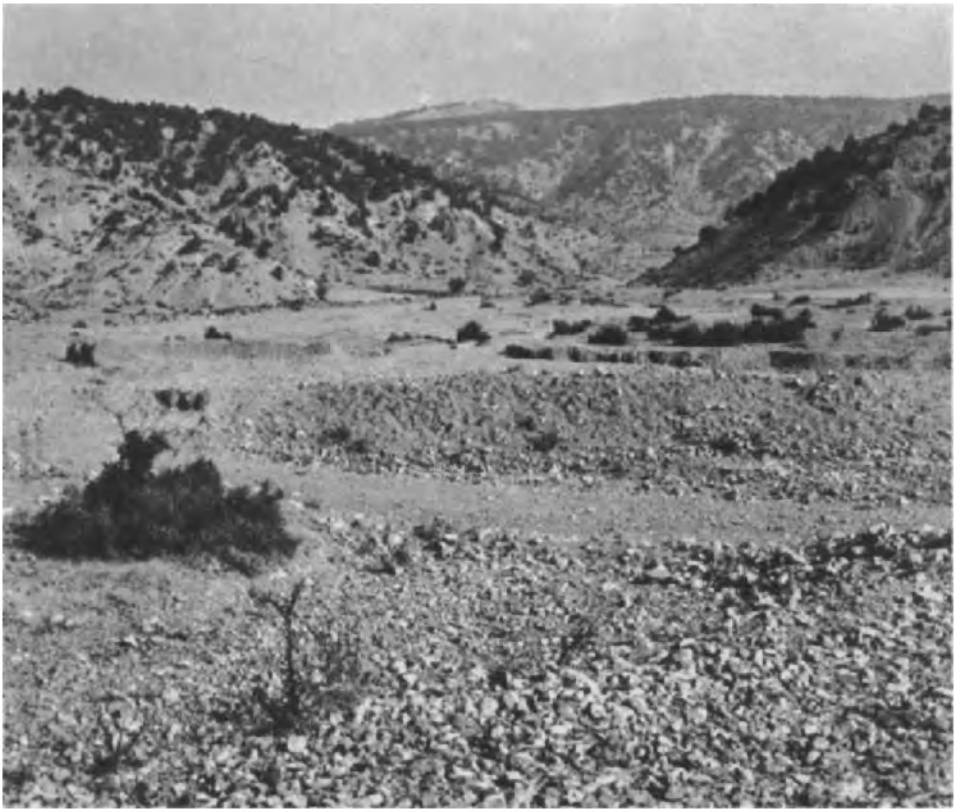


Fig. 59. Flood deposits in the semiarid region of Anatolia which cannot be cultivated unless flooding is controlled (Turkey). (Photo D. Zachar.)

about a compensatory effect on the gradation of sediments. By analyzing the particle structure of the strata it is relatively easy to determine the intensity and development of erosion processes.

It may be said in general that the origin and type of the eroded material, the type of erosion process, the nature of the sedimentation environment, the intensity of the erosion process and other factors, provide the sediments with particular properties from which it is possible to reconstruct, at least to some degree, the characteristics of the original erosion and accumulation processes. Because sedimentation influences the fertility of the soil and its utilization, due attention should be given to it. Since the properties of sediments depend mainly on the type of erosion process, a few observations relating to the latter will be made.

2.7.2 Classification of sediments according to the erosion factor

The process of deposition of sediments varies according to the type of erosion activity involved. Unfortunately, there is no uniform nomenclature for sediments. The authors's classification is based on the terminology used in the classification of erosion phenomena.

In general, sediments may be classified according to whether they arise from *gravitation* processes, decerption, solifluction, etc. Erosion sediments may be further classified as *aquatic, niveal, glacial, aeolian, organogenic,* and *anthropogenic sediments.* Aquatic sediments, which are of widely different origin, may be subdivided again into pluvial, limnic or lacustrine, and marine sediments.

Pluvial or *precipitation sediments,* being the most important, arise from the deposition of material that has been loosened by precipitation erosion. The oldest expression referring to these sediments is "deluvial sediments", in short *deluvium* (Latin *deluere* – to wash away), which, according to Pavlov (1894), comes into existence by the deposition of material transported by slope flow and builds up either on the slope, or at the foot of the slope. Later the term deluvium or diluvium was used to include all slope detritus moving under any sort of influence, including that of atmospheric water. Żolciński (1929) understood *deluvial processes* as referring mostly to wash. Because this material is genetically heterogeneous, slope and slope foot sediments are also sometimes named *colluvium* (Latin *colluere* – to wash) (Kettner 1954). This term, in the author's opinion, is an adequate expression for sediments of different origin. The terms deluvium (diluvium) and colluvium are now considered as being synonymous.

If a distinction needs to be made between the terms deluvium and colluvium, the author recommends the terms *slope* and *subslope deluvium* to refer to sediments deposited on a slope or at the foot of a slope by wash water, reserving the term colluvium for sediments deposited in sunken areas and depressions, such as hollows, dales, and dells. The term colluvium would imply that the material is heterogeneous and that it had been transported to depressions in the terrain by the confluence of slope flows. Typical colluvia in this sense occur in dry regions where slope configuration is poorly developed or there is no overall discernable slope.

Where it is necessary to emphasize the precipitation origin of sediments, the author recommends that preference be given to the terms pluvial, or *precipitation sediments.* These may be *sheet* (slope or subslope) sediments or *gully sediments* (on gully bottoms) (Fig. 58). *Cone-shaped sediments* constitute a special type which accumulates both at the mouths of gullies and at those sites in rivers where the bed levels out at the foot of a slope. These sediments, or rather alluvia, are also termed *outwash fans.* They may be divided into *precipitation (pluvial), torrential,* and *fluvial fans,* etc.

In mountain regions there occurs a special group known as *proluvial sediments,* or *proluvia,* denoting a complex of clastic sediments accumulating at the foot of mountains. According to Sekyra (in Svoboda 1961) such clastic material is carried down mainly during downpours. Proluvial sediments are only slightly sorted, if at all, and may be of different origin (pluvial, fluvial, or arising from decerption, etc.).

Fluvial alluvia develop in river-beds and floodplains, and thus bed or *alveolar sediments* (Latin *alveus* – cavity, hollow), and *channel* or *alluvial sediments* (Latin *alluere* – to wash) are distinguished. For soil formation, there are the sediments arising from river-bed deposits during inundations that are most important. Buckland, the author of the term alluvium, used the latter to denote recent fluvial sediments. Later the use of the term was unjustifiably changed in favour of the designation of younger quaternary, i.e. Holocene sediments. Inundation alluvia may vary greatly in nature; *torrential sediments* are particularly dangerous and harmful (Fig. 59).

Various sediments are typified by their particle composition and stratigraphy. Whereas in deluvial sediments the stratification is *parallel,* in outwash fans it is conical, in proluvia it is *irregular,* and in fluvial sediments it is mostly *diagonal.* The first sediments at the apical region of the cone are coarse-grained, later the texture becomes finer, ending in clay fractions. In purely gravitational phenomena the granular composition of sediments is reversed; first the lightest particles are deposited and the base is formed by detritus of the largest gravitational momentum.

In *lacustrine (limnic)* and *marine (maritime) sediments,* delta, shore, and bottom sediments are distinguished. Sediments which have been brought by river waters from far inland are called *terrigenous sediments* (Latin *terra* – earth). As a result of the build-up of these deposits the land surface grows outwards into the delta region of slow-moving water at the expense of the sea. The heavier the erosion in the catchment area and the shallower the delta region, the more intense is the aggradation action. Delta alluvia contain the finest sediments, rich in nutrients, but their economic utilization depends on effective control and drainage of underground waters. Rapid accumulation usually causes waterlogging of the floodplain and the delta. In general, water erosion is linked in eroded parts of relief to dryness, in inundated parts to a surplus of water.

No less interesting are *wind (aeolian) sediments* which are formed by the deposition of particles carried by the wind. As in water-mediated erosion, particles may be transported over short or long distances. Coarser, especially sandy fractions tend to move by a rolling or jumping action, filling holes in the ground to begin with, then moving over greater distances on the even surface and *creating various forms of drift.* Finer dust particles soar upward (efflation) and are thus transported over considerable distances before they settle in quieter conditions as *loess sediments.* The best known aeolian deposits are *sand* or *wind ripples (ripple marks), barkhans* or *horse-shoe dunes (crescent dunes),* and *sand dunes* or *sand mounds,* which are the largest formations with their long axis oriented crosswise to the

Fig. 60. Sand dunes in the desert of Iraq. The surfaces of the dunes are sculptured with typical ripples. (Photo M. Martinů.)

prevailing wind (Figs. 12, 57, 60, 61). Besides these, dunes of various other forms may be formed, both simple and complex in shape. It is a property of aeolian sediments that they are asymmetrical in cross-section, the inclination being gentle on the *windward side,* and steep on the *leeside.*

Wind sediments are also formed by concurrent soil erosion which leads to the formation of various *erosion-accumulation structures* not so frequently observed in precipitation erosion. By the action of the wind soil is deprived of small and light-weight particles which are carried away over some distance. Damage is also caused by the accumulation of sand sediments which bury the more fertile soil horizon. Most of the aeolian sediments in Czechoslovakia occur in fossil form, although active wind erosion is widely distributed and seems to be on the increase. Affected areas include, on the one hand, very poor sandy formations in the lower reaches of rivers, sea and lake shores, desert regions, and on the other, very rich loess formations upon which one of the most fertile types of soil is formed. A large part of the material in aeolian sediments has already been sorted in an aqueous environment, and other sediments (loess clay) have subsequently been partly washed and deprived of some important components – mainly $CaCO_3$ and other salts.

Fig. 61. Aeolian forms in Sahara Desert. (Photo D. Zachar.)

Passarge (1929) divides aeolian soils into two main groups: 1. *eroded aeolian soil (Rückstandboden),* and 2. *deposited aeolian soil (Ablagerungsboden).* A typical mark of a primary eroded soil is the formation of a *stone pavement (Steinpflaster).* As a result of the blowing away of the finer fractions, loose sand with characteristic drift form can arise from sandy sediments, sandstone detritus, loam, sandy marl, clay, or volcanic rock containing fine ashes. In regions of high sedimentation it is mostly fine-grained sediments that are formed, these having a completely homogeneous granular composition.

According to the quantity of *aeolian deposits* added to the soil, purely *aeolian soils, mixed aeolian soils,* and *soils with an aeolian admixture* are distinguished. *Sand drifts* and *loess* are the most typical of the first of these groups. The second group comprises soils which are also purely aeolian but of mixed origin, and the third group includes the majority of all other soils, since soil dust is carried by the wind to practically all regions of the Earth. The overall influence of wind on the soil can be summed up under the general term of *soil aeolization.*

Hellmann and Meinardus (1901) have computed that during the last 100 years 4.78 mm (or 700 kg ha^{-1} $year^{-1}$) of Sahara dust settled in southern and central

Europe, and this could be of great importance as a soil-forming factor. Leuchs (1932) points out that in 1928 about 1,130,000 tons of dust originating from the Central Asian steppes fell over 970,000 km^2 of land in Poland. The largest amount of dust mentioned in the literature was produced by the Katmay volcano from which 28.7 million tons of dust was thrown out in a few days.

Glacial, or *glaciofluvial sediments* form a special group of sediments which are mostly deposited in the form of various *moraines* (e.g. *frontal, lateral, ground moraines*), usually consisting of unsorted and unstratified material. If not sorted by water, they contain the entire range of fractions from clay to boulders. Because their cohesion is poor and they are often built up in thick layers, they easily succomb to erosion. A characteristic feature of these formations is the existence of erratic boulders which were left behind on the glacier's path as allochthonous material after its retreat.

Snow deposits occur mostly on the lower margins of snow fields; they consist of unsorted and relatively good quality material derived from the uppermost soil layers which are the richest in nutrients. However, the finest particles together with the nutrients tend to be washed out of these formations relatively rapidly by subnival erosion.

Of other types of sediment, only *anthropogenic sediments* with their multivarious forms need to be mentioned; these include *dumps, heaps, banks,* etc. In most instances their quality is poorer than that of the buried soil.

2.7.3 Classification of buried soil

From the large range of soils formed from sediments, a few remarks need to be made concerning the classification of soils formed by pluvial sedimentation, this being the most common of recent accumulation phenomena and the most important from the erosion point of view.

A classification of soils *buried by pluvial sediments* was attempted by Kostyuchenko (1937) who recognized: 1. small (malonamitie), 2. medium (sredne namitie), and 3. heavy (silnonamitie) deposits. In the first category the thickness of the deposited layer is 10 to 20%, in the second 20 to 40%, and in the third over 40% of the thickness of the humus horizon of the buried profile. In further schemes Skorodumov (1948, 1955), and later Vlasyuk (1953) classified deluvial soil by the thickness of the deposited layer (up to 25 cm, 25–50 cm, 50–100 cm, over 100 cm).

In another classification scheme Presnyakova (1959) departed from the concept of the thickness of genetic horizons, and divided the deposited strata into five classes, in which the thickness of the deposit is determined, both by the proportions of the different horizons (A_1, B_1, B_2, C) and by the absolute thickness of the deposit. She recognizes, as does Skorodumov, soils containing 1. small deposits (up to 25 cm), 2. medium deposits (25–30 cm), 3. heavy deposits (50–75 cm), 4. very heavy deposits (75–100 cm) and 5. buried soils or deposits (thickness of deposit exceeding 100 cm).

In the author's opinion, both the thickness of recent deposits and their quality (the latter depending on the origin) are important. In the initial phases of erosion the humus horizon is removed and gradually other horizons are attacked; in the fourth and fifth phases the poor-quality B horizon, or the substratum with its increasing content of clastic material and detritus are carried away. Based on the classification of eroded soil (Table 12) the author proposes, in accordance with the principles mentioned, as an example, the classification of soil (Table 17).

Table 17. Classification of buried three-phase soils

Grade	Soil burial	Description of buried soil
1	Slight	Thickness of deposit exceeding 25 cm
2	Moderate	Impoverished layer on surface, thickness of deposit exceeding 50 cm
3	Deep	Thickness of deposit reaching 75 cm, deposit less fertile
4	Very deep	Profile often buried by loose substratum material also, thickness of deposit up to 100 cm
5	Exceptionally deep	Profile buried by mineralogically less rich earth, thickness over 100 cm

Thickness data on deposits tend to be approximate and depend on the thickness of the genetic horizons of eroded soil. The critical thickness limit for deposits is considered to be about 100 cm; Gael' and Smirnova (1965) consider 300 cm as being critical for light sand, while Bennett (1939) mentions 180 cm for this limit. This critical limit refers to the thickness of deposit above which agricultural crops are no longer able to utilize the buried horizon; the roots are developing entirely in the deposit. Besides the thickness of deposits, their other properties are, of course, important also. For example, a 10 to 20 cm layer of coarse deposit may substantially impair the ecologic properties of the soil.

The classification of soil buried by aeolian sediments is discussed in more detail in Section 2.5.4.

2.8 Classification of eroded land

2.8.1 General

Erosion and accumulation phenomena do not proceed in the same way on any one surface. Not only does the intensity and form of erosion vary, but the relationships between the various stages of the erosion and accumulation processes change also. In order to quantify erosion processes occurring on large or small areas by some sort of index, uniform criteria need to be developed which can, as accurately as possible, give a general account of the actual situation. In each generalization, of course, a predetermined level for the degree of accuracy to be maintained is important.

An important requirement for classifying eroded land is the derivation of synthetic parameters from a detailed investigation in which several erosion indices (intensity of erosion, form of erosion, degree of soil wash, etc.) may be combined. For example, the method of using sample territories which are distributed according to the orographic units to be investigated is considered to be most appropriate. On these territories research is carried out to determine the relationships between erosion indices and geomorphological conditions. By establishing how do erosion intensity, or the degree of erosion of the soil depend on the relief, it is quite feasible to give an account of the surface conditions of eroded soil by morphometric methods.

2.8.2 Methods of classifying eroded land

Several methods have been prepared for deriving synthetic expressions for the degree to which the soil is affected by erosion (the *erodedness*), and for the susceptibility of a particular territory to erosion, these methods having been used mostly for the mapping of various indices on maps showing the occurrence of an erosion phenomenon. In most cases the erosion, or erodedness of a particular territory has been assessed from the surface occurrence of the prevailing phenomenon in terms of the type and form of erosion, or from its intensity.

Thus for example Sobolev (1948) uses the following criteria for judging the intensity of precipitation erosion and the degree of soil erodedness in the Soviet Union: first degree – wash activity weak or absent, terrain level, erosion not pronounced, erosion control necessary on up to 10% of the arable land; second degree – moderate wash activity considerable locally, erosion moderate, erosion control required on 10 to 25% of arable land; third degree – wash activity moderate, 25 to 40% of arable land in need of soil conservation measures; fourth degree – wash activity very heavy, over 75% of tilled land affected by erosion. This presupposes that there is a constant relationship between the proportion of the surface eroded and either the intensity of erosion, or the degree of soil erodedness.

Another method in which the erodedness of the territory (the terraces in the middle of the river Don) was classified by wind strength, was used by Smirnova (1963). Her classification (Table 18) is adapted to local conditions and includes areas damaged by the accumulation of deposits.

In Hungary, Stefanovitz (1964) expresses the different weightings of the degree of erodedness by a coefficient, multiplying the second degree of soil erodedness (in

Table 18. Classification of wind erodedness of land according to Smirnova (1963)

		Representation of erodedness of soil in %				
Grade	Erodedness of area	Slight	Moderate	Severe	Very severe	Thickness of deposit produced [cm]
1	Slight	25—100	0—25	—	—	12
2	Moderate	0—25	25—100	0—25	—	12—25
3	Severe	0—25		25—100	0—25	25—100
4	Very severe[2]	—	0—25		25—100	50—100
5	Sand drifts occurring[3]	—	0—25		25—100	100—300

1 — grades of erodedness are given in Table 14, 2 — deeply eroded soils — up to 25%, 3 — deeply eroded soils — over 25%.

terms of per cent occurrence) by the coefficient 2, and the third degree by the coefficient 4. In the first degree and in cases of sedimentation he leaves the per cent occurrence unchanged. A similar method was used in Romania by Moţoc (1963).

The author considers it better to express the surface area of soil erodedness of a particular territory, or the occurrence of soil erosion, in terms of the weighted mean computed according to the formula:

$$SE = \frac{P_1 + 2P_2 + \ldots + nP_n}{P_1 + P_2 + \ldots + P_n},$$

where SE denotes the average degree of erodedness for the whole territory, P_1 is the surface covered by the first degree of erosion, P_2 is the surface covered by the second degree of erosion, etc. The degree of erodedness or the intensity of erosion in the area covered by the nth degree of erosion (P_n) is decided according to the previous classification. The advantages of this method are that in evaluating the erodedness of the whole territory a higher degree of erodedness attracts a greater weight than a lower degree, and that the harmful effects of accelerated accumulation which can occur in several types of erosion, can also be taken into account, a single figure being used to express all the component indices. By interpolation of the aggregate indices the relative erosion damage of the territory may be established.

In the USA soil affected by erosion is classified by *land utilization capacity,* a total of eight site classes being distinguished. The first site class comprises very fertile soils which can be cultivated without any land restructuring or with only the simplest preparation such as the removal of shrubs, simple drainage, etc. The second site class includes soils of medium to good fertility. Soil utilization is made possible by simple soil conservation measures, such as ploughing along contour lines, cultivation of soil-protecting crops, and drainage by means of small channels. The third site class is formed by areas of medium to good fertility which require intensive soil conservation measures involving terracing, sowing in strips, the application of large doses of fertilizers and the drainage of surface water with the use of expensive equipment. Land belonging to the first, second and third site classes can be kept in permanent cultivation provided that the necessary soil conservation measures are observed. The fourth site class includes land of medium fertility which is suitable mainly for grazing and meadows. Such soil may be ploughed only in the first and second years of a six- to twelve-year cycle, the soil then being protected by grass for the remainder of the cycle. The next four site classes are unsuitable for ploughing and must therefore be used for pasture (with the use of control measures) and protective forests (fifth, sixth, and seventh classes) or, in the case of the eighth class left unexploited (Bennett 1955).

2.8.3 Wasteland

In this section we return to the subject of *wasteland* or *waste soil,* also known as *badland;* in Russian the term used is *brosovye pochvy,* in French – *mauvaises terres,* or *terrains nus,* and in German – *Ödland,* these terms having different implications in their meaning. The terms are therefore not directly interchangeable and their original must be taken into account together with a consideration of other terms expressing bad or degraded land. Nevertheless, these terms are often interchanged or replaced by others, and therefore a few remarks on this subject will be relevant.

First of all, it should be stressed that all these terms refer to areas in which the soil is denuded (without a cover of vegetation) or almost denuded, and for different reasons is unsuitable for cultivation and agricultural utilization, except at very high cost. The causes of land sterility are various. *Barren* and *wasteland* can generally be called *unproductive land,* a term the author suggests should include all soils which are not used for intensive production because of their bad properties.

In further subdividing this category, the author recommends that unproductive land be broken down into primary, or *naturally unproductive barren land,* and secondary, or *artificially unproductive wasteland.* Whereas barren land is unproductive in the absence of human interference, wasteland has become uncultivable and unusable as a consequence of destructive human interference with nature resulting in degradation of the soil.

Barren land can further be classified as *climatic, lithic, orographic,* or *edaphic barren land.* This includes barren land in cold, dry or warm regions, or in regions with steep terrain, high precipitation, or sterile, toxic soil. Extensive areas of the polar regions, steep and cold mountain regions, arid deserts, etc., contain land of

Fig. 62. Wasteland caused by magnesite dust and erosion near Košice (Czechoslovakia). (Photo D. Zachar.)

this type. The utilization of such areas becomes possible only when the necessary, and usually difficult measures are taken to protect and improve the land.

The author regards all areas degraded by man as being wasteland; these may differ in character according to the way in which soil has been degraded. The first group comprises soils that have been degraded chemically by accelerated *leaching, salination, contamination*, etc. In this case we may speak of the *chemical degradation of the soil.* The second group includes wasteland which has suffered from erosive degradation; the soil is in the fourth and fifth stages of erosion, or in a stage of erosive disintegration, destruction of the soil mantle, or even of the bedrock being regarded as a feature of wasteland. According to the way in which wasteland has developed, chemical erosive or other forms of wasteland may be distinguished (Figs. 62–64). The process of wasteland creation is denoted by the term *soil desertization.*

Fig. 63. Deforestation followed by grazing was the cause of erosion and the formation of this wasteland. (Photo D. Zachar.)

In a broader sense wasteland is synonymous with the German term *Ödland.* The term *badland* refers mainly to eroded land alone, but descriptions of badland soils available to date show that these include both primary, or barren land, and secondary wasteland. An essential feature of badland is highly intensive erosion, the degradation process being mediated by precipitation or wind erosion, or by some other accessory factor.

Fig. 64. Injudicious soil management was the cause of very severe erosion and the formation of wasteland over the whole of this territory which consists of soft fluvioglacial sand deposits (Melnik, Bulgaria). (Photo D. Zachar.)

2.8.4 Soil chains

Erosion also has an influence on pedogenic processes. Since gravity is an erosive force of great importance, the influence of erosion on soil development can best be observed in those parts of the relief where various stages of the denudation-accumulation process, in which erosion plays an important role, are linked one to the other. This phenomenon was well known in the first pedological investigations, and the influence of the relief on soil development has been studied in detail by Dokuchaev (1877), Vil'yams (1949), Russelle (1950), Bedrna and Džatko (1963), Mičian (1965), Pelíšek (1955, 1964), and by others.

The part played by erosion-accumulation processes as regularly occurring phenomena in the genetic differentiation of soils was first described by Milne (1935, 1936), and later by Vageler (1940, 1955). Both authors made their investigations in regions of intense erosion and established that soil on the divide, soil on the slope, and soil at the foot of the slope differ not only as a result of the uneven distribution of water, wind force and transported material, but also because of the continuous transfer of soil. Thus eluvial, deluvial, alluvial and other soils develop which, in spite of similar climatic, geological and other conditions, undergo different development. Because these soils, although different, are nevertheless genetically related and linked to one another, the above-mentioned authors named them as *soil chains*, or *catenae* (Latin *catena* – chain).

Milne and Vageler define the soil catena as a series of soil types which undergo continuous development under the influence of removal, transport, sorting, sedimentation and increasing reduction of the soil material from top to bottom of the slope, i.e. from the divide to the erosion base. The soil types of the catena series are characterized by specific plant associations. The catena as a whole is a hydro-climatic function of soil development which depends on the relief and the basic petrographic material. Erosion not only modifies the further development of soil profiles already in existence, but influences also the regional distribution of soil types over an extensive area (Milne 1936, Vageler 1955).

Depending on the existing relief, the soil catena may vary in the extent of the area it covers. Catenae arising as a consequence of the microrelief have been described in the literature under the term *soil complex* (Rode 1955, Vageler 1955). This takes into account variations in the soil-forming process resulting from small modifications of the relief, and the different degrees of influence of underground water on genetic horizons, etc. Such phenomena occur most frequently where the relief is of aeolian origin and where small changes in elevation give rise to variations in soil development. The author recommends that soils forming soil catenae over small areas, and developing independently of the influence of the mesorelief, be termed soil microcatenae, this being equivalent to soil complex.

In the mesorelief more complicated catenae develop with at least four components, the eluvium, the slope deluvium, the slope foot deluvium, and the alluvium. Within any one part of a mesocatena, microcatenae may occur. Thus for example in karst region erosion-accumulation microcatenae may occur in funnels, fields and other formations occurring in the eluvial regions of plateaus. The author recommends that the sequence of catenae in the mesorelief be termed a *soil mesocatena*, or *genuine catena*. A classic example of a genuine catena is given by Milne (1936).

Finally, chains of soil types with a large number of links in the series may occur in the macrorelief, viz. *macrocatenae* which represent the variation in soil types with vertical zoning on the macroforms of the relief.

At different altitudes denudation-accumulation phenomena often proceed differently as a result of the large differences that occur in natural conditions as the elevation above sea level increases. In this case, the high altitude system includes the lower one, i.e., the macrocatena includes the mesocatenae and microcatenae. In Czechoslovakia, the vertical distribution of soils has been studied by Pelíšek (1955, 1964), Šály (1962), Mičian (1965), Mičian and Bedrna (1964), and other authors.

Macrocatenae as conceived by the author have been discussed by Fink (1958) who described a macrocatena in the foothills of the Austrian Alps. Within the range of the macrocatena Fink recognized chernozem, leached chernozem, brown earth, illimerized brown earth. Franz (1960) also recognized similar arrangements of soil types occurring at various altitudes in the Alps.

In macrocatenae pedogenic influences are more complicated and the relief also plays an indirect role, for example by exerting some influence on the climate. Occasionally a change in soil type reflects a change in the underlying rock type. Simple catenae arise mainly under the influence of a single factor, other conditions being basically the same over the area concerned. In any case, erosion-accumulation processes play an important role.

Further classification is only in its infancy. The author suggests that *erosion catenae, solifluction catenae,* etc., be distinguished, according to the factor which gives rise to the catena. The climate, the action of underground water, and other factors may play a decisive role, and therefore the author recommends that we speak of *bioclimatic catenae, hydrogenic catenae,* etc. Haase (1961) has investigated in detail the role of catenae in the ecological differentiation of sites on slopes; he arranges a sequence of sites into a so-called *ecological catena (ökologische Catena).*

The main feature of *erosion-accumulation catenae* is the differentiation of the soil profile properties in the non-affected, disintegration, and accumulation zones. In general, as the intensity of erosion increases, the resulting changes become more pronounced; the greatest change occurring in soils of the erosion belt. The latter eroded soils are given various names, as discussed in the previous chapter. Mückenhausen (1962) writes of *ranker soil;* Fink (1958), Franz (1960), and others speak of *raw soil (Rohboden).*

Following from the terminology outlined above, the set of soil types which regularly alternate in belts over the relief could be referred to as the *chain ring,* or *crown.* This term would embrace both vertical and horizontal influences. If the relief is formed by several types of rock, a whole *mosaic* of *soil types* comes into being over the territory, and although these are genetically related, their ecological value may vary widely. Generally it may be assumed that soil types can be grouped according to their genetic relationships into *edaphic* or *ecological associations* which are always influenced to a considerable degree by the bedrock. Soils cannot therefore be studied in isolation, but should always be considered together with the bedrock on which they have developed.

The study of the influence of erosion and denudation processes on soil development is still in its beginnings, notwithstanding the wealth of experience that has been obtained regarding the influence of the relief on soil development. New problems which have to be faced in soil conservation and improvement cannot be tackled without a reappraisal of the influence of the relief on erosion, and the influence of erosion on soil development.

2.9 Conclusion

In concluding the chapter on the concepts and classification of erosion phenomena, the author wishes to stress the need for an expansion of the concept of soil erosion to include not only the present notion of so-called normal conditions occurring under the natural vegetation of the humid conditions, but rather any set of conditions under which soil is damaged and destroyed by erosion. The part played by man in degradative erosion may be either negative or positive. In a negative sense man acts upon the soil mainly by accelerating erosion processes, and in a positive sense he may slow them down by way of conservation and by bringing about changes in natural conditions.

The purpose of the definition and classification of erosion phenomena is to recognize the various erosive forces in terms of their forms, intensity, development, activity, and other criteria which are important for an understanding of their nature; such an understanding assists in the selection of the most effective and most economically advantageous erosion control measures. Soil erosion is understood as one component of the planation complementary triplet: disintegration, transport, and deposit.

The critical dividing line between harmful (malignant) and harmless (benignant) erosion depends upon the average intensity of soil formation by weathering processes. If we suppose that erosion removal equals soil formation, permissible (or tolerable) erosion occurs, and this is the state, which conservation measures should aim to achieve. The actual level of erosion achieved after the introduction of erosion control may be referred to as the level of inhibited erosion. Permissible erosion may be determined not only by the rate of soil formation, but also by the current state of the soil, the properties required of it as well as by economic considerations.

Of all erosion factors (water, snow, ice, wind, earth, living organisms, man), water, wind, and man exercise the biggest erosive influence on the soil. With regard to the influence of man, so far the tendency has been to stress only the indirect aspect of this in which the influence of water and wind is accelerated by human activity.

Erosion factors mostly affect the soil surface, and only to a small extent do they act within the soil or on the bedrock. Accordingly surface (exomorphous) and underground (cryptomorphous) erosion phenomena are distinguished. Of underground phenomena, underground water erosion (intrasoil and tunnel erosion) is the most readily investigated.

The classification of soil erosion by the forms arising from the deformation, and disintegration, respectively, of the soil mantle is important. In precipitation erosion, not only the well-known phenomena of sheet and gully erosion, but also multimorphous erosion and some other special forms are recognized. In wind erosion, deflation, corrasion (and with regard to laminar wind action) wind detersion are distinguished.

The course taken by erosion phenomena is connected with natural long-term, and current or recent factors, the principal of these being man's influence on nature. In accelerated erosion, the development course of erosion can be observed in a short period from its initial to its final stages. Under natural extreme conditions where a soil mantle with a normal profile has not developed, erosion phenomena tend to be of an irregular cyclic nature. From the economic viewpoint it is important to study the erosion process especially with regard to the course of precipitation and wind erosion.

Erosion remains form a separate topic, these being classified with respect to surface occupied, profile, or proportion of original soil mantle existing as remnants. Accordingly the remains are referred to as relics, rudiments, and selective and inhibitive remains.

Sediments too, are classified according to related erosion phenomena. Special attention is paid to precipitation (pluvial) and wind (aeolian) deposits. Of the former deluvial, colluvial, and proluvial deposits are distinguished. Wind deposits are classified mainly by their form, the proportion of aeolian admixture, and the nature of the latter. Produces of intrasoil erosion are termed eluates, products of mechanical precipitation are termed erosion deluates, those of fluvial erosion alluviates, and of wind erosion deflates, etc.

Finally, the chapter deals with the classification of eroded soil or eroded surface, and with the pedogenic influence of the erosion-accumulation process. As a consequence of this process, soil chains (catenae) or mosaics of soil types may develop. In the initial, vegetationless (aphytogenic) period of soil development from the bedrock, chain development proceeds mostly by solifluction (of moist, frozen, or dry soil) and by wind erosion; in the vegetation (phytogenic) phase, soil chains develop mainly as a result of washing (deluation) of the soil.

Chapter 3

PROBLEMS AND METHODS OF SOIL EROSION RESEARCH

3.1 General

The problems of soil erosion research are broad and varied. Any aspects of the erosion phenomenon may become an object of research, from erosion factors to erosion control measures. Considering the impact on agriculture, it is obvious that interest focuses on *precipitation* and *wind erosion* and, more specifically, on those forms of erosion that cause the greatest damage to crops. In forestry, attention is concentrated on the control of flow and load in catchment areas and on improvements by both biological and technical means.

This chapter gives a brief survey of the methods used in this research. Because the methods of investigating erosion phenomena have not yet been firmly established, a survey of papers describing variants of methods or comparing results obtained using related methods is also given. In providing a survey of the methods of research, the author hopes to compensate, at least partially, for the lack of a much needed comprehensive account of soil erosion methodology.

3.2 Problems of research in soil erosion

Research of soil erosion is difficult for several reasons, but particularly because soil erosion is an intermittent process. It is therefore extremely difficult to observe the erosion act itself, and so in most cases, only the consequences of erosion are investigated. These are *eroded soils* and *erosion forms* on the one hand, and substances removed from the soil (*eluates* in intrasoil erosion, *deluates* or *colluates* in surface and gully erosion, and *deflates* in wind erosion) on the other. In some cases, the process of erosion may also be assessed from *sediments*. It is generally the case that the greater the time interval which has elapsed since the erosion process, and the greater the transportation distance of the investigated substance from the eroded surface, the lower is the accuracy of interpretation, and the more difficult it is to assess the overall influence of erosion.

Another circumstance which renders erosion research difficult is the fact that erosion may not always be conspicuous and, moreover, that its traces may be rapidly obliterated. This is especially true of surface and wind erosion. Thus, since the direct observation of erosion losses during the erosion process is not possible, erosion can be assessed only by comparing the original situation with soil conditions created by erosion. If the original condition is not known, it is almost impossible to determine erosion-caused changes.

Relatively easy to identify are *momentary* and also *seasonal* changes, especially changes occurring in spring. The recognition of *annual changes* is more difficult, particularly in terms of the average effects which are needed for assessing damage over long periods. With varying degrees of success long-term changes are determined, notwithstanding difficulties in determining the original condition and the time period during which erosion of a particular intensity was taking place.

Finally, a *major problem in soil erosion research* is the fact that erosion does not occur as an isolated phenomenon, but takes place together with other factors. On eluvial soils, erosion effects may be considerably complicated by chemical wash and frost destruction, and on deluvial and colluvial soils by solifluction, landslides, etc. Difficulties are encountered in the determination of the proportions of erosion products, particularly in river, lake and other sediments containing deposits of diverse origin mostly transported under little-known conditions.

Notwithstanding these difficulties, soil erosion phenomena are studied by a wide range of methods. The choice of a method depends mainly on the purpose of the research, which may be conceived in various ways according to subject and area. In studies ranging from highly specific laboratory research to complex inquiries into the continuity of erosion phenomena over large areas, erodological research represents a wide scope of methods from the most simple to the most complicated, various combinations of methods used in other branches of knowledge also being applicable. Therefore, most erosion work has a particular bearing (e.g. with respect to pedology, geography, hydrology, agronomy, forestry, etc.) and a corresponding complex of specific methods.

No matter how diverse the avenues and methods of erosion research may be, the objectives pursued may be classified into the following five main groups.

Intensity of erosion

The *intensity of erosion* is expressed by the intensity of removal from, or deposit on a land surface, the attenuation of soil cover, or the size, density and areal representation of erosion forms created by erosion over a certain period of time. Erosion intensity may be assessed by measuring the amount (weight, volume, depth) of soil carried away or displaced, or of sediments created by erosion. Erosion intensity usually expresses the *quantitative* effects of erosion on soil, i.e. changes and losses in terms of *quantity*.

Qualitative effect of erosion on soil

This refers to changes in the properties of eroded soil, especially with respect to its *fertility*. These changes can be assessed by comparing the initial and the altered soil properties, or those of the initial, eroded, and deposited soil, possibly by comparing the properties of the transported and deposited material.

The quantitative and qualitative effects of erosion on soil together constitute the *erodedness of soil.* This is the state of a soil at a certain time as the result of erosion.

Susceptibility of soil to erosion

This is determined by investigating the resistance of the soil to washing, disintegration, disruption, abrasion, scouring, soil blow, etc. It may be established also by analyzing various erosion factors and conditions, such as the erosion effects of precipitation, wind, topography, plant life, human activity or other factors. The ability of soil to resist the destructive impact of erosion factors is expressed in terms of *erodibility*. Some authors confuse erodibility with *erosion danger,* or *potential erosion,* respectively. These terms have, however, a broader significance.

Effectiveness of erosion control measures

This is determined by investigating the extent to which different *erosion control* and *prevention* measures are able to bring about greater soil permeability, surface roughness, resistance of soil aggregates, mechanical soil protection, reduction and even distribution of surface runoff, reduction of wind velocity, interception of transported particles, etc. Important also is the assessment of the effects of erosion control measures on environmental improvement and the raising of crop yields.

Distribution of erosion phenomena and erosion control measures

The distribution of erosion phenomena and erosion control measures over larger areas is usually investigated. This is a synthesis of all relevant information to determine general relations between erosion and natural and economic conditions and how these vary in time and space. The final objective of this research is to classify eroded land and to develop a programme of erosion control measures that would reduce erosion to a tolerable level and bring about a continuous increase in soil fertility and an improvement of the *human environment.*

The principal objectives of erosion research may be attained by a number of methods. These include the methods of levelling, pluviosimulation, growth of vegetation, and the volumetric, deluometric, monolithic, pedologic, morphometric, hydrological, historical, photogrammetric, deflametric, cartographic, empirically mathematical and complex methods.

Besides these principal objectives, erosion research may also pursue partial objectives, i.e. the study of some aspect of the erosion process. Such objectives may take the form of inquiries into the so-called *aggressiveness of the climate,* the *erosivity* of a particular active factor, etc. Much attention has been devoted in recent years to erosiveness, or the erosion effect, respectively, of precipitation on soil. Research on the erosiveness of factors may use monolithic, deluometric (erosivity of natural precipitation), pluviosimulation (erosiveness of raindrops or artificial rain) and deflametric (erosiveness of wind) methods.

3.3 Methods of erosion research

3.3.1 Levelling (geodetic) methods

Levelling (geodetic) methods in soil erosion research are procedures designed to assess the quantitative effect of erosion by measuring vertical shifts of the soil surface. Levelling methods may be preferred in cases of *latent erosion, deflation,* and also in other erosion forms. Changes in the level of the soil surface may be determined by ambulatory or stationary methods.

The *ambulatory method* of determining changes in the soil surface (the microrelief) may be used where the original soil surface can be determined by some standard (etalon), or by the relics of erosion, or a buried horizon. Such a point of departure may be the border of a forest stand, permanent meadows and the like. Sometimes it is possible to determine erosion changes by levelling, using an imaginary line linking the original, and existing surface level with another part of the slope. This approach may be used, for instance, in investigations into the displacement of soil by ploughing terracing, soil cultivation, etc. In some cases the original depth of the soil cover may also be determined, with some degree of tolerance, from other erosion remnants, assessing the original soil level by estimation. This method may be used to advantage on pastureland where the original soil level is indicated by grass-protected soil remnants.

Ambulatory levelling has the advantage of allowing a speedy determination of erosion losses. The method may be used mainly in complex expeditionary research and surveys. Its disadvantage is that only secular changes may be determined reliably and if the age and standard are not known, the determination of erosion intensity becomes difficult and often approximative.

The *stationary method of levelling* is based on the determination of height differences in the microrelief by repeated microlevelling using a network of stationary fixed points (French *repère*). This method is suitable for the investigation of any erosional changes and should be considered at present as one of the most accurate erosion research methods. It is particularly suited for research on uncultivated land, and on cultivated soil where the etalon is not known.

As *repère, permanent* or *provisional markings* made on stones, metal pegs, impregnated wooden pins, buildings, electricity cable pylons, etc., or the triangulation network may be used. In research on precipitation erosion, the main line of guide-marks (repères) runs along the declivity, and the auxiliary line is perpendicular to the main line. In wind erosion, the main direction is indicated by the prevailing wind direction. A principle to be observed is that there should be at least two *guide-marks* on the main line for the purpose of positioning the provisional points. As provisional points, various firmly fixed pegs may be used. Changes in the microrelief may be determined both in the longitudinal and transverse directions. Negative and positive values of surface heights, ascertained at regularly spaced positions by repeated measurements at successive time intervals, are plotted on millimetre paper on which initial measurements of the microrelief are already marked.

Changes in the level of the soil surface are measured by a *levelling instrument,* in which case the spacing of points may be greater and the method may be applied also to tilled land, or may be carried out with a levelling rod and vertical post. Gerlach (1964) stretched a level wire between the pegs and measured the vertical distance between the wire and the soil surface at intervals of 20 to 50 cm.

A disadvantage of the guide-mark (repère) system is that measurements include changes caused by factors other than erosion, such as those caused by humidity variations, freezing and thawing, cultivation, etc. In order to eliminate these effects as far as possible, it is recommended that measurements should always be made in the same season of the year. The temporary effects of soil cultivation are eliminated by long-term observation.

An even simpler guide-mark method for measuring changes in the soil surface level caused by erosion and accumulation processes is to use *steel needles* or *pickets* stuck into the ground in a line perpendicular to the slope. The needles are driven into the soil by means of handles so as to protrude above the ground to approximately the same extent. During erosion or accumulation the protruding part of the needle becomes longer or shorter, and the erosion process is monitored by measuring the protruding parts of the needles. This method can only be used, of course, on uncultivated land. Similar methods are the measurement of surface level changes using the erosion gauge (Fig. 65), and the measurement of vertical soil movement by means of sticks fitted with corrugated aluminium plates (a calibrated stick with a plate on top used for measuring changes in the soil surface level, i.e. reduction of the level by erosion, or by other soil destroying processes) (Fig. 66).

In vineyards or in areas with scattered tree growth or solitary shrubs, the initial level of the soil may be marked with rings painted on tree trunks or poles not far above the soil surface. Should an accumulation of eroded material be expected, another ring may be painted in a different colour at a predetermined height. By measuring the section of the denuded object below the lower ring or of the shortened section below the upper ring, the intensity of erosion or accumulation,

Fig. 65. Measurement of changes in the soil surface with an erosion gauge; marks are made on metal rods driven into the soil (Belanské Tatry Mountains, Czechoslovakia). (Photo R. Midriak.)

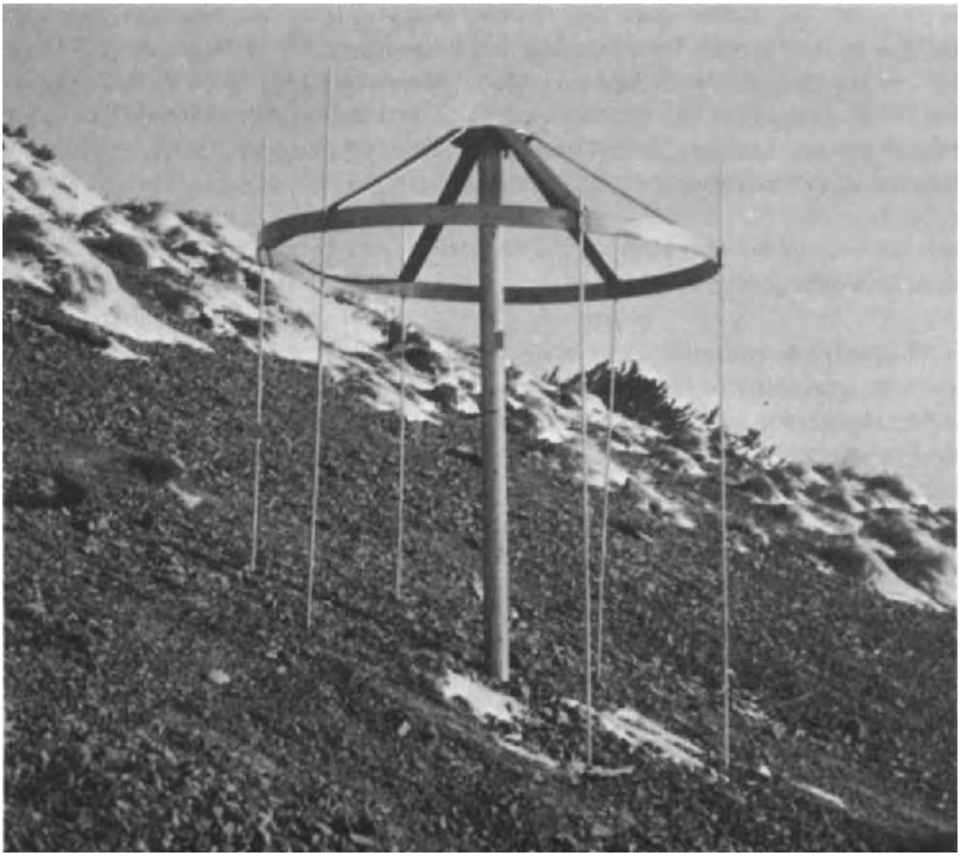
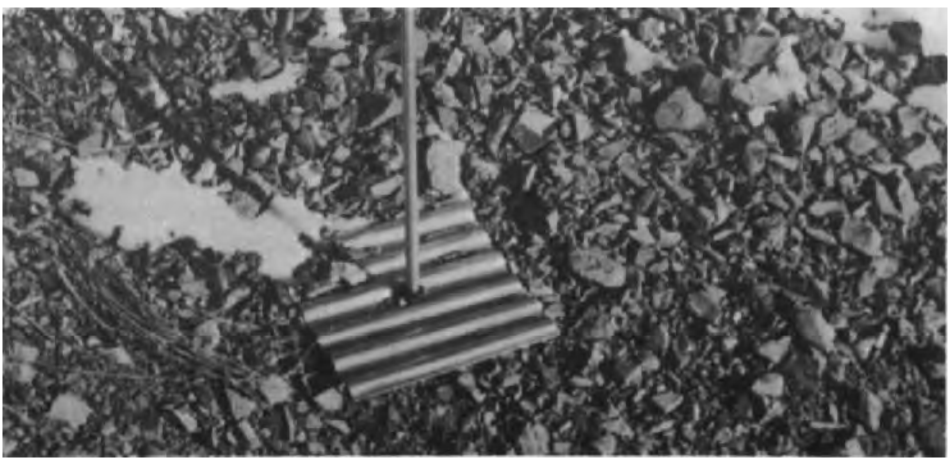


Fig. 66. Another method of measuring surface changes caused by erosion, deposition, frost, or other processes by means of a motionmeter. a – general view, b – aluminium plate – detail (Belanské Tatry Mountains, Czechoslovakia). (Photo R. Midriak.)



respectively, is determined. This method may be used to advantage mainly in vineyards, fruit groves, forest stands, and on pastures.

Levelling methods were used in erosion research by Bac (1928, 1952), Hlibowicki (1955), Pouget (1956), Gerlach (1966), in a study of wind erosion by Kunin and Petrov (1934), and the needle method was used by Gleason (1957), Galyan and Ramenskiĭ (1954), Shvebts (1957), Midriak (1972), and others.

3.3.2 Volumetric methods

Volumetric methods are used in field surveys in which changes in the volume of soil due to erosion or accumulation are measured by single *ambulatory* or *long-term stationary measurements*. Volumetric methods may be used for the observation of almost all surface erosion situations and some underground erosion and accumulation formations, the main condition being the visual expressiveness of the measured phenomena. These methods may be used to advantage mainly in the measurement of the volume of gullies and erosion rills.

Measurement of rill volume

The simplest method for determining the process of sheet erosion in the form of rills is the *measurement of rill volume* by using a meter or a templet. Because the shape of the rills changes rapidly only ambulatory research is practicable. The volume of rills is measured by taking sectors of 20 to 100 m length along the contour line, these sectors being spaced one above the other along the line of the steepest slope from the watershed down to the foot of the slope. The average erosion loss is calculated from the transverse profiles of the rills; from this is derived the volume related to a certain area. Establishing the volume and weight it is then possible to calculate the erosion loss in tons per ha, tons per acre, or any other units.

A good idea of the course and intensity of erosion losses caused by snow or rain-water under various conditions can be obtained by plotting the calculated values on to the profile of the studied slope or on a contour-plan with the known relief and vegetation. In particular the effects of gradient, length, shape and aspect of the slope, of runoff concentration, land cultivation, soil treatment and crop production on the erosion processes may be investigated volumetrically. In the same way also the contents of deposits and the whole process of erosion and accumulation may be studied. Data on rill dimensions may also be used in the calculation of the rill-damaged area and, thereby, the seasonal erosion damage.

An advantage of this method is the speedy collection of data on erosion intensity (both immediate and seasonal), mainly that caused by downpours. The effects of various factors and local conditions may be studied without the expensive constructions and costly research which is important mainly for the preparation of erosion control measures. The method is suitable for the study of soil erosion over larger areas and is usually an integral component of comprehensive research or field surveys. Together with analyses of eroded soil (from different parts of the soil profile) and transported soil, the method will give a good picture of the qualitative effects of erosion on soil, which is particularly important in the study of seasonal variations.

A disadvantage of the method is that real values are underestimated because erosion in the space between rills is neglected, as has been pointed out by several authors (Makkaveev 1955, Shvebts 1957). According to the author's measurements the values obtained by this method are underestimated by 10 to 30%. Another disadvantage of the method is that during torrential downpours, especially, the contours of the rills become indistinct, allowing only an approximate determination of their dimensions. The distinctness of the rills often depends on the roughness of the surface and its microrelief; the smoother the microrelief, the more accurate are the results of the method. Notwithstanding these shortcomings, the method is irreplaceable in erosion research, particularly as a means of determining the course of erosion in different parts of the relief and will remain one of the basic erodological methods.

Volume measurement of erosion gullies

As in the case of gullies the *dimensions of erosion gullies* may also be determined. Besides volume, the *growth* and *length (density)* of *erosion gullies* in the affected area, the proportion of the surface damaged by gully erosion, the development of erosion forms under different conditions, etc., are frequently determined. By analyzing the forms of the gullies (particularly the transverse and longitudinal profiles) in various geomorphological situations, and by complementing the measurements with further analyses, it is possible to obtain a good picture of the course and development of gully erosion, factors which are important for the preparation of erosion control recommendations.

The volumetric method, on account of its accuracy and because of the need for only one set of measurements in time when applied to gully erosion, has considerable advantages, particularly in detailed research. In open country this method may be substituted by photogrammetric methods, and, in the study of gully erosion over larger territories, by morphometric methods. But no matter which method is used, the gully dimensions are always determined in the field either by direct measuring with rod, tape or by other means, or indirectly from various maps and photographs.

By repeated measurements or gully dimensions on permanent plots using a network of points, the dynamics of erosion phenomena and the intensity of erosion during the period of investigation may be determined. In most cases the increment in length, area and volume, the development of the erosion curve, the shape of transverse profiles, etc. are observed. The nature and intensity of river and lake erosion may be monitored in a similar way.

An unusual method was adopted by Ermakov (1962), who measured the *thickness of talus-fans* and the *intensity of talus accumulation* by *electrometric sounding;* interpretation is based on differences in the conductivity of debris and bedrock. Ermakov considers this method to be appropriate and expedient mainly in high mountainous areas with difficult access. Similarly, other geophysical methods, including radioisotope methods, could also be used for the determination of the volume of deposits.

Erosion intensity by measuring the volume of gullies has been studied by Zemlyanitskiĭ (1937), Dryuchenko (1938), Bamesberge (1939), Sobolev (1941, 1948, 1960), Kozlov (1949, 1953), Presnyakova (1949, 1953), Schultze (1952), Gleason (1957), Doshchanov and Muratov (1954), Biolchev and Sirakov (1956), Zachar (1956, 1958, 1960, 1970), Gerlach (1958, 1964), Oswiecimski (1961), Woźniak (1963), Midriak (1965, 1966, 1969), and others. Gully erosion by the method of measuring the volume of eroded soil was investigated by Rozov (1927), Pronicheva (1955), Sokol (1955), Zachar (1956, 1960, 1970), Demek and Seichterová (1962), Štelcl (1962), Czudek (1962), Košťálik (1965), Midriak (1965, 1969), and others.

3.3.3 Deluometric methods

Precipitation erosion is determined by *deluometric methods,* the *amount* and *quality* of the products of *deluation,* or *deluates,* being measured. In this way it is possible to establish, with a relatively high degree of accuracy, the intensity and course of erosion under precisely fixed conditions, and to investigate erosion factors. The methods involve the interception of the surface runoff containing the eroded substance and weighing the latter. In order to relate the intercepted substance to the surface, it is necessary to know the area yielding the deluates. Therefore, the most accurate data possible are obtained from catchment areas. In this case, too, ambulatory and stationary methods are distinguished.

Ambulatory deluometric methods involve measurement of the content of silt under natural conditions using *interception* collector, or *deluometers.* At the position where the flow of deluates is to be intercepted, the container of known volume is installed. From the bottom of the container leads a pipe through which the water flows into bottles, these being filled at certain intervals. If the time taken to fill the bottle and the total time of flow are known, the flow of water and, after establishing the turbidity of the water, the flow of silt also may be computed. One modification of this method is described by Sobolev (1948).

The advantages of this method are that it is quick, expedient, allows the flow in various parts of the terrain to be established, and can be used in various agricultural situations, and in comprehensive erosion research in specific regions during one or two seasons. Owing to the simplicity of the method, one observer may attend a number of deluometers. Disadvantages are that this method is less accurate than the stationary method. In the author's studies the turbidity of water flowing from fields and other tracts of land was determined using sample bottles. Data obtained in this way serve only for purpose of orientation; moreover, only relative data are obtained by such a procedure.

Stationary deluometric methods

In the literature these methods are often described simply as *stationary methods,* or field erodological methods, and *field experimental methods,* respectively. They represent an accurate means of measuring erosion losses over elementary catchment areas. In this method, the proportion of surface runoff water is controlled together with the silt, and thus an overall picture is obtained with respect to the effects of precipitation, cropping, soil cultivation, relief, and other factors on both the course of erosion and the water balance. The surface water carrying the silt flows into the measuring device, and the quantity of eroded soil is determined from the weight of the deposited, dried soil, and from the evaporation residue of the surplus flow. This method is best carried out using sedimentation and reduction tanks with various adaptations, and channels equipped with a limnigraph. The complete outfit includes the necessary meteorological apparatus together with an ombrograph and instruments for measuring soil humidity. The intercepted soil is subjected to laboratory analyses. Comparing the soil of origin which is being eroded with the silt, the effect of erosion on the soil may be evaluated as to quality, too. Data thus obtained may also serve for more detailed analyses. The method is suitable for the long-term observation of permanent stations requiring full-time monitoring (Figs. 67, 68).

One of the instruments used for fractioning water samples and sediments is the *multishod divisor.* This is a board with a series of vertical slots mounted on the end of a rectangular box between two interception tanks.

Later the *Coshocton octave hyperbolic rotating vane sampler* was developed. In a water channel equipped with a flowmeter the flow-rate, the total runoff, and the soil loss during one downpour are monitored. A rotating device collects an aliquot sample of the flow. If there is a great deal of sediment, a flow retarding device is needed in order to prevent sedimentation in the channel (Fig. 69).

In some types of work on permanent experimental plots, mainly erosion capability and erosion caused by precipitation or snow melt water (both of which are the basis of the assessment of the aggressiveness of the climate), are evaluated.

Fig. 67. View of a stationary runoff plot with a simple retention trough in use on an erosion control research station in China. (The author's collection of photographs.)

A disadvantage of the stationary deluometric method is the fact that the surface runoff from a limited area does not take place like in a natural way, since the isolated area under observation is functioning more like the upper, evenly inclined part of the overall slope, thus giving values lower than the true values. The smaller the catchment areas, the more distorted are the data. In one series of experiments the effects of precipitation, cultivation and slope length may be observed, whereas another series of experiments must be carried out in order to compare the effects of inclination, aspect, slope profile, geological substratum, soil type, etc. Consequently research becomes expensive and on large areas costly equipment must be installed; on small areas accuracy is lost and it is difficult to simulate soil cultivation by heavy machines at consecutive yearly intervals. Notwithstanding these disadvantages the method yields exact data and is also being used in special research on the effectiveness of control measures.

A simplified method of measuring erosion intensity by determining the removal of washed soil on uncultivated land was described by Gerlach (1964), who

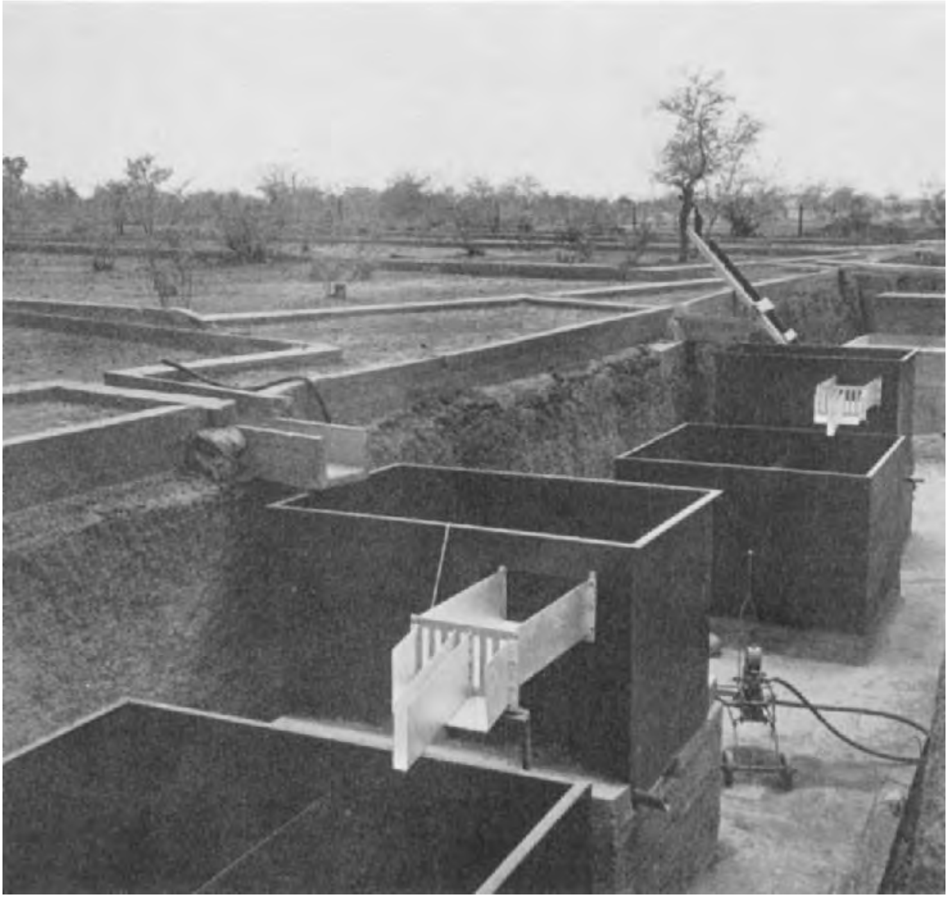


Fig. 68. A well-equipped erosion control research station with a series of suitably large elementary plots with reduction vessels (Quahiquoya, Upper Volta). (The author's collection of photographs.)

arranged the various catchment areas on the slope one below the other, each area being equipped with an interception container. In this method Gerlach used, instead of the original angular container (Schmid 1925 in Gerlach 1964) measuring 34 × 16 × 5 cm, a different container in the form of a trough – 50 cm long, 10 cm broad, and 8 cm deep. In order to obtain more representative results, Gerlach recommends that two or three troughs should be placed at each measurement point. The intercepted water is emptied through an opening in the bottom into bottles. An advantage of this method is that it allows the measurement of wash water on different parts of the slope; one disadvantage is that the soil cannot be cultivated when the measuring device is installed permanently, and therefore the method is recommended only for research on uncultivated slopes. Also the terrain

Fig. 69. A water and sediment measuring station, northern Mississippi. Above – H type galvanized stand flume with O Coshocton-type sediment sampler. The wheel diverts a sample from the total flow into a container where the sample is subdivided. Below – device for measuring runoff and sediments in an artificial catchment area. (By courtesy of U.S. Forest Service.)

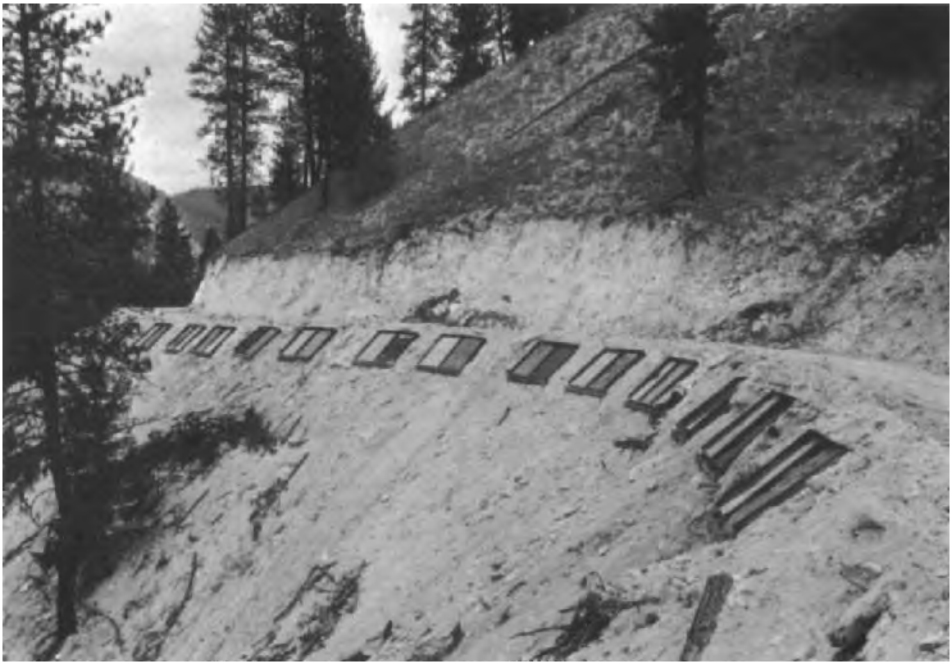


Fig. 70. Research on the protective effect of surface mulching. A series of small experimental plots are used: measurements of runoff and silt are not made (Idaho, USA). (By courtesy of U.S. Forest Service.)

studied with this method should be even, because local accumulation of water could distort the results.

The first experimental plots were established by the Forest Service in Utah in 1915. From among the many research workers who used stationary deluometric methods, mention should be made of Neboľsin and Nadeev (1937), who used 80 × 31 m areas, and Yunevich (1937) and Shcheklein (1937, 1938), who worked with 50 × 2 m runoff areas; valuable results were obtained by Manilov (1939) on 40 to 500 m^2 areas, by Roshchin (1938) on 15 × 8 m areas, by Kazakov (1940) on 20 × 3 m areas, by Firsova (1949) on 20 × 6 m areas, by Gonchar (1956) on 100 × 2 m areas, by Doshchanov (1957, 1962) on 3 × 5, 10, 20, 30, 40 m areas, by Aĭrapetyan (1962) on 50–100 m^2 areas, by Chernyshev (1964) on 165, 185, 30 m areas, and by Burakovskaya (1963) on 4–6 m × 30–40 m runoff areas. Mustafaev (1962) adopted the stationary method in afforestation research, using areas of 270 m^2.

In the USA, New Zealand, Africa, and other countries (see the works of Bennett 1939, 1955, Borst 1945, Browning 1948, Hays 1949, Musgrave 1954, Wischmeier

Fig. 71. Soil Conservation Research Stations such as this one at Wagga carry out research and investigations aimed at improving techniques of erosion control. (By courtesy of Soil Conservation Service of N.S.W., Australia.)

1955, Hudson 1957, Hayward 1969) runoff plots used in the study of soil erosion were mostly set out with dimensions of about 76.6 × 6 ft (equivalent to 22 × 2 m, 0.01 acre, or 0.004 ha, approximately). The U.S. Forest Service used stationary microplots in their soil conservation research on the mulching of road banks in Idaho (Fig. 70). Kuron (1954, 1956), Jung (1956), and Schreiber (1956) from Germany report results of erosion research on plots of 8 × 2 m, while Niewiadomski and Skrodzki (1959) from Poland used plots of 120 × 7 m, and Słupik (1973) used plots of 110–230 × 5 m and 115–130 × 13 m. In Czechoslovakia stationary research areas were established in the Ještěd Mountains on pentagonal experimental plots of 3,000 m^2 area (Mařan 1957b), or on rectangular plots measuring 19.8 × 6 m (Holý 1965). Recently, standardized stationary plots have been established as part of a comprehensive erosion research project in Australia (Fig. 71).

It should be added that data from stationary experimental stations have revealed some precise mathematical relationships between erosion factors and erosion losses and have led to the setting up of equations which enable calculations of precipitation sheet erosion to be made, and which maintain their validity. The most thorough analysis of these relationships was made by Wischmeier (1955), who based his investigation on the records of 65,000 storms, 8,250 plot-years, and 2,500 watershed years. The Soil Conservation Service of the USA was at one time operating 44 experimental stations, and very many others are operated by the Forest Service and other Federal Government Agencies, and by the State Extension Services, Universities, and Colleges (Hudson 1971). An extensive network of experimental stations has also been established in the USSR, Yugoslavia and many other countries.

It should be added that in the case of stationary methods, an important question is what size of experimental plot to use particularly what length is best. It has been shown that the shorter the experimental plots, the larger are the errors that are made in the generalization of the results, even though results from experimental plots of equal length are being compared. The effects of the relief on water runoff, and consequently on the erosive activity of surface water are very complicated and cannot be investigated using short plots. Musokhranov (1976), by comparing results from plots of equal length, established that experimental plots on long slopes exposed to torrential rain and snow should not be less than 600 m long and 25 m wide. Differences arising on plots of various lengths under different conditions of soil, climate and topography, were found to be very large.

3.3.4 Deflametric methods

The success of measures taken against *wind erosion* of soil may be monitored by volumetric, pedological, morphometric, photogrammetric and historical methods, as well as by nivelation and vegetation growth methods. Beside these, wind erosion may be investigated using a number of specific *deflametric methods* which focus mainly on the exact determination of the properties of the deflates, viz. the particles carried by the wind. By analyzing eroded and blown soil with respect to granulation, structure, and nutrient content, the effects of wind erosion on the soil may be established. These methods may be divided more or less into field and laboratory methods.

The most important data to be obtained on a terrain concerns the quantity and quality of particles carried by the wind under different conditions, and at different heights above the ground. Quantitative data on removal are required for determining the intensity of wind erosion and its relationship with other factors and conditions; qualitative data are required for assessing the selective effects on the soil. To this end, various instruments called deflameters are used, such as *Uggla's deflameter* (Uggla and Nozyński 1962) and *Znamenskii's deflameter* (Gael' and Smirnova 1963). Uggla's instrument for measuring deflation consists of four interception vessels, of traps attached to a vertical revolving axis. By means of wind vanes the traps revolve into the wind.

Znamenskii's device consists of a tube 67 cm long and 10 cm in diameter. The rear part of the tube (30 cm of its length) is cylindrical, and the front section tapers conically so that the admission hole is 3.5 cm in diameter. The airflow inside the tube is retarded by four discs inside the wide cylindrical section, thus reducing the wind velocity and its carrying capacity. In this way conditions are created for the sedimentation of deflates.

Gall (1953) used an *interception box* 10 cm high, 8.5 cm wide and 10 cm deep. In the lower part of the box of the deflameter was a wind screen 4 cm broad. The boxes were placed one above the other at intervals of 10 cm, the lowest being at ground level. This instrument was intended for qualitative studies.

3.3.5 Climatological methods

Pluviological methods used in soil erosion research, and for making measurements of rain erosivity, respectively, belong to the group of *climatological methods* by which the *erosivity of climatic erosion factors* (mainly precipitation and wind) and the effects of *climatic conditions* (mainly temperature, humidity, drying tendency, etc.) are determined. In this respect it would be possible to speak about *climate erosivity* and to refer to the complex of climatic factors and conditions which make up the overall *erosive force* of the climate; some authors speak in this connection of the *aggressivity of the climate.* A more accurate term would be the erosion aggressivity of the climate, which would refer to the degree of erosion that might be expected to arise from the climatic conditions.

These methods are used in all work involving the determination of the erosivity of rain and wind under various climatic conditions.

The methods for determining rain erosivity area largely based on the effects of the energy dissipated during maximal downpour conditions [e.g., with the maximal intensity lasting 15 minutes – Goujon (1968), or 25 minutes – Hudson (1971), etc.].

In the USA a period of 30 minutes maximal intensity rain (Wischmeier et al. 1958) is taken as the basis fot the assessment of the erosivity of precipitation. According to Wischmeier, the product of the energy and the intensity of a downpour expressed in terms of the index EI_{30} (for 30 minutes) varies from 100 to 10,000 J m^{-2} for different downpour conditions. By dividing these values by 100 a range of values for the erositivity of downpours ranging from 1 to 100 is obtained. By summing the EI values for all downpours occurring during a given period, the erosivity of the rain in that period is obtained. This is given as the mean annual or seasonal value; the method of calculation is explained in the next section.

According to the values obtained, maps are drawn which are incorrectly referred to by some authors as maps of *isoerodents*. An isoerodent is a line connecting points of equal erosion intensity rather than points of equal erosivity of downpours or periods of rain. Lines connecting the latter system of points are more correctly referred to as *isopluvioerodents*. In addition to this points of equal precipitation erosion may be said to be joined by *isoplurients*, i.e. lines of the same degree of pluvial erosion (or plurosis).

A disadvantage of these methods is the fact that the indices and other indicators of the erosivity of rains that have so far been used, do not accurately convey the erosion effects of so-called "non-erosive" rains, and still less do they represent the erosivity of snow melt water. A further disadvantage is that the erosivity of rain itself is only one indicator which, although it is of primary importance, does not represent the variability of climate which ultimately determines the degree of danger caused by rain erosion. The greatest levels of erosion do not always occur where the erosivity of rain is highest, but may occur instead where ecoclimatic conditions are unfavourable, as in mountain regions above the timberline or vegetation belt, and in arid or semiarid regions, with the highest values of rain erosivity being found in the tropics. Neither do these indicators express the interactions between rain erosivity and other destructive factors which increase the erosive effects of climatic factors.

Fournier (1960) studied in detail the method of assessing the relationship between climate and erosion, and introduced the term *aggressivity* of the *climate;* this was expressed by the coefficient

$$G = \frac{p^2}{P},$$

where G is the aggresivity of climate, p the precipitation in the rainiest month (in mm) and P the annual precipitation (in mm).

In this case too, of course, only the relationship between the effects of precipitations of different erosivities is expressed. In addition to these highly simplified methods a number of other methods has been developed, but in the author's opinion these require further refinement. Munteanu (in Ionescu 1972) expressed the coefficient of aggressivity of the climate *(K)* as a function of erosion potential as follows

$$PT = \int \frac{KTVL}{DA}$$

Table 19. Classification of K factors in terms of the agressivity of the European climate

Grade	Subgrade	Pluviometric range	Type of climate
K_1	—	15— 50	Semidesert
K_2	$K_{2.1}$	51— 75	Xerothermomediterranean
	$K_{2.2}$	76—100	Thermomediterranean
	$K_{2.3}$	101—125	Mesomediterranean
K_3	$K_{3.1}$	126—150	Submediterranean — temperate-warm
	$K_{3.2}$	151—175	Submediterranean — temperate-cold
K_4	—	176—200	Mediterranean — humid-mountainous
K_5	—	>200	Moderately humid oceanic, or continental

where PT is the potential erodibility of the region (or catchment area) by torrential rain, K the coefficient of aggresivity of the climate, T the topographic factor, V the vegetation and soil utilization factor, L the lithic factor, D the factor denoting erosive properties of catchment area, and A the ablation factor (pertaining to the transport of eroded particles; this depends on various degradation processes).

In this equation Munteanu used a scale for the expression of the aggressivity of climate (coefficient K) which was drawn up specifically for European conditions by Barnouls and Gaussen (Table 19).

Again, in expressing the aggressivity of the climate, emphasis is laid on the erosivity of precipitation expressed mainly in terms of pluviometric indicators. Notwithstanding the method of expressing the erosivity of precipitation, the method takes no account of wind and other climatic conditions as erosion factors. Problems of *wind erosivity* have been studied independently by a number of authors who determined the *erosive effects of wind* by different methods, using the so-called *threshold velocity* (Chepil 1945) or other indicators. Nevertheless, detailed study of wind erosivity lags behind the study of precipitation erosivity, and as yet no methods for the aggregate assessment of wind erosivity are available; consequently, the mapping of *aeolian erosivity* as *iso-aeoloerodents* (in short, *iso-aerodents*) is not yet possible.

In order to express the global erosive "aggressivity" of the climate, i.e. the degree of influence of climatic factors and conditions on erosion, it would be necessary to include all the climatic factors involved in erosion and other destructive phenomena (such as cryogenic disintegration, decerption, etc.).

3.3.6 Pluviological methods

3.3.6.1 Methods of researching the erosivity of natural precipitation

The purpose of *pluviological methods* in erosion research is to determine the erosive effects of precipitation, especially the effects of *raindrops.* In the literature the latter phenomenon is generally referred to by the terms *erosivity of precipitation* and *potential ability of precipitation to eroded soil,* respectively. These terms are given a quantitative value in terms of the quantity of soil eroded by a certain amount of rain, although it would be appropriate to include also the erosive effects of snow melt water. In the most up to date research the erosivity of precipitation is being used for determining the so-called index of erosivity of precipitation, which serves as a basis for the calculation of potential soil erosion, if soil erodibility and the kinetic energy dissipated on the relief are also known.

Because the *kinetic energy of rainfall* is a basic factor in the calculation and determination of the erosivity of rain (according to present understanding), pluviological methods concentrate on determining the relationships between the kinetic energy of erosive rainfall and erosion processes. Since there are as yet no instruments available for the measurement of the kinetic energy of rainfall, the erosive effect of rain is assessed indirectly, by methods which may be referred to as *pluvioenergetic* methods.

For the purpose of determining the erosive effect of precipitation, *pluviometric, pluviographic, pluviodistributive,* and other methods have been developed and used. These methods, which are discussed in detail in texts on climatology, meteorology, hydrology, hydrometrics, and other disciplines, are based on finding relationships between the size of the raindrops and the velocity and kinetic energy of the raindrops, and between rain intensity and the structure or proportion of raindrops with critical and larger than critical kinetic energies; interrelationships occurring between the intensity of the rain, its duration, the total amount of rain, the area over which the rain falls, etc., are also considered.

The first measurement of the sizes of raindrops was made in 1892 by Lowe, who measured the size of the stain made by a raindrop on a slate. By further investigation raindrop growth, drop disintegration, and other processes governed by meteorological conditions became known (in Hudson 1971).

Of the research that has been done on this subject, mention should be made of the detailed studies carried out by Laws (1941) and Gunn and Kinzer (1949) concerning the velocity of raindrops, by Laws and Parsons (1943) on the structure of raindrops in rainfall of different intensities, and by Hudson (1971) on the average size of raindrops and other of their properties. In the world literature there is a great number of publications of this kind and the determination of the erosive effects of precipitation is based on these.

Fig. 72. Soil trays for measuring splash and wash erosion. The porous tiles in the false floor can be seen in the empty tray on the left. (Photo N. W. Hudson.)

A more concrete measure of rain erosion has, however, been gained from research on the *kinetic energy of rain.* The first attempts at measuring this value involved the use of a sensitive balance, or measuring the effect of rain on various revolving devices; the recording and transformation of acoustic effects into electric impulses, for example by placing recording sound-level metres beneath a membrane on which the raindrops fall, has also been tried.

An important step forward in understanding the relationships between the kinetic energy of rain and its erosive effect was made by Ellison (1944), who carried out the first investigation of the process of raindrop erosion particularly with regard to the disaggregation of soil by the *splashing* of *raindrops.* For recording the splash propensity of rain, brass cylinders of 77 mm diameter and 50 mm height are used with a fine wire net welded across the bottom. A thin layer of cotton wool is placed on the net, and the containers are filled with sand or other soil and placed in shallow water so that the soil sample attains a moisture content close to capillary saturation. The containers are dried out and weighed both before and after exposure to rain; the weight difference indicates the *soil loss by splashing.*

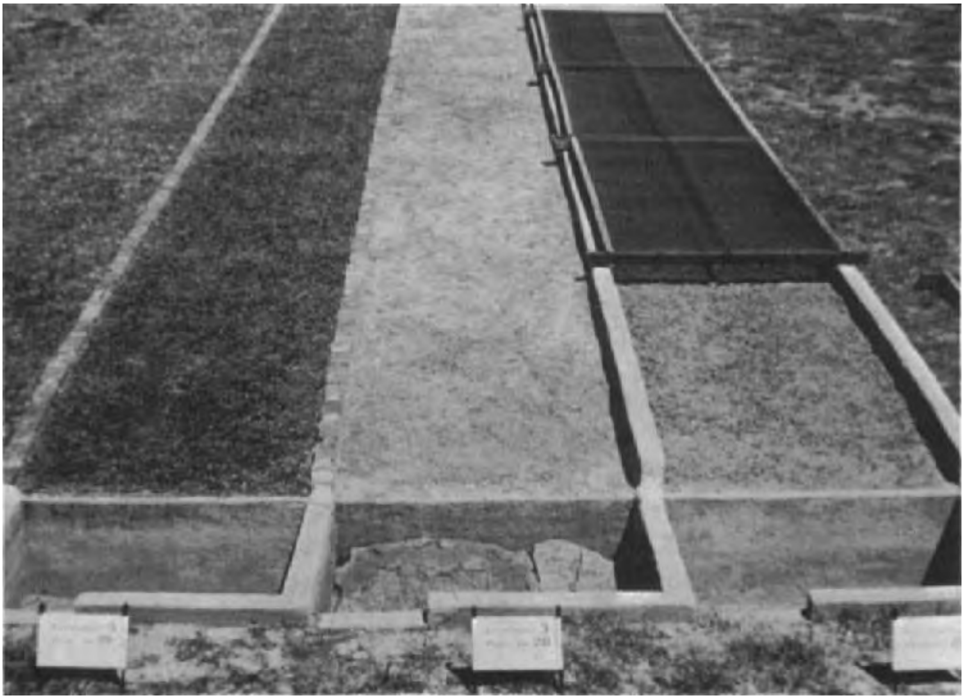


Fig. 73. The field plot which was used to compare soil losses from a plot with estimates of erosivity made from rainfall records. (Photo N. W. Hudson.)

Gradually the technique of using the splash gauge was improved, and data on splash erosion were combined with data on wash erosion; monitoring stations were set up specifically for the *measurement of the erosive effect of rain.* Interesting measurements carried out on parallel plots have been made by removing the effects of disaggregation caused by the splashing of raindrops using a tissue or other cloth. Later, the effects of splash erosion were investigated on the inclined surfaces of soil mounds in order to establish soil transfer by raindrops falling on a slope. At the same time, the effect of a surface layer of water as a protection against splashing, and the rolling of disaggregated particles in water, as well as other phenomena were investigated.

Much attention has recently been given to these methods by Hudson (1971). In vessels similar to those used by Free (1952) he placed soil instead of sand. The pans were 100 cm long, 30 cm wide and 10 cm deep, and were arranged on a 1 : 20 slope (Fig. 72). The pans were filled with a sieved clay loam soil which was maintained at a constant moisture content by connecting a tray to a constant head reservoir. The soil washed from the pan was collected in a trough at the lower edge.

After two years the relationship between soil loss and rain erosivity (expressed in terms of the KE > 1 index) was established.

A further step forward was the use of a field plot 27.5 m long by 1.5 m wide on a 1 : 20 slope (Fig. 73). This was similar to the plot used by Free (1952) for the determinaton of deluation (wash), except that in additon to a bare soil plot and a grass covered plot, a third plot was used which was bare but covered by cloth in order to protect the soil against the effect of raindrops. Soil loss was measured by the amount of soil intercepted in containers installed at the edge of the plot. The results also showed a close correlation between rain erosivity and the measured soil losses.

The author recommends the use of the KE > 1 index rather than the EI_{30} index proposed by Wischmeier (1955). Whereas Wischmeier together with his collaborators (Wischmeier et al. 1958) expressed precipitation erosivity in terms of the greatest average rainfall intensity experienced in any 30-minute period during the storm and the kinetic energy of this rain, Hudson takes the total kinetic energy of precipitation which has an intensity exceeding one inch per hour as the indicator of precipitation erosivity.

Hudson (1971) demonstrates the calculation of precipitation erosivity by both of these methods in the following examples:

a) Using the EI_{30} method

Intensity [in h^{-1}]	Amount [in]	Specific energy [ft-t $acre^{-1}$ in^{-1}]	Col.2 × Col.3 [ft-t $acre^{-1}$]
0—1	1.5	816	1224
1—2	1.0	974	974
2—3	0.75	1048	786
>3	0.25	1096	274
Total	3.5		3258

I_{30} is 0.6 in h^{-1}.

$$EI_{30} = E \times I_{30} = 3258 \times 0.6 = 1954.8 \text{ ft-t acre}^{-1}$$

Calculated from the expression

$$\text{Specific kinetic energy of rain} = 916 + 331 \log_{10} I \text{ ft-t acre}^{-1}\text{in}^{-1}$$

when kinetic energy is expressed in ft-t $acre^{-1}$, and intensity in inches per hour (Wischmeier et al. 1958).

b) Using the KE > 1 method

The energy of 1.5 in of rain falling at less than 1 in h^{-1} is ignored and the remainder summed,

$$KE > 1 = 974 + 786 + 274 = 2034 \text{ ft-t acre}^{-1}.$$

Note on units

The EI_{30} method was developed in foot-pound-second units and since these units have always been used in its application there seems little point in changing over to SI units. The KE > 1 method is readily usable with SI units, and thus becomes the KE > 25 method. A sample calculation is as follows:

Intensity [mn h^{-1}]	Amount [mm]	Specific energy of rain [J m^{-2} per mm of rain]	Total energy Col.2 × Col.3 [J m^{-2}]
0—25	30	—	—
25—50	20	26	520
50—75	10	28	280
> 75	5	29	145
Total	65		945

A first approximation may be made by assuming a single mean value for the energy of all rain falling at intensities greater than 25 mm h^{-1}.

Thus

$$35 \text{ mm rain} \times 28 \text{ J m}^{-2} \text{ per mm rain} = 980 \text{ J m}^{-2}.$$

Numerical expressions of precipitation erosivity are discussed in Section 3.3.14 and elsewhere (Hudson 1971).

It must be added here that these methods for the determination of precipitation erosivity are not yet very far advanced, although it must be conceded that simply by their use new discoveries about erosion processes have been made, especially with regard to the importance of the erosive effect of raindrops. It is due to experience gained with these methods that *raindrop erosion* may be classed as an important erosion phenomenon and an important factor in the action of torrential rain on the soil.

A disadvantage of these methods is that in their present application the quantitative proportion of soil eroded by raindrop erosion (and the subsequent stages of the erosion process) cannot be measured and expressed numerically. Generalizations made from results obtained by these methods may lead to distorted conclusions, as will be discussed in Chapter 4.

3.3.6.2 Pluviosimulation methods

In order to overcome the disadvantages of stationary deluometric methods involving long observation periods and expensive installations, a speedier approach is to use the technique of sprinkling water on experimental plots, measuring the amount and intensity of the *artificial rain*, the energy of the falling drops, the quantity of surface runoff, and the flow of the deluates. Such methods are called *pluviosimulation methods.*

A container of water is supported usually on a simple stage 2 to 3 m above the ground, and the water is conveyed by pipes to a set of sieves or other devices (Hudson 1971). The energy of the drops is regulated by adjusting their size and the height of the sprinkler. Generally available sprinkling equipment is often used in artificial rain research, the water being dispersed under pressure from a pipe with outlets of different sizes.

Instruments which simulate rain are called *spraying simulators* (Wilm 1943), *rainulators, rainfall simulators* (Meyer and McCune 1958), and *artificial rainfall simulators;* automatic instruments for measuring surface runoff, infiltration of water into the soil, and erosion losses are also required (Hudson 1971).

An advantage of pluviosimulation methods is that in a relatively short space of time which allows for necessary repeat measurements, it is possible to investigate a wide range of factors and conditions affecting the course of erosion. It is possible, for example, to study the effects of different total quantities and intensities of precipitation, of different drop sizes and velocities of fall, of kinetic energy differences in the precipitation, of various conservation measures, and also to study the intensity of erosion on different slope inclinations and under various conditions, for a predetermined type of precipitation.

A disadvantage of pluviosimulation methods is that it is almost impossible to imitate the combined erosive influence of natural precipitation and wind acting together. It is therefore practically impossible to observe wind-influenced impact and splash erosion, and any such observations must be subject to certain reservations. In addition, runoff erosion proceeds differently in artificial rain than compared with natural rain. Another disadvantage of the method is that the erosive effect of snow melt water cannot be investigated. Finally, the small size of the plots used in this method increases the possible deviation of simulated conditions from natural conditions. In series of repeat experiments there is also an increase in the soil moisture content which influences the results. Consequently it is not the actual erosion that is determined in pluviosimulation methods, but rather the erodibility of the soil under certain conditions. It is however true that the erodibility values ascertained in this way are nearer to reality than values obtained by other methods, and therefore the pluviosimulation method may be preferred in erodological research.

Valuable information from pluviosimulation methods was obtained by the following authors: Shaposhnikov (1940) (4 × 2 m plots), Falesov (1939) (1.6 × 2.5 m plots), Polyakov (1939) (2.5 ×1 m plots), Voznesenskiĭ et al. (1940) (1 × 1 m plots), Kobezskiĭ (1949) (4 × 2 m plots), Kozlov (1953) and Burakovskaya (1963) (using microplots of 0.25 m^2), Surmach (1955) (4 × 4, 4 × 2, and 1 × 1 m plots), Nefed'eva (1958) (140 × 80 cm plots). In the American literature the following work is well-known: Duley and Hays (1932) (7.50 × 0.85 m plots), Hendriksen (1934) (3.20 × 0.99 m plots), Zingg (1940) (2.40 × 1.22 m plots), Meyer and McCune (1958) (25 × 3 m, i.e. 75 ft by 14 ft plots). In Czechoslovakia, Mařan (1956, 1957a) used a sprinkling technique on 50 m^2 plots.

Besides pluviosimulation methods, *irrigation methods* are also used in the study of *irrigation erosion*. Again these may be divided into laboratory and field methods. In laboratory research, the soil's resistance against wash is studied, whereas field research investigates the effect of irrigation on the soil and its resistance to erosion. Among these methods fluviosimulation techniques can be included. In this field, successful research has been carried out mainly in the Soviet Union (Mirtskhulava 1970), and also in Bulgaria (Tatrova-Krusteva and Tzonev 1970).

3.3.7 Monolithic methods

The difficulties of setting up sprinkling experiments in the open, and the fact that the effects of slope inclination cannot be studied on the same soil, are problems which are overcome in the so-called *monolithic methods* in which a soil monolith with its structure as intact as possible is subjected to laboratory sprinkling tests. In these methods the conditions under which erosion occurs differ still more from those in nature, and therefore the results that are obtained are only of comparative value. This means that only the erodibility of soil is determined by monolithic methods and this value deviates from actual erosion still more than the values obtained by pluviosimulation method.

An advantage of the method is that by using large soil samples, the inclination of the soil surface, the duration of the precipitation, the intensity of precipitation, and the size of the drops may be varied as desired, and thus the relationships between erosion and these factors may be established under fully controlled conditions. However, the simulation of natural vegetation and soil cultivation on soil monoliths is more difficult. Apart from this, the monolithic method is suitable for the detailed study of the course of erosion, using delicate microscopic and photogrammetric methods of observing the effects of water drops and the fracture of soil aggregates, etc. With this method it is also possible to observe, to a limited extent, the formation of surface runoff on the collected samples, and to subject eroded soil containing silt to pedological tests in order to determine the influence of erosion on the quality of the soil, also.

Of the many publications dealing with soil erosion on experimental monoliths, mention should be made of the pioneering work carried out by the German pedologist Wollny (1895), who used monoliths 80 × 80 cm × 25 cm tall, the extensive work carried out by Gussak (1937, 1945, 1950) with 100 × 40 × 50 cm monoliths, the work of Manilov (1939) also on monoliths measuring 100 × 40 × 50 cm, and the important contribution made by Neal (1938) who worked with 3.6 × 1.1 m monoliths.

Interesting experiments using the monolithic method were carried out by Takiguchi and Namba (1964) on monoliths of 40 × 35 × 15 cm; the effects of varying monolith inclination and precipitation intensity were studied together with the efficacy of erosion control using artificial coverings.

In Czechoslovakia the effect of the degree of slope inclination on erosion intensity was investigated on a hydraulic trough of variable inclination filled with earth. The dimensions of the trough were 12 × 2.5 m with an effective surface area of 14.1 m^2 (8.55 × 1.65 m). The inclination of the trough containing the soil monolith could be changed hydraulically from 0 to 1 : 11 (Holý and Vítková 1970).

3.3.8 Pedological methods

Besides volumetric, deluometric, and monolithic methods, including nivelation and pluviosimulation methods, many techniques for studying erosion and its consequences have been developed with a *pedological basis*. These methods involve either the determination of specific soil properties relating to the *susceptibility*, or *resistance* of the soil to erosion, or the measurement of *quantitative* and *qualitative soil changes* caused by erosion processes. If the initial state of the soil is known, particularly the depths of the soil profile and the various soil horizons, then by comparing this state with the eroded profile the extent of erosion losses may be established, especially in cases of sheet and wind erosion.

Erodibility, as the principal parameter of erosion, can be examined in different ways, the chief of these being granulometric, structural, physical, and chemical methods. The effects of erosion on the soil and on the intensity of soil removal is investigated largely by current pedological methods, emphasis being laid on those properties of the soil which are most changed by erosion, and those which are important to soil fertility. In the overall assessment of the effects of erosion on the soil, various modified techniques are adopted including the extensive comparative method and the pedogenic method.

3.3.8.1 Research on the erodibility of soil

While erosion research was still in its beginnings it was established that each soil type has a certain ability to withstand erosion, displaying an *erodibility* which is closely related to specific soil properties. As far back as 1926, Bennett pointed to the fact that those soils which are resistant to erosion have a good structure, are easily permeable, have a profile with few genetic horizons, and are mechanically homogeneous, too. Furthermore, it has been established that soils in which the $SiO_2 : R_2O$ ratio is less than 2 are more resistant to erosion than other soils.

Bennett also ascribed great importance to the humus content and the mechanical structure of the soil.

After Bennett's work, a number of other studies carried out in the nineteen-forties attempted to establish the relationships between soil erodibility and other soil properties, with the intention of finding some pedological and mathematical factor of sufficient precision to make possible the laboratory determination of soil resistance to erosion, especially precipitation erosion. It should be added that although this work has not yet met with success, the research has nevertheless produced very valuable theoretical information which has been of great importance in furthering erosion research.

One of the pioneers of this work was Middleton (1930), who together with his colleagues established that soil erodibility depends on several factors which are important properties of the soil types he investigated, namely 1. the mechanical structure, 2. the colloid content, 3. the moisture content, 4. the soil density, 5. the capillary water balance, 6. the plasticity (according to Atterberg), 7. soil swelling capacity, 8. soil shrinkage (according to Middleton), and 9. the dispersion ratio. He recommended calculation of soil erodibility by means of the formula

$$\mathrm{ER} = \frac{\mathrm{DR}\ \mathrm{ME}}{\mathrm{Col}},$$

where ER is the erosion ratio, DR the dispersion ratio, ME the moisture equivalent and Col the colloid content. The dispersion ratio is expressed as the ratio of the clay content (according to Middleton this includes particles smaller than 0.05 mm) as measured by chemical soil treatment to the particle content as measured by dispersing 10 g of soil in 1,000 g of water.

It follows from Middleton's research that soil which is resistant to erosion has a larger content of clay particles, a higher colloid content to equivalent moisture ratio, a greater specific weight with respect to the soil phase of the soil, a lower plasticity limit, a lower dust content, a smaller dispersion and erosion ratio. In non-erodible soils ER < 10, and in erodible soils ER ranges from 12 to 115.

Slater and Byers (1931) recommended that for the determination of soil erodibility, account should be taken of the permeability of the soil expressed as the so-called percolation ratio, i.e. the ratio of the content of suspended silt and clay particles to the content of colloids divided by the moisture equivalent. Calculated in this way, the percolation ratio is very near to Middleton's erosion ratio.

Lutz (1934, 1935) came to the conclusion that aggregation of the finest fractions is of considerable importance to soil erodibility and that the life time of aggregates depends on the coagulation of non-hydrated colloids, or expressed in another way, soil erodibility (according to Lutz) is related to the ability of colloids to hydrate.

Bouyoucos (1935) suggested that soil erodibility could be assessed in terms of the ratio of the clay content (particles below 0.002 mm) to the sand (particles from

0.06 to 2.0 mm) plus loam (particles from 0.002 to 0.006 mm) content established in the fine earth fraction

$$E = \frac{\text{sand} + \text{loam}}{\text{clay}} .$$

Bouyoucos intended the *erosion parameter E* to express the proportion of sandy substance in the colloid-bound proportion of the fine earth. Further research has shown that the ratio of the clay content to the remainder of the fine earth is a reliable indicator of erodibility only in some cases. In some papers, the erosion indicator is referred to as the *erosion index of the soil.*

In the Soviet Union, Middleton's work has been continued by Voznesenskiĭ and Artsruni (Voznesenskiĭ 1938, 1940, Voznesenskiĭ and Artsruni 1936, 1938, 1940). They came to the conclusion that the best indication of erosion is given by the indicator of aggregatedness, dispersivity, and hydrophily according to the formula

$$E = \frac{dh}{a},$$

where a is the quantity of aggregates larger than 0.25 mm which remain intact after an hour in a water stream flowing at 100 cm min^{-1}, d the ratio of the fraction of particles of diameter larger than 0.05 mm (determined without chemical preparation) to the same fraction after treatment by the NaCl (international A method), and h the indicator of hydrophily expressed as the water retention of the soil relative to that of 1 g of colloids. On the basis of long-term experiments, the authors recommend the assessment of soil erodibility as shown in Table 20.

Table 20. Classification of soil susceptibility to erosion according to Voznesenskiĭ and Artsruni (1940)

Susceptibility parameter	Susceptibility of soil to erosion			
	I	II	III	IV
	Low	Medium	High	Very high
Aggregation	Under 0.1	0.1— 0.3	0.3— 0.6	>0.6
Dispersivity	Under 0.6	0.6— 0.8	0.8— 0.9	0.9—1.0
Hydrophily	Under 1.0	1.0— 1.25	1.25— 1.5	>1.5
Erosivity	Under 1.0	1—10	10—100	>100

Together with physical and chemical methods, other methods were developed which relied mostly on an analysis of the *physical structure of the soil.* The use of an indicator of soil resistance in erosion research was studied by a number of workers; the most extensive study was carried out by Vilenskiĭ (1935, 1936, 1937, 1938, 1945), who based his work on the hypothesis that the structure is the most important property of cultivated soil, the quality and quantity of the soil aggregates ultimately determining both the rate of infiltration of rain-water into the soil, and the resistance of the upper soil layers to the action of raindrops and soil wash.

In his recommendations of 1938, Vilenskiĭ based the determination of the soil's resistance to erosion on the following measurements:

1. The structural content of dry soil using sieves with 0.25 to 15 mm meshes; the most valuable aggregates were thought to be those of diameter 1 to 10 mm.
2. The density and porosity of aggregates determined with a special volumetric instrument.
3. Rate of water uptake by aggregates, (a) after preliminary capillary saturation, and (b) by dry aggregates in a crystallizer.
4. Rate of water uptake by aggregates during their bombardment by two drops per second from 5 cm height (transverse section of drops 0.03 cm^2). The resistance of the aggregates is expressed in terms of the amount of water required to saturate the aggregate.
5. Intensity of surface wash in a cylindrical sample of soil collected with minimum disturbance of the structure, and at different soil moisture levels. Samples of 10 cm diameter and 3 cm height were exposed to artificial sprinkling in a special apparatus. The rain intensity in Vilenskiĭ's experiments was 1 l per minute, duration 15 minutes. Wash intensity was assessed by the amount of washed earth collected.

As can be seen, Vilenskiĭ's method consists of the direct testing of the erosion resistance to water of aggregates of undisturbed soil, and this has certain advantages over previous methods. The fifth and last test involves pluviosimulation on a small soil sample (micromonolite) under constant conditions in the laboratory. Sobolev (1948) considers Vilenskiĭ's method to be the most reliable presently in use, and Burykin (1962) and other authors have based their determinations of soil erodibility on this method. Partial indications of the degree of soil resistance to erosion can be obtained also by other soil aggregate experiments as were proposed by Savvinov (1931), Novák (1932, 1942), Tyulin (1933), Tsyganov (1935), and others. A remarkable proposition in this respect was made by Ponomareva (1957), suggesting the determination of the *water stability of soil aggregates* by pluviosimulation.

A weakness of the soil *aggregate method* is that it relies on a property which is very unstable both under natural and laboratory conditions, and which is greatly influenced by the moisture content of the sample at the time of collecting, preparing and processing. Moreover, the aggregate method is not suitable for the investigation of soil resistance to gully erosion, in which disintegration usually occurs in the unstructured substratum. It should be added finally that by means of these methods only an indirect assessment of the rate of infiltration can be made, yet infiltration is a critical factor that governs surface runoff. Notwithstanding these disadvantages the soil aggregate methods make a valuable tool both for establishing soil resistance to erosion and for studying erosion processes. Besides the methods mentioned, erodibility as an internal soil property can also be assessed from other indicators such as the granulation of the soil, its porosity, permeability, genetic type, depth, and site conditions, etc. But none of these indicators taken alone without a comprehensive assessment of other soil properties, provides a satisfactory answer.

Similar attempts have been made at assessing *soil resistance to wind erosion,* or the *aeolibility of the soil.* In this case too, *granulation, structure,* and *soil moisture* are the most important factors. In addition to these basic soil properties, the climate, the vegetation cover, the relief (especially the degree of roughness), and other conditions play a role.

In order to standardize the relationships between soil properties and resistance to wind erosion, various formulae have been derived which in most cases make possible the determination of resistance to wind erosion over a wide range of soil conditions. It appears that under some circumstances all soils may be eroded by wind; it is therefore necessary to determine the minimum or *threshold wind velocity* which starts the movement of soil particles of a certain size, under a given set of conditions.

Chepil (1945) established that the critical wind velocity for the smallest grains (diameter 0.1 to 0.15 mm) is 8 to 9 miles per hour at a height of 6 coles above ground level. As grain size increases over 0.1 mm, the critical wind velocity increases with the square of the product of the specific weight and the diameter of the grains; the specific weight of soil grains is usually within the range 1.65 to 2.65.

Because soil properties are so variable, Shiyatyĭ (1972) recommended that the erodibility of an investigated soil should be compared with that of an etalon which, in our opinion, should be represented by a soil with a potential erodibility equalling compensation erosion. In this case, the factor P which is an expression of those soil aggregate properties that have a bearing on soil erodibility, could be derived from the formula

$$P = \frac{P_p \cdot 100}{P_{\mathrm{et}}} \ [\%] \, ,$$

where P_{p} is the indicator for the investigated soil, and P_{et} the indicator for the soil etalon.

According to Chepil et al. (1962), soil aeolibility depends mainly on the *effective soil moisture* level since the latter determines the effectiveness of cohesive forces; the minimum effective value of these forces occurs at a moisture content approximately one third of the permanent wilting point of the soil. Following from this, the rate of removal of soil particles by the wind may be expressed by the formula

$$q = \mathrm{f} \, \frac{v^3}{w^2} \, ,$$

where v is the wind velocity, and w the soil moisture required for effective cohesion.

The dependence of the soil's resistance to wind erosion on granulation was expressed by the relation (Shiyatyĭ 1972)

$$P = \frac{(100 - s)_p \,.\, 100}{(100 - s)_{et}} = \frac{[100 - (a + bx_1 - cx_2 - dx_3)]_p}{[100 - (a + bx_1 - cx_2 - dx_3)]_{et}}\, 100 \,,$$

where $s = a + bx_1 - cx_2 - dx_3$, p refers to the investigated soil, et refers to the etalon, a, b, c, d are regression coefficients, x_1 is the clay content of the soil (in %), x_2 the sand content from 0.05 to 0.25 mm (in %), and x_3 the sand content > 0.25 mm (in %).

Shiyatyĭ expressed the relationship between the rate of soil removal by the wind or soil structuring by the equation

$$q = 10^{a - bk},$$

where q is the rate of removal, k the soil structure (the percentage of fractions of particle size 1 mm diameter occurring up to a depth of 5 cm), a and b are constants. It is supposed that the larger the value of k, the more resistant is the soil to erosion.

Besides these basic indicators, a high degree of dependence of soil aeolibility on the quantity and quality of humus, on calcium content, on clay content, on the dispersiveness of aggregates or microaggregates, and on the creation of soil and ice aggregations, etc. has been established.

Also the regressive assessment of aeolian soil erodibility from the degree of soil aeolization observed directly in the field, is both possible and practical.

Among the special methods that have been devised, attention should be given to the study of wind erosion in aerodynamic tunnels; a survey of this approach is given by Chepil (1945). Suction and circulation tunnels are used for the study of different types of air current, and laboratory, field, and combined tunnels may be distinguished. Aerodynamic tunnels are also used in wind erosion research by the Institute of Deserts of the Academy of Sciences in the Turkmen Soviet Socialist Republic in Ashkhabad.

In Czechoslovakia, a suction tunnel powered by an electric motor has been used. This consisted of a deflation chamber, a filtration chamber, and a ventilator. A sample measuring 1 × 1 m was placed in the tunnel, and the wind velocity could be regulated within the range 2 to 20 m per second. The time of exposure was 15 minutes (Pasák 1966).

The aerodynamic method has also been used in research on the effects of shelterbelts (Smolař 1955) and in research on the effectiveness of avalanche prevention by the control of snow deposition (Blahout and Pacl 1965). Whereas

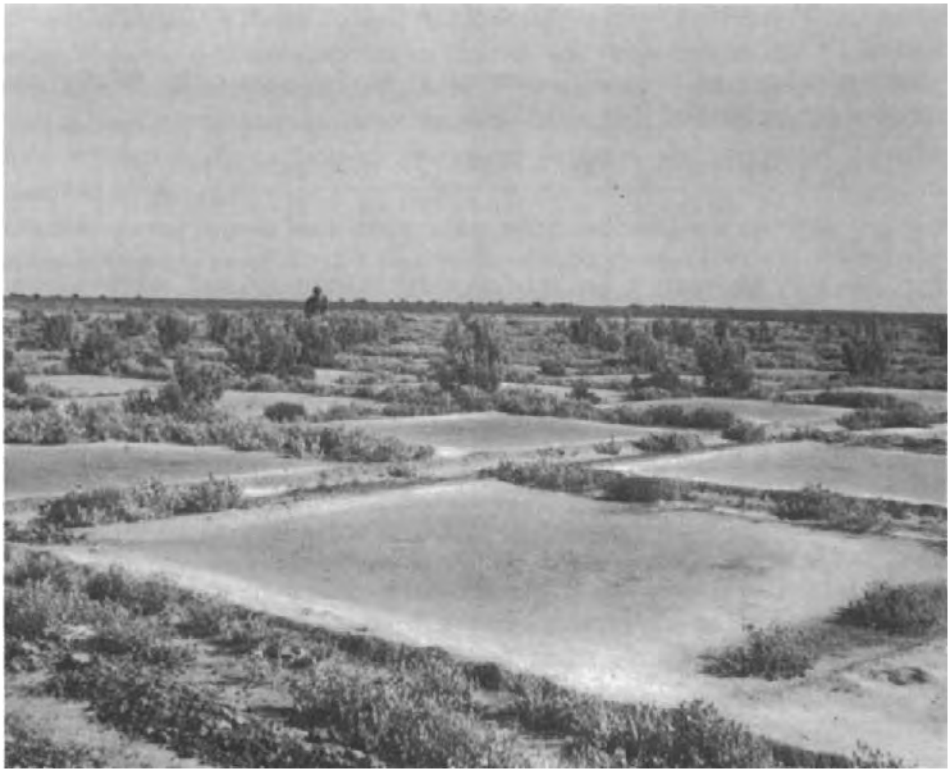


Fig. 74. Trials in the reclamation of a wind-eroded area by the checker-board system of furrowing. (By courtesy of Soil Conservation Service of N.S.W., Australia.)

deflametric methods are suitable for the quantitative and qualitative study of the effects of deflation on the soil, tunnel methods, being similar to monolithic methods, are used for the determination of aeolian soil erodibility.

Very important also is research on the build-up of soil resistance, or reduction of soil erodibility by various measures, such as making adjustments to the properties of the soil or making correct economic use of eroded soil (Fig. 74).

3.3.8.2 Research on soil erodedness

The second group of pedological methods deals with the investigation of the *overall effect of erosion on the soil,* and the determination of soil damage caused by water and wind erosion. Thus the extent of soil erosion caused by *pluviation, aeolization,* and other erosion processes is the subject of research. As has been mentioned already, these erosion phenomena proceed according to quite different time scales, and the method has to be chosen according to the effect that is to be investigated. It is an advantage, if possible, to compare results obtained from investigating the same process at different times, so that more general information is gained on the erosion process. Of considerable importance are the results of analyses of active phenomena from which secular phenomena may also be explained.

If it is not possible to observe the recent effects of erosion on the soil, research into secular effects (which is by and large the main objective in erosion research) may be supplemented by the simulation of erosion processes on the investigated soil using pluviosimulation or monolithic methods, or in the case of deflation, artificial wind force on the soil, and in this way qualitative changes caused by erosion under different conditions may be determined.

Notwithstanding the many snags inherent in pedological methods, and the unsatisfactorily slow advances made in this area, the investigation of soil as a substrate remains the main objective of erodological research. This is so because the final goal of all erosion research is to conserve soil, to combat erosion and to improve the soil; and these goals can only be achieved it appropriate information about the properties of the soil and the relationships between these and erosion is available. The purpose of pedological methods of measuring the extent of soil erosion is not only to investigate the current degree of soil damage but also to study the tendencies of erosion development and the possibility of changing erosion conditions by the application of erosion control measures. It would be desirable if at some time in the future the pedological methods concerned with soil typology could be unified with the assessment of the ecological value (fertility) of the soil by pedoerosion methods.

Among the many methods used in erodological research with their various modifications, soil indicator methods, pedogenic methods, and extensive comparative methods may be distinguished.

Methods using soil indicators

Many authors have used the *soil indicator methods* in investigations of the degree of erodedness of soil. This approach was used, for example, by Kornev (1937), Konova (1937), Sobolev (1948), Mamaev (1954), Ionova (1956), Bennett (1939), Kuron (1947, 1954), Jung (1953, 1954, 1956), Kuron and Jung (1958), Żiemnicki (1949), Ostromecki (1950), Dobrzanski and Zbysław (1955), Mattyasovski (1957), Gračanin (1962) and in Czechoslovakia by Dvořák (1955), Holý (1955), Janáč (1958), Zachar (1958, 1960, 1970), Midriak (1965), Košťálik (1965), and others.

The main objective in the use of the *soil indicator method* is to determine, using basic or modified methods, the effect of erosion on the soil and to express these changes in terms of pedological indicators. Moreover, attempts have been made to use various indicators for assessing the degree of damage caused by erosion, and even for determining erosion intensity. Some authors have classified soils on the basis of analyses of eroded soil, dividing them into different categories according to the level of threat caused to the soil by erosion, and the degree of urgency of erosion control measures.

The problems arising from these methods were tackled very effectively by the German pedologists Kuron and Jung. According to them, the course that erosion will take on slopes is best indicated by the levels of *humus, nitrogen, potassium,* and *phosphorus* in the soil, as well as by the *granulometric texture* of the soil, in some cases the *calcium carbonate* content and the *rate of infiltration* have also been taken into consideration. Detailed investigations of these properties carried out on soil profiles located along the lines of steepest inclination on various parts of the relief make it possible to predict the likely course of erosion and deposit of sediments on the slopes. Areas damaged to different degrees by erosion and by the selective grading of transported material are carefully defined, together with areas of temporarily deposited sediments, and accumulation areas. With respect to these different areas (for an account of their occurrence in the USA see Bennett 1955), individual site indices are established on the basis of soil yield, which on slopes is determined mainly by soil removal (Jung 1954). The level of danger to the soil by possible erosion can be classified, according to the above authors, as follows:

1st degree – erosion danger slight or nil; soil conservation measures required only to a limited extent,

2nd degree – erosion danger moderate; soil conservation crops should be inserted in the crop-rotation,

3rd degree – increased danger from erosion; in addition to soil conservation crops, contour ploughing should be introduced, and ploughing up and down the slope should cease,

4th degree – high degree of danger from erosion; cultivation and other techniques of soil conservation are required, together with the introduction of a soil conservation crop-rotation,

5th degree – very great danger from erosion; soil utilization is only possible with the cultivation of perennial crops or forests.

The degree of danger can be outlined on a contour map. An advantage of the method is that it takes account of the required conservation measures, but on the other hand it is laborious and better suited for the detailed survey of small territories. A weak point of the method is the fact that in spite of the detailed pedological data collected, the final assessment of the degree of erosion danger is more or less a subjective one.

Kuron's, Jung's, and especially Bennett's recognized levels of erosion danger to the soil, and the assessment of erosion danger and the degree of urgency of erosion control measures, all require in additon to the soil indicators mentioned, knowledge of a number of other soil properties, these methods belonging more to the subject of the classification of eroded land and territory. Nevertheless, the basis of these methods is the classification of soils (including non-eroded soils) by means of soil indicators.

In Czechoslovakia, Dvořák (1955) proposed that erosion be assessed by the so-called k index, which is the ratio of the granular fraction of the eroded soil to the granular fraction of the same soil before it was eroded. He established experimentally that at "a lower level of erosion" only the first and second fractions of the fine earth (granulation measured according to Kopecký's method) are perceptibly changed, whereas at "a higher level of erosion" the washing of all fractions occurs, but mostly the first, second and third fractions are affected; the fourth fraction, according to Dvořák, is affected less at "the higher level" and only the finer fractions from the skeleton are influenced. The borderline between "lower" and "higher" erosion, according to Dvořák, is represented by the sum total of the k indices for the first, second and third fractions, given by the values $(0.77 + 0.87 + 0.96) = 2.6$. At "lower levels" this sum total will be higher than 2.6, and at "higher levels" it will be smaller than 2.6.

According to the author's findings, Dvořák's method can be used to advantage for coarsely grained soils within one type. The selective process involved in the method may be rather different under different conditions.

Like Dvořák, also Holý (1955) analyzed sheet erosion in terms of changes in the texture of the topsoil, and suggested that the intensity of water erosion be assessed by the so-called "maximum relative change of texture" which "characterizes the degree of intensity of water erosion". The characteristic is calculated as follows

$$\mu = \frac{1}{R_1} m_1 + \frac{1}{R_2} m_2 + \ldots + \frac{1}{R_s} \cdot m_s ,$$

where R_1, R_2 etc. are so-called "common radii" of grains of fractions 1 and 2 etc., and m_1, m_2 etc. are the percentage weights of the respective fractions. This means that Holý introduces into the calculation the reciprocal value of the "common radius" R, instead of the simple ratios of weight percentages.

It appears that the search for mathematical relationships between erosion and granulation encounters certain difficulties arising from the fact that soil granulation varies, both with the changing geological features of the territory, and with the origin of detritus. Since eroded soils vary with respect to origin and quality (e.g. eluvial, deluvial, colluvial, aeolian soils), being formed on slopes and substrata which in the past may have been sorted and selected by erosion, transport, and sedimentation, the use of only one indicator as a criterion for assessing soil erodedness, important as it may be, is very difficult.

Therefore in most methods several indicators are taken into consideration. The most important of these are the change in *soil granulation,* the change in the *water regime,* the change of *nutrient reserves,* and finally, the change in *soil fertility.* It should be stressed that these changes are not always directly proportional to the erosion intensity, hence the use of soil changes as indicators of erosion intensity is generally almost impossible. It seems that for any degree of damage caused by erosion on one particular soil type there is possible a corresponding range of changes that will differ widely under different conditions.

Pedogenic methods

Unlike the previously discussed methods, *pedogenic methods* of classifying eroded soils are based mainly on genetic soil profiles. When the serial arrangement, expressiveness, and thickness of various horizons in non-eroded soils are known, it is possible to determine the thickness of the layer of soil that has been removed in the erosion remnants of an eroded soil profile. In this way the total quantity of soil removed over a long period can be determined. It goes without saying that the pedogenic method of determining the degree of soil erosion is based on a detailed analysis of soil types, and is particularly appropriate for deep soils with a "normal" profile. By establishing the relationship between the degree of erosion and the slope inclination for various types of soil and relief, it is relatively easy to mark various degrees of soil erosion damage with accuracy on a map showing the relief of the territory.

With these methods, too, problems arise and thus when they are applied to soil on sloping ground, firstly because of the heterogeneity of the underlaying strata and the slope material, as Šály (1974) correctly pointed out, and secondly because of the considerable changes brought about by crop cultivation in the stratigraphy of genetic horizons; the latter effect, although recent, affects the soil at ever increasing depths.

The classification of eroded soil on a genetic basis has been studied in detail mainly in the Soviet Union, where the prevailing natural conditions are very unfavourable for applying these methods. For general information on the subject, a survey of variants of the methods proposed at different times is given in Table 21.

The table shows that the pedogenic classification is based mainly on the upper horizon, which tends to be relatively thin in Czechoslovakia, its conservation being reduced to the task of preserving the soil's fertility. More detailed classifications distinguish separate subhorizons which differ in their fertility. A similar classification with a genetic basis is being introduced in Hungary (Mattyasovsky 1957, Stefanovitz 1964), in Bulgaria (Milchev and Andonov 1957), and in Czechoslovakia (Košťálik 1965, Bedrna 1974). A more detailed account of the pedogenic method of classifying eroded soils is given in the previous chapter.

An advantage of pedogenic methods is that they make it possible to determine erosion losses, especially on well-expressed profiles, and thus the patterns of erosion development on different parts of the slope can be compared; in addition,

Table 21. Survey of pedogenic methods of classifying eroded soils

Author of classification	Year of publication	Soil type for which the classification was proposed	Number of grades	Main criteria and characteristics of classification
Kasatkin	1937	Podzol soils	3	Colour, humus content, and other properties
Kostyuchenko	1937	General	3	% of removal of A hor., humus content
Pankov	1938	General	6	Degree of removal of A hor. and B hor.
Sobolev	1939	Chernozem, podzol and similar soils	5	Degree of removal of A hor. and B hor.
	1946	Ditto		
	1948	All soils	3 (4)	Degree of removal of A hor. and B hor.
Shaposhnikov	1947	Chernozem	3	% of removal of A hor.
Kozmenko	1948	Gray forest and chernozem soil	4	% of removal of A hor.
Sil'vestrov	1949	Chernozem	3	Part of removed A hor.
Lidov	1956	General	4	% removal of A hor.
Surmach	1954	Gray forest and chernozem soil	5	% and thickness of $(A_1 + A_2)$, and $(A + B_1)$ horizons
Naumov	1955	Chestnut soils and chernozem	4	Ditto + humus content
Presnyakova	1956	Medium podzols and chernozem	4 (8)	Set of selected characteristics

these methods allow forecasts to be made of likely erosion damage in terms of denudation of the illuvial horizon or of the bedrock. By detailed study of pedogenic changes it has been established that erosion not only affects soil properties and soil fertility, but also changes the soil from one type to another, degrading it from higher to lower quality, until destruction of the soil is complete and a *soil wreck* or *debris* remains. In this sense we may speak of *erosion degradation of the soil.*

There are certain difficulties in the application of these methods to shallow soils (e.g. rendzina and mountain soils), or heavily eroded soils where a comparative etalon that has not been affected by erosion cannot be found. One of the chief problems is to distinguish between recent and fossil changes in soil catenae and coronas; mistakes are made in pedogenic methods, mainly as a result of comparing differently placed links of soil catenae. Thus, for example, eroded soil on a southern slope may be compared with non-eroded soil on a northern slope.

Comparative methods

Comparative methods involve mutual morphological comparisons of the state of the soil in terms of thickness of soil cover, particle content, colour, response to acute erosion, etc. By comparing soils eroded differently, the effects of the relief, vegetation, cultivation, the type of geological substratum, conservation measures, etc. are gradually excluded, so that finally it is possible to determine the so-called *critical inclination of the slope,* the inclination at which acute erosion begins in forests, and on grassland, fields, and roads. By mapping areas of a particular inclination on a map showing the steepness of slopes and drawing in the lines connecting points of critical slope inclination, information may be obtained on the susceptibility of soil to erosion under various conditions, and in different geomorphological regions and soil types. This susceptibility will, of course, be considerably influenced by the state of erosion at the time of observation.

The comparative method is very quick and well-suited for field work, surveying, and for mapping operations. It requires, of course, some experience on the part of the observer and the definitions of criteria must be precise. The greatest disadvantage is that it does not permit the direct determination of erosion intensity. This method was used successfully by Wandel (1950) for making a comparison between soil erosion on forest land and erosion on agricultural land in the northern Rheinland, also by Grosse (1950, 1951) for erosion mapping in the GFR, by Stefanovitz (1964) in Hungary, and by Schultze (1952) in the course of soil erosion research in Thüringen.

Grosse (1950), in assessing the degree of soil damage caused by erosion, based his work on the assumption that all forms of erosion (i.e. sheet, rill, and gully erosion, according to Grosse) ultimately lead to a reduction in soil thickness over the affected area, and that consequently, erosion damage too, may be assessed in terms of area. For the purposes of mapping, which was based on the results of comparative research, he used the following classification of eroded soil:

1st degree – less than one tenth of the soil profile removed, damage not observed, inclination small,

2nd degree – from one tenth to one third of the profile removed, moderate deposits of humus and fine soil particles occurring in sheltered depressions,

3rd degree – from one third to two thirds of the soil profile removed, on the upper part of the slope heavy deposits in sheltered depressions,

4th degree – over two thirds of the soil profile removed by erosion,

5th degree – bedrock becoming exposed on slopes, deep layers of deposits building up in depressions,

6th degree – soil almost completely removed, surface rocks appearing on the slopes, with heavy deposits at the foot of slopes.

Stefanovitz (1964), investigating the extent of soil erosion by the comparative method, used a three-category classification of soils damaged by erosion:

1st degree – up to 30% of the soil profile removed,

2nd degree – from 30 to 70% of the soil profile removed,

3rd degree – over 70% of the soil profile removed.

A scale of three categories was also used in the classification of territory. A low degree of erosion of a territory, according to Stefanovitz, corresponds to 10% erosion of the soil, the intermediate degree corresponds to soil erosion from 10 to 30% and the highest degree over 30% soil erosion.

Schultze (1952) mapped erosion in Thüringen, designating it as land affected by acute erosion, all territory with an erosion intensity exceeding 0.4 mm year^{-1}; the following four degrees of susceptibility of a territory to erosion were distinguished:

1. susceptibility low or nil, damage local,
2. susceptibility moderate, damage occurring over 5% of the territory,
3. susceptibility high, locally heavy damage occurring on 10 to 15% of the territory,
4. susceptibility very high, active erosion leading to badland, about 20% of the territory affected.

Because in many instances Schultze did not succeed in determining erosion intensity, either directly or indirectly, he used a geomorphological assessment of soil erosion to establish the critical inclination. Schultze identified susceptibility to erosion with real damage to the soil from erosion; this generalization has no proper basis, and can only be accepted in making rough estimates of erosion intensity.

Mizerov (1966) used the comparative method successfully in soil erosion research in the southern regions of the USSR, Far East and on the island of Sakhalin. Comparing the thickness of the soil cover on virgin land and on ploughed land, he was able to establish with a high degree of accuracy soil losses caused by accelerated erosion. It was fortuitous that he made his investigation in regions where virgin land had been freshly cleared, and where the timing and type of cultivation carried out over the past few decades were known. Thus the intensity of erosion could be calculated, and the data obtained in this way were very valuable. Mizerov referred to the method as the comparative method otherwise known, according to Pankov, as the *historical,* or *historiocomparative method.*

With respect to these proposed terms, the author observes that the comparative method is concerned only with the determination of overall losses, or with changes in those soil properties which are usually assessed visually. In the historiocomparative method, knowledge of the timing of events is essential to the determination of erosion intensity. Consequently, the method used by Mizerov is more appropriately referred to as a historiocomparative method.

For the sake of completeness, it must be added that various *eluometric methods* are being used in intrasoil erosion research, involving the use of lysimeters. These methods are fruitful, but unfortunately they have been little used so far in erosion research.

Besides the classical pedological methods some modern radioisotope methods are also in use. One of the first attempts at the autoradiographic labelling of soil components involved in erosion and solifluction processes was made in Hungary by Kazó and Grubner (1960, 1962). Woodrige (1965) used a $^{59}FeCl_3$ solution along a contour line, and observed the movement of labelled soil particles by means of Geiger–Müller tubes and scintillation counters. Nowland (1962) used radioisotopes in an investigation of concentrated runoff. In addition to these authors, Kandil (1966), and Coutts and Tinsley (1970) in Great Britain investigated the movement of particles in soil erosion using ^{59}Fe; (^{59}Fe has a half-life of 46 days and emits easily detectable gamma radiation). These methods are also applicable to the study of intrasoil erosion and underground erosion processes in general.

3.3.9 Hydrological methods

Hydrological methods are of particular interest where it is desired to determine the intensity of soil erosion in a defined region or catchment area. These methods are similar to deluometric methods with the difference that it is not deluates or the products of precipitation erosion that are observed, but rather the overall effect of precipitation erosion and river erosion together is assessed. Since in the transport of solid matter by rivers all kinds of denudation products combine, the determination of erosion intensity by measuring the flow of silt and bed load meets with considerable difficulty. Despite this, hydrological methods provide an important means of studying erosion (Fig. 75).

Hydrological methods have been used by quite a number of authors. As far as we know, the Chinese were the first to make measurements of silt. Pan having studied

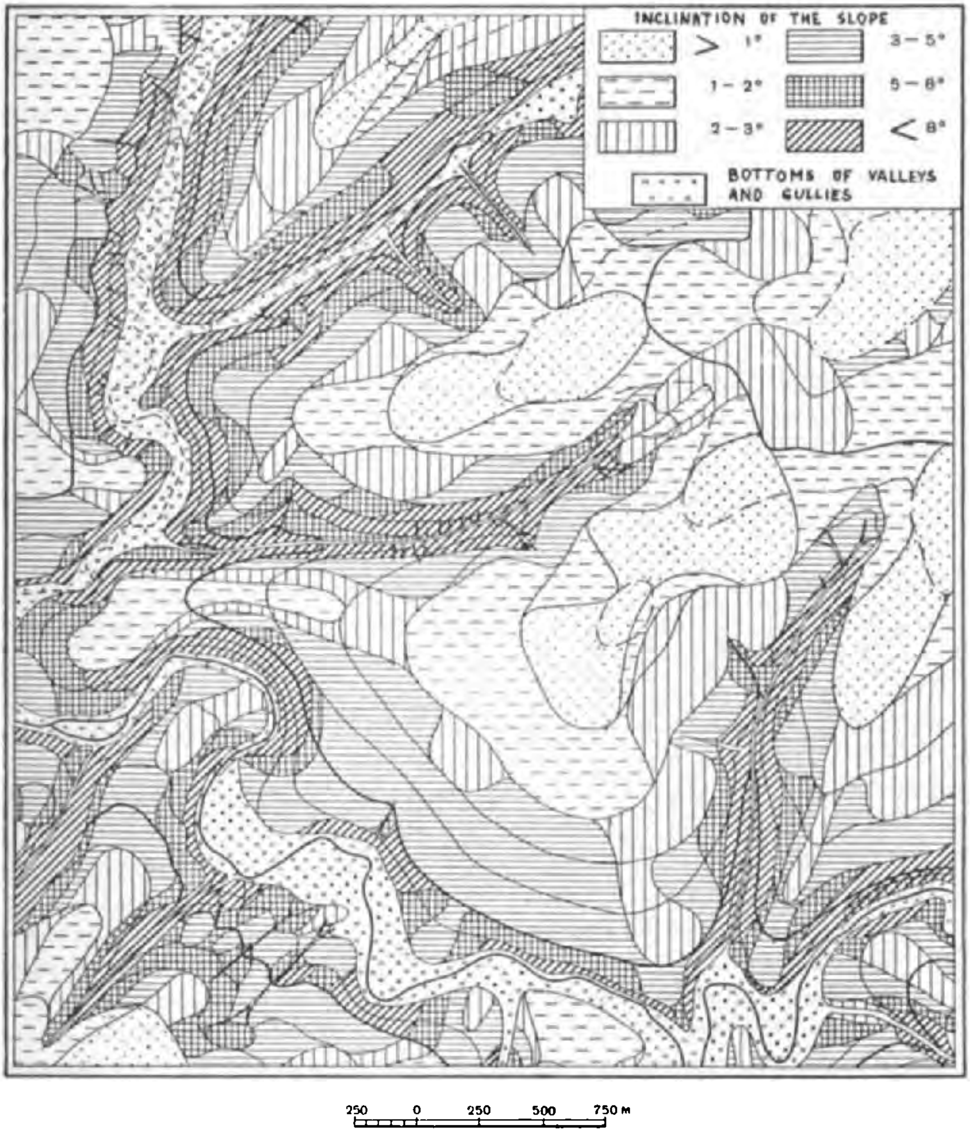


Fig. 75. Plan showing gradients on an experimental plot on the Novosil' Research Station (according to Lidov et al. 1956).

these problems for 26 years (1565–1591) in the Yellow River basin (according to Szolgay 1960). Systematic observations of silt flow started in 1911 when permanent monitoring stations were set up in Russia. Today over 700 stations in the Soviet Union are engaged in recording data which make possible a thorough study of the origin of silt and how this is related to soil erosion. The largest publications relating to this kind of work are Lopatin's (1952) and Shamov's (1954) monographs. A major contribution towards developing the methodology of this subject was made by Polyakov (1940). Interesting studies involving hydrological methods of erosion research were made by Hénin et al. (1954), Tixeront and Berkaloff (1954), Glymph (1954), and others.

In Czechoslovakia, the first interpretation of silt measurements was made by Smetana (1957) and Dub (1954, 1955). In Slovakia, systematic observations began in 1952 and the results were evaluated in publications by Almer (1955), Szolgay and Náther (Szolgay and Náther 1954; Náther and Szolgay 1958; Szolgay 1960), Hampl and Koščo (1961), and others.

The quantity of silt is recorded by taking samples of water and pouring them into containers called *bathometers* in which the turbidity of the water is measured; from this, and from measurements of the flow rate of the water, the total flow of silt can be calculated. This is the *bathometric method,* which is essentially a *turbidimetric method* of assessing erosion intensity from water turbidity. However, most bathometric studies of silt are carried out for the limited purpose of assessing sedimentation in watercourses and water reservoirs, and therefore the method is adapted for this purpose so that only silt deposited in sedimentation cylinders during periods of 24 hours is taken into consideration. Data obtained in this way may represent as little as one half of the total flow of silt.

More up-to-date methods are based on modern techniques of measuring silt flow, such as *photoelectric* devices which continuously record changes in the total flow of silt by means of one or several photocells placed under the water. The most modern device of this kind is the French turbidimeter. Shvebts (1957) investigated the rate of removal of soil from runoff plots by means of a photoelectric turbidimeter. Finally, attemps have also been made to measure water turbidity by means of *radioisotopes* (Szolgay 1960).

In Czechoslovakia, the first attempt to establish the intensity of erosion from silt data was made by Dub (1955). His suggested procedure for assessing *erosion intensity from the flow of silt* is based on the method of Polyakov (1940), who considered that the flow of silt *(H)* is a function of the water flow *(W),* the mean slope of the watercourse *(I),* and of the erosion coefficient *(A);* the mean annual turbidity ϱ per m^3 of water $= H/W$. Polyakov expressed the erosion coefficient in terms of the relationship

$$A = \frac{\varrho}{k\sqrt{I}} \,,$$

where $k = 10^4$. For calculating the erosion intensity, *E,* in any region, Polyakov derived the equation

$$E \doteq H = AkIW \; [\mathrm{t \cdot ha^{-1}\ year^{-1}}] \,.$$

Contrarily to Polyakov, Dub proposed an erosion coefficient, C, which relates to various local conditions and which is expressed by

$$C = \frac{g}{k l^n} ,$$

where k is the constant of proportionaly ($k = 10^4$), l the specific work of water in kW per km^2 calculated from the specific runoff and the corresponding depth of the erosion base, $n < 1$, and g the amount of material carried in tons per km^2, calculated from $g = G/F$, where G is the quantity of silt discharged from the catchment area, and F the surface area of the catchment area.

In other words, if the average flow of silt, the inclination of the watercourse, and the quantity of water are known, the intensity of erosion may be established indirectly, using a proportionality factor. Polyakov assumes here that there is a direct relationship between water discharge and the discharge of silt. Dub assumes this relationship to exist between the creation of silt and the *work done by the flowing water.*

It has been established that these relationships are valid only for measurements taken in large streams over long time periods, and at the same measuring place, whereas for small watercourses they are less likely to hold true. Let us consider the situation in which the water-flow causes erosion only in the upper part of the catchment area, while in the sedimentation region the quantity of silt derived from the upper region gradually declines. This decline is particularly pronounced where the water causes an area to be flooded, and silt is deposited both in alluvia and on the river-bed. Examples of silt decline in river water are cited by Szolgay and Náther (1954), who refer to measurements made by the Hydrological Research Institute in Budapest of amounts of silt carried by the Danube at different points along its direction of flow:

Vienna	540,000 m^3 year^{-1}
Bratislava	370,000 m^3 year^{-1}
Palkovičovo	120,000 m^3 year^{-1}
Komárno	38,000 m^3 year^{-1}
Nagymaros	15,000 m^3 year^{-1}

This means that even the most accurate measurements of silt flow will be affected considerably by the selection of the measuring site on the watercourse.

The main difficulty encountered in this method is in finding the nature of the relationship between the quantity of silt flowing and the quantity of soil displaced by erosion.

This is an indication of the need for further development and refinement in hydrological methods, so that the determination of erosion intensity in the catchment area from silt flow data becomes possible. It would thus seem necessary to establish, by making special measurements, the relationship between the contents of the bed load, the silt flow, and dissolved matter in different sectors of watercourses and at different flow rates; it is also necessary to obtain data from long-term observations, and to define with greater precision detailed parameters expressing natural and economic factors, and the conditions that affect erosion in the catchment area.

A further hydrological method for the determination of erosion intensity involves measurement of the *quantity of deposits* intercepted in the retention spaces of dams or barriers that have been built across watercourses to create reservoirs and ponds. The precision of the method may be increased by establishing the turbidity of the flowing water. In fact, this is a modified form of the volumetric method, the quantity of soil that is eroded in the catchment area and the hydrographic network usually being observed over a long period. The result is influenced mainly by the morphology and hydrological structure of the catchment area. The use of these methods is well illustrated by research carried out by Dvořák (1962), Veselý (1964), and others. The methods depend on information on the increase of the erosive effect of surface runoff as slope length increases, as determined by volumetric, deluometric, pluviosimulation, and other methods.

3.3.10 Vegetation methods

The previously described methods are closely related to the *vegetation method,* by which determinations of erosion intensity, the effect of erosion on the soil (especially on soil fertility), and the protective effects of vegetation under various conditions can be made.

Erosion intensity or the *accumulation of deposits* may be investigated by the vegetation method, mainly in those cases in which the stand or crop plant protects the soil sufficiently to provide a suitable comparison with soil surface changes in the surroundings. Mature and older trees are best suited for this purpose. In the author's research, the conditions of soil movement under forest trees were used as a control for the determination of erosion removal on pastures and for the measurement of the rate of deposit accumulation on the lower stretches of slopes (Zachar 1970) (Fig. 76). Forested land was used as a measure of the original state of the soil mantle by Wandel (1950) and Midriak (1969). Sobolev (1945, 1948), in an investigation of the intensity of wind erosion, used erosion remnants protected by vegetation for the purposes of comparison. The thickness of the eroded layer is shown on Fig. 42.

Vegetation can be used to even greater advantage in research on the effects of *erosion on soil properties,* especially those relating to the *decline of soil fertility.* In this type of erosion problem the vegetation method is almost irreplaceable because

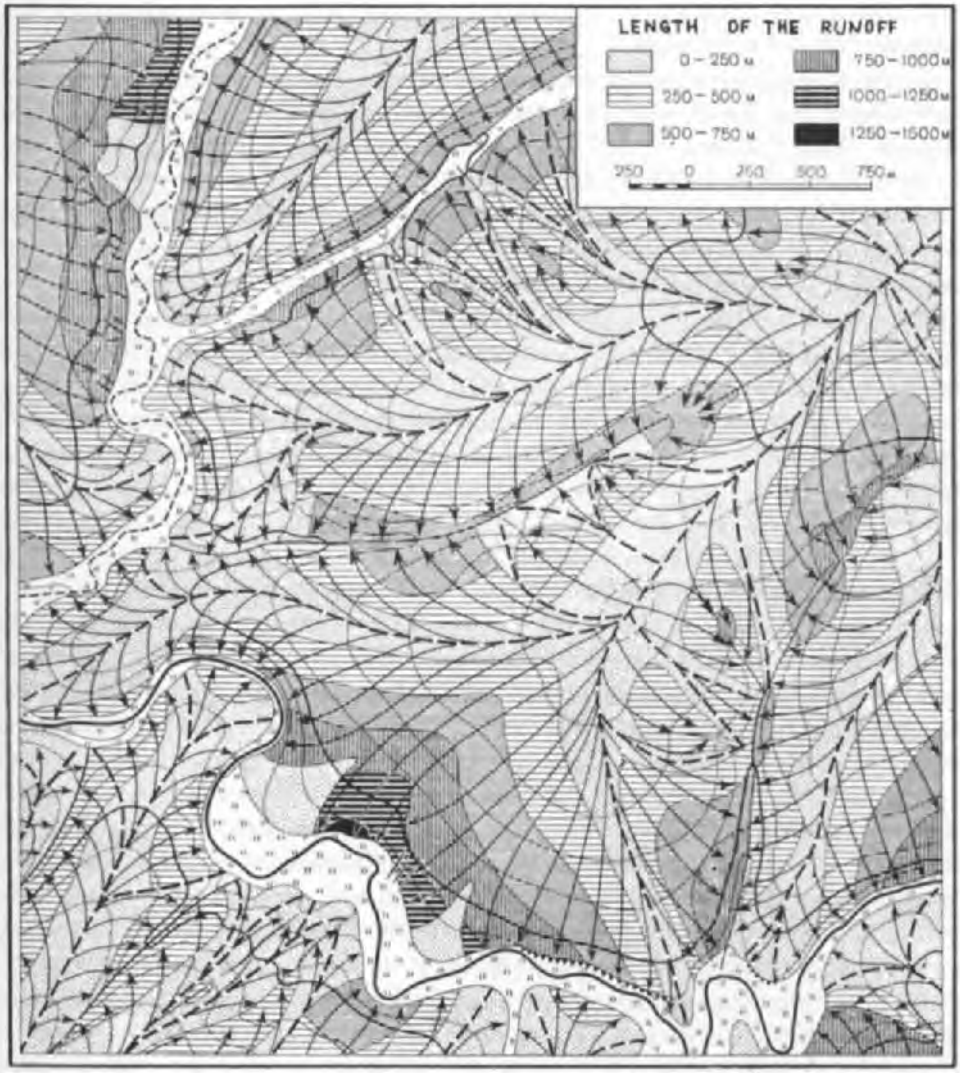


Fig. 76. Plan showing lengths of runoff lines on the Novosil' experimental plot (according to Lidov et al. 1956).

no pedological method can reflect changes in the soil with respect to its ecological value as comprehensively as the plant itself. Vegetation has been used to monitor both seasonal and long-term changes, as well as changes that were predicted by other means. Of particular interest are studies which inquire into the relationship between erosion and plant growth and how this is disturbed by grazing and other human interference. Valuable theoretical research going beyond the scope of the subject of erosion has been carried out, in which the growth and development of crop plants was observed on soil originating from different genetic horizons that became exposed by erosion.

Of the many studies concerned with the effects of erosion on *soil fertility* and *agricultural crop yields,* mention should be made of the work done by Braude and Gussak (1938), Kornev (1937), Konova (1937), Klepinin (1937), Kozmenko (1949), Shaposhnikov (1940), the important contributions made by Presnyakova (1948, 1953), Antropov (1957), Tikhonov (1958), Jung (1953, 1956), and the monograph of Bennett (1939, 1955). The fertility of the various genetic soil horizons, and the application of the vegetation method to erosion problems were studied by Gedroïts (1909), Sinclair and Sampson (1931), Sabinin et al. (1936), Kirsanov (1936), Mosolov (1937), Presnyakova (1953), Latham (1940), and others. Finally, the relationships between soil erosion and different kinds of plant association were studied by Yakovlev (1940), Shalyt (1949) Semenova-Tyan'-Shan'skaya (1949, 1951), Nadezhdina (1956), Tkachenko (1956), Zapryagaeva (1964), and in Czechoslovakia by Šmarda (1964) and Midriak (1972), and others.

3.3.11 Historical methods

Historical methods are based on various records, maps, and other documents giving evidence of changes in the surface relief, or of soil movements caused by erosion. *Paleontological studies* have not been widely used as yet for the purposes of erosion research.

A good example of the use of the historical method in research on sheet erosion is cited by the Polish scientists Bac (1928) and Hlibowicki (1955), who assessed the intensity of erosion by the changes that had taken place in the microrelief since the latter had been recorded in accurately surveyed contour maps; the survery was carried out after 19, and 45 years, respectively. Żiemnicki (1949) determined erosion intensity by measuring the thickness of slope sediments deposited over 28 years in the surroundings of a small church. Rozov (1927), Kozmenko (1949), Sobolev (1948), and others have used historical data in gully erosion research. In Czechoslovakia, Láznička (1959) used historical information for assessing soil erosion in the Brno region, and Zachar (1960, 1970) used this method in gully erosion research in the Rakovník region. In the English literature, Bennett (1939) describes one of the most detailed historical approaches to erosion research.

Interesting results were obtained by Karl (1970) who used the historical method to investigate periglacial valley deposits along the northern border of the Alps. He found that the displacement of glacial deposits had intensified in the latter 150 years as a consequence of glacier recession and accelerated erosion. The sources of the deposits were growing exponentially and the eroded surface increased threefold during the 150-year period. The present rate of displacement of deposits in the Halblech region is approximately 60,000 m^3 $year^{-1}$ these deposits originating over an area of about 50 km^2. The creation of these deposits is largely associated with glacial erosion, followed by water erosion.

Ložek (1963) successfully applied the historical method in soil erosion research of the Holocene.

3.3.12 Morphometric methods

The *morphometric* or *geomorphometric methods* involve the investigation of erosion phenomena by means of morphometric factors, such as the inclination, length, aspect, and shape of the slope, the form of the relief, the depth of the erosion base, the form, length, activity, and density of erosion gullies, the proportion of ploughed land, etc. By these methods the nature of the relief – an important factor governing the activity of exogenous erosion factors – may be studied. If the relationships between the relief and erosion taking place within various formations are known, morphometric data may serve as an excellent basis for investigating the distribution of erosion over a given territory.

Morphometric data were used in erosion research by Sobolev, who with his colleagues prepared a whole series of maps for the study of erosion. These maps show the different types of erosion formation occurring on the relief, the length of the network of gullies, the distribution of erosion formations over the territory, the depths of local erosion bases, and the mean angles of inclination of the surface; all these factors have been surveyed in the European part of the USSR (Sobolev 1948) and provide a basis for the preparation of a *map of the USSR* showing the *distribution of erosion phenomena.*

In a similar way Lidov and colleagues (Lidov and Setunskaya 1959) used special maps in an analysis of the relief of small territories. They successfully used *maps showing the inclination* (Fig. 77) and the aspect of the land, *maps showing runoff lines,* and *maps showing the accumulation of surface runoff* (Fig. 78). By looking at the degree of washing and the distribution of gully erosion in relation to various elements of the relief, interesting relationship were observed which gave a deeper understanding of the more regular aspects of the development of erosion phenomena.

A method devised by Sil'vestrov (1955) also belongs to this group of methods; it is based on the experience gained by Kozmenko in developing a method for the determination of the so-called *erosion coefficient,* given by the following expression

$$E = \frac{HRS}{10\sqrt[4]{P}}$$

where E is the erosion coefficient, H the depth of the erosion base [m], R the configuration of the catchment area expressed in terms of the density of the

Fig. 77. Measurement of water runoff and silt sedimentation on an experimental plot. (By courtesy of U. S. Forest Service.)

Table 22. Relation between erosion and erosion coefficient according to Sil'vestrov (1955)

Degree of danger from erosion	Characteristics of erosion	Erosion coefficient
1	Territory almost unaffected by erosion, neither sheet nor rill erosion visible	<0.20
2	Territory affected by slight erosion, sheet erosion slight on exposed sunny aspects, rills not developing	0.20—0.49
3	Territory affected by moderate erosion, sheet erosion occuring over the whole territory, more intensively on sunny slopes where rills begin to form	0.50—0.99
4	Territory badly affected by erosion, locally severe on all slopes with moderate development of rills	1.00—1.41
5	Territory affected by heavy erosion, both sheet and rill erosion widely distributed	>1.50

Fig. 78. The stem of a walnut tree *(Juglans regia)* buried by deposits gives a relatively reliable indication of the period of sediment accumulation and intensity of erosion in the catchment area of the gully (Tematín Hills, Czechoslovakia). (Photo D. Zachar.)

hydrographic network [km km^{-2}], S the cultivation coefficient (the ratio of the area of tilled land to the entire surface of the catchment area), and P the surface of the catchment area [ha]. It has been established empirically that for steppe and forest-steppe regions there exists a relationship (Table 22) between the degree of danger threatened by erosion and the "erosion coefficient".

A similar method was used by Kozlík (1958) who suggested that the erosion susceptibility of a territory could be assessed from the average angle of inclination.

The essence of the method is the determination of the relationship between the angle of inclination and the erosion intensity.

As well as being used by Kozlík (1958), morphometric methods have been used by Mazúrová (1955), Bučko (1956), Košťálik (1965), Bučko and Mazúrová (1958), Holý (1958), Midriak (1965), Gam and Stehlík (1956), Lochmann (1964), and others.

Of these investigations the collective work of Bučko et al. (1964) is the most important study of erosion distribution in Czechoslovakia, representing a summary of results obtained in different investigations. The methodological basis of morphometric research was laid in Czechoslovakia by the work of Bučko and Mazúrová (1958) on gully erosion. The essence of this work was the determination of the dimensions and inclinations of erosion gullies, and the mapping of these on a scale of 1 : 25,000, the working plot covering an area of 4 km^2. The total gully length (considered by the authors to be the most important factor) was computed for areas of 1 km^2, and from distribution curves 6 categories of gully network density were set up as follows: 1. 0.0 to 0.1 km per km^2, 2. 0.1 to 0.5 km per km^2, 3. 0.5 to 1.0 km per km^2, 4. 1.0 to 2.0 km per km^2, 5. 2.0 to 3.0 km per km^2, and 6. over 3.0 km per km^2.

In this way, Bučko and Mazúrová obtained data on the average density of gullies in different regions, in which they were then able to study the relationships between the occurrence of gully forms and such factors as the nature of the geological bedrock, the relief, the percentage forest cover, etc. They also made conclusions about the intensity of gully and sheet erosion, based on the density of gullies.

3.3.13 Photogrammetric methods

The study of erosion phenomena from either aerial or ground-level photographs has considerable advantages. Both kinds of photography may be used in the preparatory stages of terrain studies, and in research on erosion, erosion gullies, eroded soils, etc., as well as in all types of map-making. Aerial photographs have the advantage of being useful throughout the duration of the research project. Of greatest value of course are special maps which are made after the evaluation of aerial photographs and which are used in detailed studies of erosion.

Another advantage of photogrammetric methods is that they may be used provided that certain conditions for the measurement of very small formations on the soil surface, and by taking repeat photographs for identifying changes in these forms are fulfilled. All types of maps can be produced from aerial photographs;

from contour maps to maps of erosion forms. Research based on aerial photographs requires little field work, and evaluation is quick and reliable, making this technique an indispensable part of any erosion research project.

However aerial surveys are rather expensive, and cannot be undertaken unless weather conditions and seasonal factors are favourable. The best time for flying and photographing is in spring when the soil is least obscured by vegetation. Aerial photographs are most valuable in studies of heavily eroded and threatened soils in mountainous and alpine terrain with difficult access. Also the study of erosion phenomena by means of aerial photographs, and photography in general, is carried out to best advantage in cases of natural disasters, torrential downpours, floods, dust storms, etc., where a survey of the course and consequences of erosion is needed over a large area in the shortest possible time.

The recording of such occurrences in the understanding of erosion is of greater value than long-term and often costly research into less distinct erosion phenomena. Aerial photography has been used in the investigation of erosion and erosion control measures by Gorrie (1935), Troll (1939), Cooper (1942), Smith (1942), Schumann (1943), Magruder (1949), Stübner (1955), Pronicheva (1955), Vageler (1955), Steinmetz (1958), Semenova (1959, 1960, 1962), Andronikov (1959), Afanas'eva and Lidov (1962), Rasmusson (1962), Ionescu and Blegu (1963), and others.

Of particular note is the publication by Stübner (1955) who used the photogrammetric method in his research on soil erosion and erosion control measures in Thüringen. His work contains discussions of procedure, the principles of photographing, the interpretation of photographs, and research results which can be compared with those of the comparative method that was being used at about the same time by Schultze and his collaborators (1952). Stübner comes to the conclusion that all forms of erosion with the exception of hidden erosion, which is not very conspicuous, may be evaluated by aerial photography. According to Stübner, stereoscopic observation of photographs taken from a height of 330 m (enlargement × 3.5) can reveal a 10 cm long object with an image size of 0.105 mm (using a RMK camera). The resolving limit of the eye is from 0.07 to 0.21 mm.

Detailed methodological research was carried out in the mountain regions of the West Carpathians in Czechoslovakia by Midriak and Petráš (1972), who compared results obtained by *aerial photogrammetry* with data from *ground-level stereophotogrammetry*. It appeared that in the evaluation of *destructive erosion phenomena* in the *alpine belt*, the use of pictures taken by ground-level stereophotogrammetry gave higher and more accurate values, indicating the greater degree of faith that can be placed in ground-level photographs (Figs. 79, 80). A combination of the universal photogrammetric and ground-level methods is ideal, the universal method being of advantage in the detailed mapping of the territory at a scale of 1 : 10,000 or 1 : 5,000, and ground-level being suitable for the detailed evaluation of erosion phenomena at scales of between 1 : 200 and

Fig. 79. Territory of the Western Tatra Mountains (Czechoslovakia) where erosion phenomena were evaluated by ground-level photogrammetry. (By courtesy of Technical University, Bratislava.)

1 : 1,000. Results obtained by photogrammetric methods are more accurate and contain substantially more information than those obtained by the *classical geodetic methods*. What is more, in difficult terrain they are approximately three times cheaper and require less time.

Of work concerned with special problems, mention may be made of the investigations of Karl et al. (1960) into the use of aerial photographs in *torrent control*, and the work by Kuhn (1953) on the relationship between *erosion* and *terracing;* there is also the work of Stübner (1956) dealing with the *diagnostics of impending erosion damage* by means of aerial photographs, the work of Hassenpflug (1971) on the counter-deflation effect of shelterbelts, and the work of Richter (1963) on the use of aerial photography in *practical soil conservation*. The latter two authors (Hassenpflug and Richter 1972) also dealt in their publication with the interpretation of aerial photographs in the investigation of water and wind erosion (Fig. 81). In Poland Obraczka (1970) studied the problems of aerial surveying and the use of *photogrammetric mapping* as aids to the *improvement of eroded land* and *erosion control*.

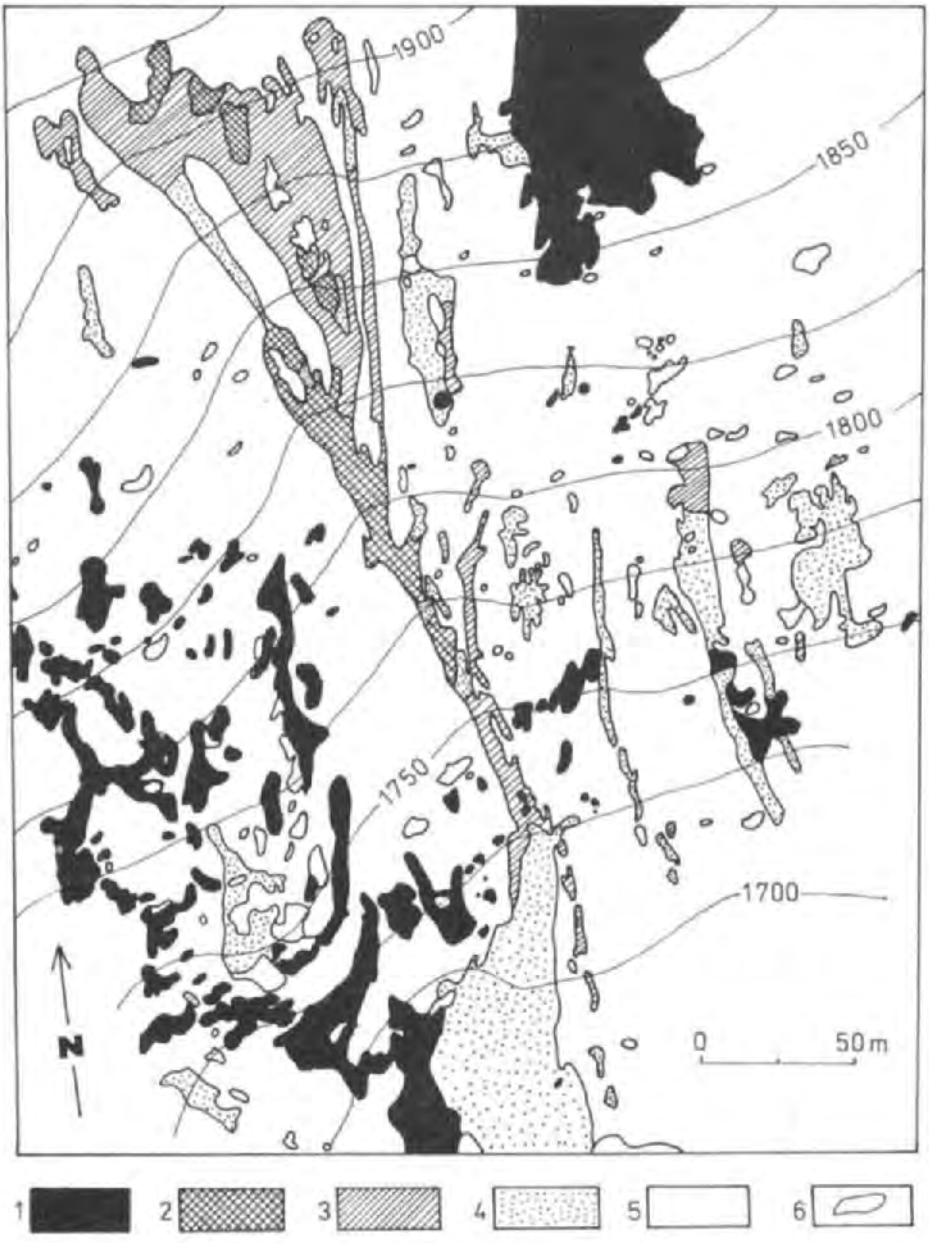


Fig. 80. Quantitative evaluation of phenomena. 1 – granitic rocks (5.87%), 2 – bedrock exposed by snow erosion (1.57%), 3 – soil severely eroded by snow abrasion (nival detersion) (5.08%), 4 – nival and pluvial deposits (6.89%), 5 – areas protected by tussocks (77.49%), 6 – area of dwarf pine (1.81%) and isolated remnants of grass (1.3%). Area severely damaged by nival erosion – about 11%.

Fig. 81. Aeolian formations to the west of Nordhackstadt (GFR) shown on an aerial photograph taken on 7th April 1969. Eroded land, accumulation formations, and the protective influence of shelterbelts can be seen. (By courtesy of Landesvermessungsamt Schleswig-Holstein.)

3.3.14 Cartography of erosion phenomena

The mapping of erosion phenomena has been mentioned already in connection with morphometric methods. The surface distribution of an erosion phenomenon can be expressed very clearly on maps using *cartographic methods*. By means of these the distribution of erosion factors and conditions, the distribution of erodibility and eroded land, the distribution of erosion gullies, the use of certain conservation measures, etc., can be represented. By superimposing different maps upon one another the dependence of erosion on various factors and conditions may also be established. For example, the map showing the occurrence of erosion may be superimposed upon maps showing slopes, aspects, vegetation types, or land use, respectively (Midriak and Zachar 1973). Relationships that come to light from the detailed study of smaller territorial units may be generalized, and maps covering larger areas may be drawn, but the degree of generalization must be kept within the scope of the data obtained because abstraction involves the multiplication of errors.

Of chief importance are maps showing the localization of *actual, potential,* and *forecasted erosion;* these maps being prepared from the results of comprehensive research, or by the use of empirical formulae (see the next section). The basic means of expression used in the preparation of *erosion maps* are lines connecting places of equal erosion intensity; such lines could be called *isoerodents.*

The first precondition for successful mapping is the definition of objectives, and secondly, a clear cartographic symbol for the investigated factor must be selected. In detailed research, the type of erosion, its form and intensity are usually indicated. Mapping becomes difficult if several types of erosion occur in various forms on the same territory, because the more symbols are represented on a map, the less easy it is to read the map. Still more difficult is the representation of the activity and intensity of erosion on a map. It is almost impossible to represent the less conspicuous yet harmful forms of erosion such as hidden erosion, the effects of which can be detected only after it has been acting over a long period. Therefore, in sheet and wind erosion, it is usually the degree of soil damage, or the susceptibility of the soil to erosion (its erodibility) that is mapped, whereas in gully erosion the density of gullies is represented.

The cartography of gullies and erosion-derived sands dates from the 17th century, when erosion phenomena were included within the scope of general mapping. According to Sobolev (1970) the first photogrammetric maps produced for the purpose of showing sand and gully phenomena were made in 1898. These maps were used in the stabilization and afforestation of land carved by gullies. The first mapping surveys in central Europe and Russia date from the 16th century.

An example of an orientation map is given by the general map of the distribution of soil erosion in the Soviet Union (scale 1 : 5,000,000), showing 5 grades of sheet erosion, 2 grades of gully erosion, 4 grades of wind erosion, a territory in which there is an accumulation of eroded material, and mountain regions, etc. The map showing gully networks in the Soviet Union was produced with a scale of 1 : 420,000 (Sobolev 1948). Spiridonov (1952) suggested the preparation of a scale 1 : 100,000 map with a grid of 4 cm^2 (representing 4 km^2), showing the distribution of erosion on the relief.

In the USA the Soil Conservation Service has produced a general map of the distribution of soil erosion, showing territory with little or no erosion, and six grades of sheet, gully, and wind erosion. Three of these grades embrace minor erosion, two grades embrace major erosion, and one grade refers to mountain regions and wasteland (Bennett 1939). Similar general maps of erosion have been prepared in Bulgaria, Romania, and other countries (Biolchev 1955, Moţoc 1963).

An interesting map of soil erosion in Thüringen was prepared by Schultze (1952), who distinguished between territories which were either susceptible or very susceptible to erosion. In addition, Schultze marked areas of actual damage on the map. His working map had a scale of 1 : 25,000 and a generalization of erosion conditions was represented on a comprehensive map (scale 1 : 500,000).

Krumsdorf and Beer (1962) consider a scale of 1 : 5,000 to be the most appropriate for the mapping of pedoerosion phenomena, as was found earlier by Hempel (1951, 1954) also. In Hungary, a soil erosion map of scale 1 : 500,000 was produced (Stefanovitz 1964).

In Czechoslovakia, erosion was mapped as a part of the State Water Management Plan on maps of scale 1 : 25,000 and 1 : 75,000 (Holý 1958), and maps showing the density of erosion gullies were produced with scales of 1 : 25,000, 1 : 200,000, and larger (Gam 1957, Gam and Stehlík 1956, Bučko and Mazúrová 1958, and others). In the author's research maps with scales of 1 : 1,000 up to 1 : 5,000 were used for more detailed investigations. Exceptionally, when preparing maps from ground-level photographs, a scale of 1 : 500 was used.

The retrieval of data from maps is carried out by various well-known methods. *Curvimeters, planimeters, templets, weighing of paper cut-outs, quadrangular templates, point grids,* and *statistical techniques* are commonly used. The application of cartographic methods is presented in detail in the publication *Voprosy metodiki pochvennoerozionnogo kartirovaniya.*

3.3.15 Empirical mathematical methods

Empirical mathematical methods form an inseparable part of any erosion research in which the erodibility of the soil, the state of erosion, the erosion intensity, the expected effects of conservation measures, and other factors that are important in the understanding of erosion processes and erosion control need to be expressed in figures. This type of research may form part of a limited, specific research project, or may be required as part of the generalization of investigated phenomena. These methods should therefore be regarded as being basic to all areas of erosion research and an important aid to practical soil conservation.

Mathematical methods are commonly used as means of expressing *existing (actual), expected (forecasted),* and *possible (potential) erosion.*

The basic entity is always the *potential erosion* which is a function of the intensity of the erosion process without the protective effects of vegetation and special land use schemes. This *potential erosion is the maximum erosion* expected to occur as a result of natural abiotic factors. According to the factor concerned, pluvial, surface, linear, channel, and aeolian potential erosion may be distinguished. Empirical formulae have recently been derived for other types of destructive action, in particular for the process of erosive solifluction. The potential erosion represents the extreme degree of erosion that can be expected on a particular area, and as such is the opposite of the natural erosion which is currently affecting a given territory, or which would occur in the absence of human interference and without the presence of domestic animals. In desert areas devoid of vegetation potential erosion and natural erosion are identical.

Actual erosion may be computed from the potential erosion by multiplying potential erosion by the coefficients of natural, biotic, and anthropogenic factors: in most cases the value of the coefficient is less than unity, and therefore actual erosion is usually of a lower value than potential erosion. Where potential erosion is reduced, this is mainly on account of the influence of vegetation, whereas animals and man can increase the potential erosion. The latter situation may arise from the mechanical compaction of the soil by cattle, or from a decline in the natural resistance of the soil to erosion, as a result of industrial contamination by fumes in artificially high accumulations of surface runoff, or from increased dynamic wind turbulence, etc. The determination of actual erosion gives an indication of the actual danger from erosion in the investigated area. The nearer the level of actual erosion to that of natural erosion, and the farther it is from the level of potential erosion, the greater is the protective effect of vegetation and man-made conservation schemes.

Actual erosion may fluctuate within wide limits between the levels of natural and potential erosion. If land utilization and applied soil conservation measures in a particular region are known, a *diagnosis* as well as a *prognosis of erosion phenomena* can be made with a view to improving soil management and reducing erosion to a *harmless,* or *tolerable level.* Expected or facultative erosion is referred to by the author as *forecast erosion.*

A special situation arises in semiarid, arid, and desert regions where natural erosion exceeds the tolerable level, and consequently, the forecast level of erosion is lower than that of natural erosion, let alone accelerated erosion. In such cases the reduction of actual erosion to a tolerable level is usually difficult and costly, although imperative from a soil conservation point of view.

The merits of mathematical methods, in which empirical formulae are derived for the purpose of calculating erosion, are obvious, and further progress in erosion research and practical soil conservation is hardly imaginable without these methods. They represent the highest degree of generalization of research information, serving as a means of gaining additional knowledge, and of planning, projecting, organizing, and implementing systems of erosion control measures; mathematical methods have also opened erosion research to the application of computer techniques.

The disadavantages of mathematical methods are their present lack of precision and the principal need for rather exacting data relating to local conditions – data which are often not available. In such cases the use of equations is limited, and they may not even be of use for making approximate calculations.

Several authors have undertaken the task of setting up universal equations for the calculation of erosion intensity. In this chapter only a short survey of the most important work carried out in this field is given. In the following chapters the calculation of various values is discussed in greater detail.

Musgrave (1947) was among the first to devise a universal equation for the calculation of erosion *intensity* and *erosion losses* in *precipitation erosion.* This equation was later modified to give greater precision of results by Wischmeier (1955), Wischmeier and Smith (1965), and other versions of it were adopted by Frewert et al. (1955), Kohnke and Bertrand (1959), and others. For the USA, the *universal soil equation* ARS (Agricultural Handbook 282, 1965) is written

$$A = RKLSCP,$$

where A is the soil loss [t acre $^{-1}$], R the rainfall erosivity index – a number which indicates the erosivity of the rain on a scale based on the EI_{30} index, K the soil erodibility factor – a number which reflects the susceptibility of a soil type to erosion, L the length factor – a ratio which expresses the soil loss relative to that from a field with a specified length of 72.6 ft (22.6 m), S the slope factor – a ratio which expresses the soil loss relative to that from a field with a specified slope (9%), C the crop management factor – a ratio which expresses the soil loss relative to that from a field under a standard cultivation treatment, P the conservation factor – a ratio which expresses the soil loss relative to that from a field deprived of conservation practices (i.e. a field ploughed up and down the steepest slope).

It is essential for the successful use of this, or indeed any other equation, that sufficiently precise parameters are available; the measurement of these parameters has been refined in various countries. In Czechoslovakia the search for erosion coefficients most closely relating to central European conditions was undertaken by Holý (1970), Pretl (1970), and Stehlík (1970, 1975a, b), and for forest soils by Míchal (1973) and Midriak (1975a, b). For the most part, modifications of Wischmeier's and Frewert's equations were proposed. In Romania local erosion parameters were established by Moţoc (1963, 1970) and the theory of water erosion was studied by Ionescu (1972). A promising approach to erosion research is the use of mathematical models of erosion processes (Holý 1970).

The most detailed and complete discussion on *calculating* and *forecasting* the effects of *water erosion* by *engineering methods* was presented by Mirtskhulava (1970). He based his calculations on the *critical velocity of rain* and *snow runoff water* at which erosion starts (many authors erroneously take only rain-water runoff into consideration). Thus erosion is calculated from its dependence on the critical velocity of water and the carrying force, in the same way as in calculations of river erosion Mirtskhulava also derived formulae for the washing of natural water channels, including their banks, a formula for establishing the intensity of gully formation, a formula for the calculation of irrigation erosion, and a formula for computing the erosion control effectiveness of conservation measures. He successfully evaluated and enriched existing theoretical information on the activity of water erosion over the whole catchment area. Many of the broader aspects of water erosion problems were also overcome successfully by Jůva and Cablík (1954), Gavrilović (1972), Riedl, Zachar et al. (1973), and others.

In a similar way mathematical methods for the calculation of the rate of *debris flow* (mudflow, aquasolifluction), its transporting capacity and other characteristics were devised. Thus Makkaveev used diffusion theory in the study of the turbulent mixing of materials, Velikanov used gravitation theory, etc. (Bogolyubova 1957). The theory of debris flow was improved with the derivation of mathematical relationships these being developed mainly by Kherkheulidze (1967).

Finally, experimental methods have also been used in research on *wind erosion, deflation,* and the *accumulation* of *aeolian deposits*. This subject was treated in detail by Chepil et al. (1945), Yakubov (1962), Zakharov (1965), Pasák (1962), and others. Comprehensive problems of water and wind erosion were tackled by Zvonkov (1962) using mathematical methods.

3.3.16 Complex methods

The methods mentioned in the previous sections are seldom used in isolation. Usually, and according to the objectives of the research, a combination of methods is chosen which makes for deeper investigation of the erosion phenomenon and for the prospect of more effective control measures. Despite this, most work is narrowly oriented and of local significance only. Sometimes erosion research does little more than complement pedological, land improvement, hydrological and other branches of research.

In selecting the method of research a statement of objectives and required precision is very important. A narrowly angled project involves the risk of producing distorting results, whereas broadly conceived research, on the other hand, makes heavy demands on labour, finance, and time. In all these considerations it is important to maintain a certain degree of comprehensiveness which allows a full assessment of the investigated phenomena.

For the purposes of undertaking *complex erosion research,* independent *research organizations* or *specialized agencies* have been set up, with highly sophisticated laboratory apparatus, *experimental research stations,* and a network of other installations. The most extensive erosion research is organized in the USA by the Soil Conservation Service and the U.S. Forest Service, in the USSR by specialized institutes of the Academy of Sciences, various ministries, colleges, etc., and in Australia by the Soil Conservation Authority. Intensive research is also carried out in China, Japan, New Zealand, in many European and African countries, and more recently in Central and South America.

In this chapter brief mention is made of the main principles and procedures used by the author and his colleagues since 1956 in their research on erosion in Czechoslovakia.

The first objective of this research was to identify, in the shortest possible time, the *occurrence of erosion* and to *assess its effect on the soil.* For this investigation, *experimental areas* were selected in the various geographic regions, where the relationships between the effects of erosion and erosion factors, and between the effects of erosion and natural conditions, respectively, were studied in detail. Special emphasis was laid on establishing the critical slope at which acute erosion starts on unprotected or poorly protected soils (average soil loss exceeding 0.5 m^3 ha^{-1} $year^{-1}$), and on identifying high levels of erosion that are not compatible with ploughing and raising agricultural crops. Inclination of the ground exceeding the critical slope was considered as being an important factor in deciding on the need for complex erosion control measures; a knowledge of erosion levels incompatible with cultivation served to identify land better suited to permanent stands of vegetation such as forest.

Special attention was paid to *erosion-degraded wasteland destined for afforestation.* These were the lands most severely affected by erosion in Czechoslovakia, and therefore they were well suited for the study of erosion phenomena. Simultaneously with the research on erosion and its effects on soil, methods of soil stabilization and afforestation were investigated on the wastelands.

In addition to permanent plots selected for particular attributes, erosion was also studied on temporary plots which were unexpectedly hit by extraordinarily heavy downpours or strong winds. During such downpours or gale conditions, it is possible to determine the maximum soil losses that may be expected; the erosion occurs in pronounced forms which are easily observed in the field. Finally, it is during such emergency conditions that the effectiveness of erosion control measures can best be tested.

The complex research in the various localities was based on *maps* of scale 1 : 20,000 and *aerial photographs,* by means of which plots were chosen and observed. Having established the boundaries of the plot, a contour map was prepared for each plot from aerial photographs, usually with a scale of 1 : 5,000 or 1 : 10,000; all visible types and forms of erosion, parcels of land (lots), the hydrographic network, communications, and other features were drawn in. This map served for guidance in the field, for the preparation of further maps showing ground inclination and slope aspects, and for the evaluation of erosion phenomena.

Erosion intensity was established in most cases (a) by direct volumetric measurement of the sizes of rills and gullies, (b) by levelling over the ground surface, and (c) by the vegetation method, with the aid of trees growing on erosion remnants. Topographical changes occurring on eroded land were monitored using maps prepared from ground-level photographs. Values obtained from volumetric measurements were checked with measurements of silt flow made by the bathometric integration method.

The *erodibility of the soil* was determined mainly by granulometric and structural analyses. Great importance was attached to the investigation of changes in soil properties brought about by erosion. This research proceeded along the lines of general pedological research with the difference that sample collection, soil analysis, the study of soil ecology, etc., were limited to the surface layers of the soil profile. Soil probes were distributed in the field so as to gain the best possible information on differences in soil catenae. The effects of erosion on granular composition and soil structure, on levels of organic matter and nutrients, on the water regime and the microclimate, and the damage caused by erosion to crops, were determined with special care. Of the various types of erosion control measures, soil-protecting afforestation was investigated in greatest detail.

The research also included the *photographic documentation* and collection of other *data required* in the study of erosion phenomena. Simultaneously with this research work, general data relating to the geology, soil, climate and economic utilization of the investigated territory were studied and evaluated.

3.4 Conclusion

To the methods already listed, others could be added, each of the latter being used only for one particular purpose. The significance and utility of a method varies, of course, according to the nature of the erosion phenomena which is under investigation. In most cases, it is an advantage to combine several methods, both in the field and in the laboratory, provided that this does not deviate from the principal objectives of the research. Erosion research usually entails studying the intensity of erosion *(quantitative research),* the effect of erosion on the soil *(qualitative research),* the damage caused by erosion (a synthesis of the foregoing investigations), the susceptibility of the soil (or territory) to erosion, the outcome of erosion control measures, and finally, the distribution of erosion phenomena and areas in need of erosion control measures. These objectives may be pursued directly or indirectly.

The *intensity of soil erosion* can be determined either directly by establishing the rate of soil loss and the form of the erosion phenomena (using levelling, volumetric, pedological, morphometric, photogrammetric, and vegetation techniques), or indirectly by analyzing the constituents and quality of removed soil (deluates, deflates, etc.), or in the case of sediments, the constituents and quality of deposits (using levelling, volumetric, deluometric, deflametric, pluviometric, or climatological and hydrological techniques). Of special importance are historical methods, which may be used for the direct determination of erosion intensity, provided that sufficient reliable data are available. The advantage of direct methods is that the precise location of an erosion phenomenon and its extent across the terrain can be established. Indirect methods, particularly stationary measurements, in which the movement of eroded material is recorded, make possible the observation of changes in erosion phenomena with time.

The *qualitative effects of erosion* are studied primarily by pedological and vegetation methods. Some aspects of qualitative changes may also be observed by monolithic, hydrological, deluometric, deflametric, or even photogrammetric methods. Although there are endless possibilities for methods of researching the effects of erosion on the soil, the most reliable methods remain those involving soil indicators. It should be noted that qualitative research is of importance mainly in situations in which erosion acts selectively, the comprehensive evaluation of the qualitative effects on the soil is requiring data on erosion intensity or total soil loss, because the smaller the thickness of soil, the smaller its ecological values. Therefore, in assessing damage caused by erosion, quantitative and qualitative research should be combined.

The *susceptibility of the soil* to erosion may be investigated either directly by comparing erosion intensities under different conditions, or indirectly by observing erosion under artificially created and controlled conditions. Consequently, all methods used for the determination of erosion intensity are also suitable for gauging the susceptibility of the soil to erosion. Pluviosimulation and monolithic methods are special methods for the determination of soil erodibility under controlled conditions. Valuable information on erodibility has also been obtained by pedological methods. Besides the determination of the relationships between the soil and erosion factors (water, wind, etc.), relationships between the soil and soil-forming, or other erosion conditions are also of importance. The evaluation of these relationships is therefore very important in any study of the susceptibility of the soil to erosion. Accordingly, one may speak of climatic, hydrological, or geomorphological susceptibility of the soil to erosion, etc. The most frequently used indicators of soil susceptibility to erosion are amounts and intensity of rainfall, the frequency and velocity of the wind, the steepness of the terrain, the runoff coefficient, or roughness coefficient, respectively, and the disaggregation tendency of the soil under the influence of water and wind, etc. As a rule, special methods are used for the determination of these indicators.

Effectiveness of erosion control measures by determining the quantitative and qualitative aspects of erosion phenomena and the susceptibility of the soil to erosion, the effectiveness of erosion control measures frequently becomes evident also. If, for example, erosion intensity is determined on unprotected soils and on soil protected by various crops, the protective effect of each crop can be assessed. However, in some cases the measures that are recommended for introduction into an already economically viable complex must first be tested. The effectiveness of control measures may be investigated with models either in the laboratory or directly in the field, using some of the standard methods of erosion research or modifications of these.

Exploring the *distribution of erosion phenomena* and identifying areas in need of *erosion control measures* usually constitute the final stage in an erosion research programme. In most cases practical applications based on information obtained also need to be supported by knowledge of the extent of erosion phenomena with regard to area. On larger areas, it is generally necessary to establish only the susceptibility of the soil to erosion or the degree of danger from erosion posed on the soil or territory, taking into account the current set of natural conditions and the expected economic utilization of the soil. The actual occurrence of erosion can be determined only by concrete measurements made on small areas, these measurements being valid only for a limited period since erosion activity is relatively variable. The most important indicators of erosion phenomena which express their distribution and which are easily recorded with respect to the surface distribution, include the degree of soil erodedness, the density of erosion gullies, and the susceptibility of the soil to erosion expressed in terms of edaphic, climatic, hydrological, geomorphological, vegetation, or other criteria.

The research of erosion distribution usually culminates in the drawing of *maps* which show the surface extent of erosion, and thus also its harmfulness. According to the indicator used, maps of the distribution of soil erosion may show erosion intensity (maps of isoerodents), grades of soil damage, the type, form, and age of erosion as well as the flow of silt, the susceptibility of the soil or terrain to erosion or the areas in which erosion control measures are in effect. Substantial assistance in cartographic work is afforded by aerial photographs and currently existing maps containing information on precipitation, temperature, hydrogeology, pedology, topography, vegetation, etc.

In addition to these objectives, erosion research may pursue other inquiries of narrower, or more local significance. The more important of these include inquiries into the effects of erosion on the water regime, on the yields of agricultural crops, on natural vegetation and the secondary effects of the latter on diseases and vermin, on human health, on damage to constructions, on the choking of rivers and silting of reservoirs, etc. This research is mostly concerned with the indirect consequences of erosion and borrows methods from other branches of science which have some bearing on the wider aspects of erosion.

In conclusion, it should be stressed that the methods of erosion research are not yet fully developed, one of the reasons for this being that research is carried out in institutions which have different areas of interest. A greater effort is needed to *integrate research* and *standardize criteria for evaluating erosion.* One advantage of greater standardization in erosion research, apart from the gains to be made in making wider adaptations of already existing methods, is the possibility of making observations and comparing results from larger areas, with the result that our understanding of erosion may increase more rapidly. Some progress may also be expected from the introduction of more up-to-date methods of research based on new or as yet incompletely established principles. Mathematical and statistical methods have a very important part to play in the improvement of techniques and should form a part of all methods.

A general chart of all methods discussed and the main objectives of soil erosion research is given in Table 23.

The methods described in the foregoing could be discussed in much greater detail, which would, of course, require a separate study. The author hopes that the survey given here is sufficient to show clearly the aims and state of development of the *methodology of soil erosion science.* The range and standard of information is

Table 23. General chart of methods and main aims of soil erosion research

Research methods	Research objective									
	I		II		III		IV		V	
	1	2	1	2	1	2	1	2	1	2
Nivelation, geodetic	x	x		x		x	x		x	
Volumetric	x			x		x	x		x	
Deluometric		x	x		x		x			x
Deflametric		x	x		x		x			x
Climatological		x							x	x
Pluviological			x		x		x			x
Monolithic			x		x			x		x
Pedological	x	x	x	x		x		x	x	x
Hydrological		x		x		x		x		x
Vegetation	x			x		x		x		x
Historic		x		x		x		x		x
Morphometric	x		x			x		x	x	x
Photogrammetric	x	x	x			x		x	x	x
Cartographic		x		x		x		x	x	x
Mathematical	x	x		x		x		x		x

I — erosion intensity, II — qualitative effect of erosion on soil, erodedness, III — susceptibility of soil to erosion, erodibility, IV — effectiveness of erosion control measures, V — distribution of erosion and erosion control measures, 1 — direct determination, 2 — indirect determination.

increasing very rapidly indeed and the scope of erosion research is becoming wider, as indicated by the numbers of research staff engaged in this work. Thus in the USA about 4,000 specialists are working on erosion research (Hudson 1971), and in the USSR more than 6,000 persons in 52 colleges of the Ministry of Agriculture of the USSR are engaged in research on water and wind erosion based on well established principles (Dryabezgov 1976). Methodological matters have been discussed at several conferences devoted to the subjects of research methods, conservation measures, economic evaluation of losses caused by erosion, and the economics of erosion control measures, etc. Thus for example, at the Second College Conference in Moscow in January 1976, more than 270 papers dealing mostly with the methodology of *erosion science* (eroziovedenie) were presented.

This means that since the instigation of the first pedoerosion research expedition in the USSR in 1939 (Sobolev 1939), the methodology of erosion research has seen very rapid development, and it is desirable that much greater attention be given to this aspect of the subject so as to provide a firm foundation for the theory of soil erosion.

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Chapter 4

EROSION FACTORS AND CONDITIONS GOVERNING SOIL EROSION AND EROSION PROCESSES

4.1 Introductory remarks

As we have seen in the foregoing chapters, soil erosion is an unusually complicated process which exercises considerable influence over the properties of the surface layers, as well as those of the deeper layers of the soil cover, and the underlying bedrock. Thus precise expression of the significance of the various factors and conditions is by no means simple, especially when attempting a universal generalization of the relationships between them and erosion processes.

By the *word factor,* the *active agent of erosion,* e.g. water or wind, is understood, and correspondingly different types of erosion may be distinguished, such as water or wind erosion. By the *word condition,* we are referring to an environmental component which influences the intensity, form, and other characteristics of the erosion process by modifying the action of the erosion factor; these conditions are either *natural* or *man-made.* Natural conditions are further divided into *living (biotic)* and *non-living (abiotic)* conditions.

As already mentioned, erosion caused by an erosion factor under natural abiotic conditions, i.e. without the protective effect of vegetation, animal or human interference, represents the maximum possible erosion on a particular site, and is called *potential erosion.* Erosion occurring under natural abiotic and biotic conditions is called natural erosion; if the latter is increased by man, it becomes accelerated erosion, or if it is slowed down by man, it is referred to as inhibited erosion. An existing state of erosion is called actual erosion, an admissible state of erosion is *tolerable erosion,* and an expected state of erosion is called *prognosticated erosion.*

In this chapter it is the main intention of the author to give a summary of available information on those *erosion factors* and *conditions* which govern the intensity, form, and other characteristics of erosion processes. The overall purpose of collecting and analyzing this information is to determine (a) the *degree of erosion danger,* of which the maximum is equivalent to potential erosion, (b) the *damage* caused by *erosion to the soil,* from small changes through *deterioration* to *degradation,* the final stage of damage being a total *destruction of the soil,* (c) the degree of *soil conservation* which can be achieved by a permanent cover of vegetation of suitably high density, and (d) the *improvement in the properties of eroded soils* and entire areas achieved by erosion control schemes.

In this way the analysis of erosion factors and the conditions governing erosion is aimed at making a *diagnosis of actual damage* and a *prognosis of possible erosion damage* to the *pedosphere.* This makes possible the more effective protection of the soil against erosion, and the rational utilization of land without endangering its substance, improving at the same time both the production potential of the soil and its value as a part of the environment. In fulfilling these aims also conditions in which decrease in pollution of the human environment (mainly water and air) by erosion products will be created.

Again, most attention is given to precipitation and wind erosion.

4.2 Precipitation erosion

The main erosion factor in *precipitation erosion* is *rainfall* moving with vertical and horizontal components. Rain affects soil both by the influence of raindrops and by the influence of surface runoff, and subsurface runoff, respectively. Another factor in precipitation erosion is *hail* which has a severe effect on the soil surface on account of its kinetic energy being several times greater than the energy of rain; thus the soil surface is completely destroyed and much material may then be washed away. *Hailstorms* give rise to the greatest danger. The importance of rain and its influence on the soil is often exaggerated; rain being considered as the most important erosion factor under all natural conditions.

In some regions erosion is caused largely by *snow water* which during thaw conditions (especially in continental regions where temperatures rise quite rapidly in the spring) often gives rise to greater erosion losses than those caused by rain. In cases of rapid erosion of the soil by water, the process is often enhanced by the total disaggregation of the soil by frost and the saturation of the surface layers with water which sets off a process of mass flow (cryosolifluction or also aquasolifluction). The saturation of the soil by water is assisted not only by freezing, but also by snow water which if it cannot enter the frozen ground saturates the upper unfrozen layers producing a highly liquid state which favours erosion.

Horizontal precipitation indirectly assists erosion in regions with a high occurrence of fog. The horizontal component of precipitation is known to account for as much as one third or a half of the total volume of precipitation in some European mountain districts, thus causing greater runoff and consequently a greater degree of erosion.

4.2.1 Raindrop erosion

Over large parts of our planet *rain* is the most important erosion factor. The erosive effect of rainfall depends mainly on its physical characteristics, such as the rate and extent of the rainfall, the velocity of raindrops and their direction of fall, electric potentials in the atmosphere, pattern fluctuations in the rain and its frequency of recurrence, the degree of coincidence with other factors, especially hail. It could be generally stated that the erosive effect depends, in the first place, on the *kinetic energy of the precipitation factor.* Some authors consider the kinetic energy of the rain to be the basic issue, regarding the discovery of a correlation between kinetic energy and the erosive effect of raindrops as the most important one.

Stallings (1957) writes that new understanding of *raindrop splashing* has meant the end of an era in the fight against water erosion, and sees the start of a second age in which, for the first time, there is hope of a successful solution to the problem. According to Stallings, Ellison was the first to appreciate that the falling raindrop represents all aspects of the erosive effect of rain. The protection afforded by vegetation is based on the fact that the falling raindrop is stripped of its kinetic energy.

Since it was established that the concentration of soil in the runoff water increases with the *energy of the raindrops* (Laws 1941), much attention has been directed at establishing the role of the kinetic energy of raindrops. In experiments on artificial rain, Borst and Woodburn (1942) found that a straw mulch placed at a height of 1 inch above the surface of bare soil diminished erosion by 95%. It was established that on small plots the removal of a large amount of soil from unprotected areas was determined by the impact of the drops and not by the surface runoff. Similar results were obtained by Hudson (1971), and other workers.

Ellison (1944) established by detailed study of the *erosive effect of raindrops* that the *disaggregation* caused by the raindrop is the initial stage of the erosion process. His investigations, which were based on direct measurement of soil disaggregation, showed that with a slope inclination of 1 : 10, 75% of the splash material is transported downhill and 25% moves uphill. Similar conclusions were arrived at by Ekern (1950), Mihara (1959), and other authors.

Soil disaggregation by the impact of raindrops is referred to by some authors as *impact erosion,* and the spattering of released particles as *splashing,* or *splash erosion.* This view of the erosive effect of raindrops is considered by many authors to be an independent aspect or process of water erosion which can occur without the process of runoff (Ellison 1944, Stallings 1957, Hudson 1971, and others).

Another result emanating from raindrop research is the finding that soil material becomes *selected* when raindrops fall on the soil. A part of this selective process is the mechanism by which fine soil particles are forced by the impacting drops deeper into the soil, that they can then be carried away by infiltration water into pores and cavities with the result that the soil surface becomes muddy and infiltration then decreases. The transport of fine particles into the soil pores is referred to by the term *puddle erosion.* Stallings (1957) describes this as follows:

"The sharp impact, as the drops beat on the naked earth during violent storms, shatters the clods and soil crumbs and breaks down the soil structure into a puddled condition. The beating and churning action of these drops compacts the soil's finely broken parts into an impervious layer of surface mud. This compacted surface layer is made denser and more impervious as it collects colloids and other particles from the turbid rain-water that filters down from the surface. Eventually, the porosity of this surface layer is materially reduced by the infiltration of muddy surface materials...".

With regard to this phenomenon it should be noted that some authors incorrectly include within the meaning of puddle erosion the sedimentation of finely broken material in the accumulation zone. Erosion, of course, means "eating away", and thus refers to processes of degradation and not to accumulation or aggradation of soil. The term puddle erosion could be accepted, within the limits allowed by the terminology, for referring to only one of the forms of *erosion degradation* caused by *rain falling on the soil,* but on no account should it be used to refer to soil enrichment with washed material and nutrients.

Another aspect of this grading process is the variation in the distance over which soil particles are carried when they are splashed into the air by raindrops; this distance depends on the weight and size of the particles. It was observed that the diameter of splashed particles is usually less than 2 mm, and it is the finer fractions of the fine earth that tend to be moved away. The smaller the particles, the farther they are carried and the finest material may even find its way into watercourses. The grading of soil particles by raindrops, according to some authors, is the most important factor governing not only the transport and wash of the finest soil fractions, but also the decline in the soil's fertility. Therefore Stallings (1957) and other authors refer to this type of precipitation erosion caused by raindrops as *fertility erosion.* It makes the soil surface coarse.

In the author's opinion there is a tendency to associate this phenomenon too closely with the erosive influence of raindrops. In the light of the author's research in every type of surface erosion, including subsurface water erosion and wind erosion, the grading of material can clearly be seen to occur together with a decline in soil fertility in every case. It would therefore be more correct to speak of selective erosion (e.g. selective drop erosion, or selective precipitation erosion). The term "fertility erosion" is more appropriate for expressing the influence of erosion in reducing the fertility of eroded soils in general, including the depletion of soil nutrients and fine material, changes in the physical, chemical, and biological properties of the soil, and the decrease in the depth of the soil profile caused by erosion process, etc.

The grading of soil material is not confined to the process of raindrop erosion alone, as shown by the fact that in sheet erosion, easily soluble and light, mostly organic material and nutrients are *washed out* of the soil. By far the largest part of this material is already loosened before rainfall or dissolved by rain and surface water containing weak acids. Thus in selective erosion, particles and substances released by *chemical* action, as well as *mechanical action* are washed away. In addition to this, soils which have formed on loess and loess clay may be very heavily eroded without the selective effect of raindrop action.

Further evidence that the selective effect of raindrops is not the sole factor involved in the selective action of precipitation water, is provided by the fact that some degree of selection also occurs in sheet erosion arising from snow melt water, and that the intensity and qualitative effects of the selection process depends mainly on the granular structure of the soil, the stability of the soil aggregates, the disaggregating effects of rain or frost, and the kinetic energy and transporting capacity of flowing water. The same is true of the wind.

As an example of the many observations made by the author during the course of his work (*Erózia pôdy,* 1970), the change in the granular structure near the surface of the soil brought about by spring snow melt waters on the experimental plot in Závadka (Low Tatras, CSSR) may be mentioned. The slope inclination was 9°12′, and the soil loss was 63.65 m^3 ha^{-1}. Further data are given in Table 24.

Table 24. Change of soil granulation on the research plot in Závadka during the spring snow thaw of 1958

Depth of sample [cm]	Size of soil particles [mm]					
	<0.01	0.01—0.05	0.05—0.1	0.1—2.0	2.0—10	>10
	Percentage present					
0—2	14.66	9.32	5.18	28.34	18.54	23.16
10—15	35.01	14.38	6.77	28.41	11.32	4.11

By plotting these data semilogarithmically and subtracting the value d_{40} (Fig. 82), it was established that the diameter of grain for equal totals of weight per cents increased with the erosion of soil by snow water approximately 20 times, according to the formula

$$\frac{d_{40}^{(2)}}{d_{40}^{(1)}} \doteq \frac{0.4 \text{ mm}}{0.02 \text{ mm}} \doteq 20,$$

where $d_{40}^{(2)}$ is the diameter of soil grains affected by selective erosion, and $d_{40}^{(1)}$ the diameter of soil grains not affected by erosion.

The data indicate that the soil surface was *depleted of the finest material* and *reduced to a skeleton.* Heavy rainfall together with a hailstorm produced a deterio-

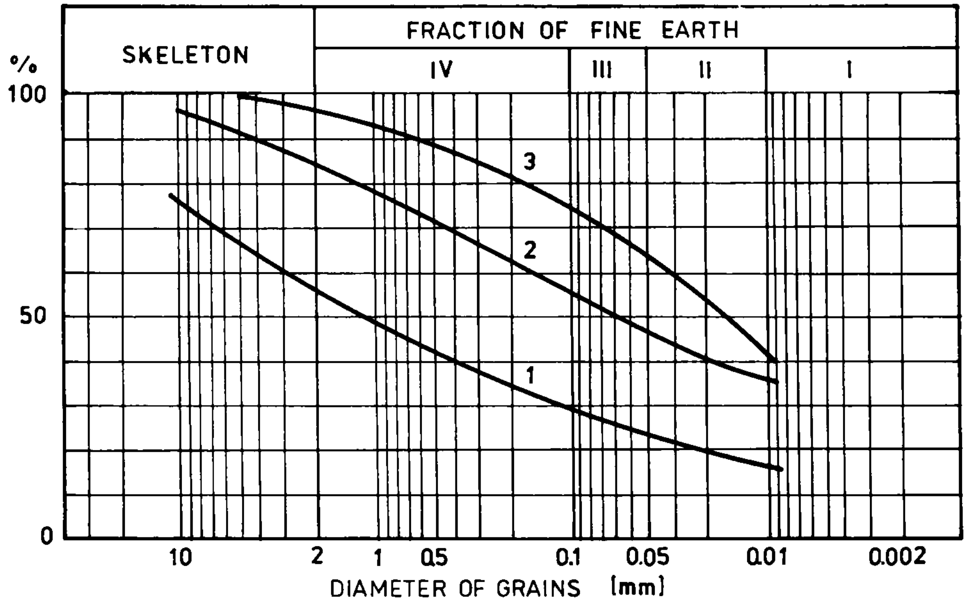


Fig. 82. Grain size curves of the soil: eroded (1), control (2), and deposits (3). Locality Závadka (Czechoslovakia).

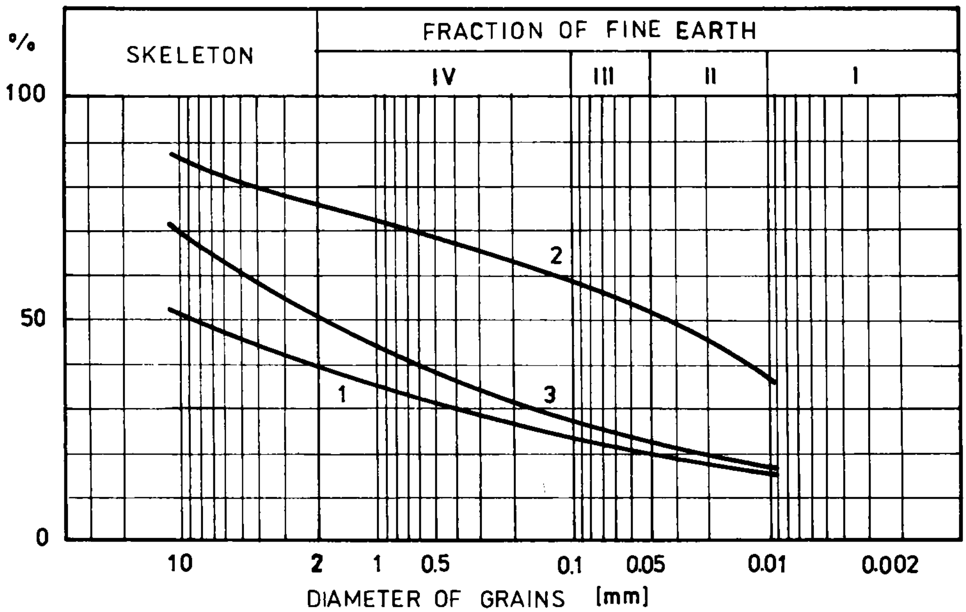


Fig. 83. Grain size curves of the soil: eroded (1), control (2), and deposits (3). Locality Lučatín (Czechoslovakia).

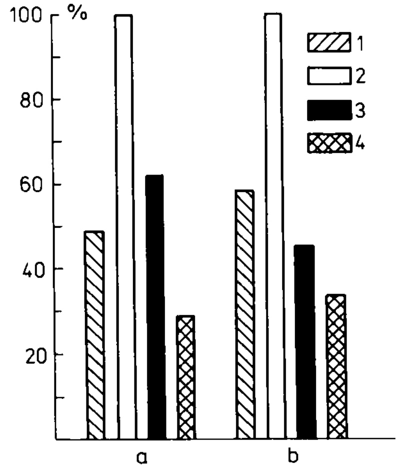


Fig. 84. Relative content: total of N, P_2O_5, K_2O in the localities Lučatín (a) and Hiadeľ (b) (Czechoslovakia). 1 – surface layer 0–3 cm, 2 – tilled soil 3–15 cm, 3 – subsoil under 30 cm, 4 – deposits. Surface layer was heavily damaged by rainstorm. For the purpose of comparison – tilled soil presents the situation before erosion.

ration of granulation expressed by d_{40} up to 200 times (Fig. 83), lighter downpours only about 10 times. The erosion also brought about a large decrease in the humus content and available nutrients (Fig. 84).

The author's measurements, both in the case of downpours and snow melt water runoff, revealed first a decrease, and later an increase in the *turbidity of water down the slope,* this being in proportion to the increase in the amount of eroded material being carried. Sometimes the turbidity exceeded the proportional relationship with the amount of eroded material, and this was explained by the fact that the amount of soil carried by surface water depends mainly on the flow and not solely on the erosive action of raindrops, although raindrops do have a highly disaggregating effect. As will be explained later, a lamina of water is created over the soil during rainfall and the soil is thus protected against the impacts of raindrops, so that the amount of soil loosened and displaced by raindrops decreases down the slope, particularly in places where the flow begins to gather into channels. In spite of this, the total losses due to erosion generally tend to increase with distance down the slope.

In considering the erosive effect of raindrops it is important to remember that the disaggregation and splashing caused by raindrops is only the first stage in the erosion process, this being followed by the washing away of the loosened particles, the dissolving of easily soluble substances, and further erosion caused by flowing water. Ellison (1944) demonstrated that erosion can occur without the action of flowing water, but he points out that maximum splashing occurs shortly after the soil surface has been moistened; then splashing gradually diminishes with the duration of the rain. Kuron and Steinmetz (1958) proved that soil losses due to erosion increase, despite reduced splashing of the soil, this being explained by a growing turbulence in the flow of water. Makkaveev (1955) observed that the turbulence of released soil particles reached a maximum at a water depth of

10–12 mm, at which depth the splashing effect of raindrops is already strongly inhibited by the water layer (Mirtskhulava 1970).

Thus the erosive effect of raindrops on the soil decreases with distance down the slope, whereas the erosive effect of flowing water increases. The more permeable the soil, the greater is the relative erosive effect of raindrops, while on the other hand, any factor which contributes to the increase of surface runoff, and therefore also to the "protective" effect of the surface layer, tends to diminish the impact of raindrops. It follows, that where water gathers in depressions (rills and furrows, etc.) the influence of raindrops will be strong in the elevated areas between channels. Under such conditions the erosive effect of flowing water is also greater, and consequently the total erosive effect of both raindrop action and flowing rain-water on exposed soil is considerable.

Perhaps the best evidence of the combined erosive effects of raindrops and flowing water is given by eroded soils which have suffered from the long-term aggregate action of both of these stages of the erosion process. It has been shown that elevated areas are mostly eroded by splash erosion, whereas wash erosion is the predominant influence on those parts of the slope where the relief is uneven and the kinetic energy of surface water high, regardless of whether the latter is attributable to an increase in steepness or length of the slope, coarseness of the surface, varying permeability of the soil on differently managed fields, or to some other factor. These observations have been confirmed by measurements of the intensity of erosion processes following immediately after downpours and the thawing of lying snow (Zachar 1970).

From among many known examples of the correlation between soil erosion and the inclination and length of the slope, the results of research on the influence of erosion on soils in the Perm region of the USSR may be mentioned. These soils have poor permeability and are mostly heavy; erosion is caused by downpours and snow melt water. Annual losses on arable land amount to 20–60 t ha^{-1}. Pronounced erosion on the more permeable ground occurs at an inclination of 8° and a precipitation intensity of 0.3 mm min^{-1}; on heavy impermeable soil (which covers about 65% of the land surface) pronounced erosion occurs at a rainfall rate of 0.05 mm min^{-1}, that is, at precipitation levels with very small kinetic energy and thus little splash effect. During thaw conditions when the rate of runoff was still smaller and was measured on a straight slope of 4°, a soil loss of 5.16 t ha^{-1} was established at a distance of 50 m down the slope and a loss of 53.6 t ha^{-1} was recorded at 200 m. Over a period of 200 to 300 years of agricultural utilization the soils of this region have been heavily eroded; detailed investigations (Skryabina 1972) have established a correlation between slope length and the degree of soil erosion (Table 25). The degree of soil erosion was assessed by the method of Presnyakova (1956).

The data show that as the distance down the slope increases, the soil damage caused by erosion increases in proportion with the increasing erosive effect of

Table 25. Relationship between soil wash and length and inclination of slope

Soils	Slope inclination									
	1.5—3°			3—6°			6—12°			>12°
	Runoff distance [m]									
	<100	100—500	>500	<100	100—400	>400	<100	100—400	>400	
Sod podzol, heavy mechanical composition	I	II	III	II	III	IV	III	IV	V	IV—V
Sod brown soil, medium mechanical composition	I	II	—	I—II	II—III	—	II	II	—	III–IV–V

I — soil unwashed, II — soil slightly washed, III — soil moderately washed, IV — soil heavily washed, V — soil very heavily washed.

flowing water. Similar conclusions were reached by Flohr (1962) in the German Federal Republic in an investigation of soil erosion caused by spring rains of low intensity. Observations were made when the daily total rainfall was between 5 and 20, and at most 30 mm, the effects of erosion appearing mostly in the form of wash and rill erosion.

For these reasons the author suspects that some research workers tend to overestimate the erosive influence of raindrops, to the extent that *impact* or *splash erosion* become identified with *sheet erosion* and the actual importance of the latter process is underestimated, or is even regarded as superfluous.

In classifying *precipitation erosion,* the position of *rain erosion* as a subtype is justified, encompassing *impact erosion* as the first stage of *drop erosion,* with the concomitant stirring-up of sediment *(puddle erosion);* the second stage is represented by *splashing* followed by selective dispersal erosion which has, of course, broader significance. These two stages of rain and *hail erosion,* respectively, are followed by erosion caused by rain-water runoff which may therefore be termed *runoff erosion, surface-flow erosion, rainwash erosion,* etc.

Some authors also use the term *slope erosion* to refer collectively to all the types and forms of erosion caused by precipitation water. By using this term, attention is drawn to the nature of the site rather than the erosion factor, and therefore the expression is not quite correct. In any case, soil erosion on slopes involves erosive action by both raindrops and flowing water; the proportions and significance of the effects of each of these components under various conditions will be discussed later.

It can be generally stated that in climates which are more continental in character, and under conditions of reduced disaggregation by frost, the erosive effect of raindrops becomes more important. The effect of flowing water derived from rain and thawing snow also varies over different parts of the terrain and depends on the configuration of the relief, the permeability of the soil, etc. Whereas drop erosion has a greater influence on the upper parts of slopes and fields, flowing water is more important in the lower parts. As to soil permeability,

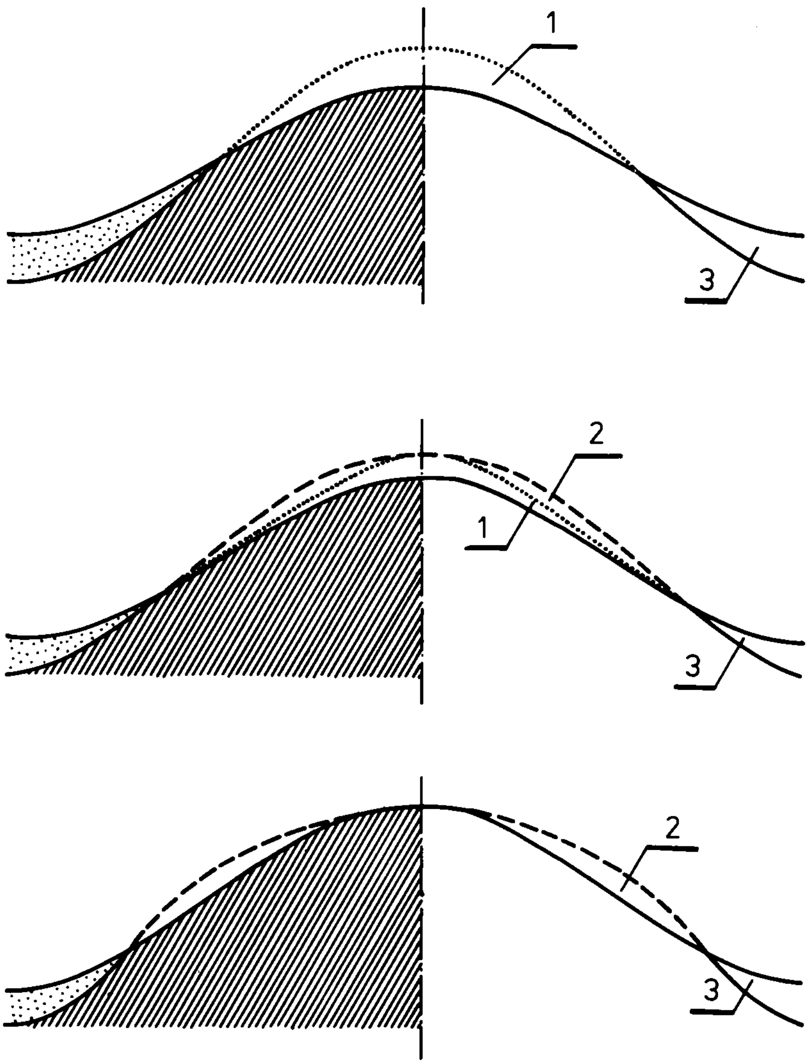


Fig. 85. Schematic representation of the erosive effect of precipitation. a – according to the theory of raindrop action (1), b – according to the theory of raindrop action (1) and surface water action (2), c – according to the theory of surface water action (2). Cases a and c are exceptional. Points of inflection change according to the acceleration or inhibition of erosion and deposition processes. Deposits are designated by the number (3).

the more permeable soils are affected to a greater degree by drops, the less permeable soils are affected more by flowing water, and on very permeable, cracked soil, subsurface runoff within the soil being the decisive factor. These processes may also be substantially influenced by man's activities. Therefore in situations of accelerated erosion, not only the intensity, but also the form of erosion is changed. The schematic representation of the erosive effects of precipitation in Fig. 85 shows the most common and frequently occurring situations, and the theoretical possibility of some extreme situations.

From both a theoretical and a practical point of view it is desirable to establish the proportions of the total erosion attributable to raindrops or water flow under different conditions. Unfortunately, with respect to these two components, sufficiently precise parameters for the construction of empirical formulae, such as those used to express total precipitation erosion, have not yet been established. For the purpose of approximation it may be worthwhile considering the kinetic energy of raindrops and the kinetic energy of rain-water running down an inclined plane.

The *kinetic energy of raindrops* is determined mainly by their size and form. In general, the greater the intensity of the rain, the larger are the diameters of the drops, the velocity of the drops, and consequently also their kinetic energy.

According to Mirtskhulava (1970), the *size of raindrops* depends on the *intensity of precipitation,* as demonstrated by Obolenskiĭ's and Nikandrov's data (Table 26).

Table 26. Relationship between size of raindrops and rain intensity

Characteristics of rain	Intensity of rain [mm min^{-1}]	Size of drops typical [mm]	Velocity of drop fall [m s^{-1}]	Distance between drops [mm]	Water contents [g m^{-3}]
Fog	—	0.01	0.003	4.3	6×10^{-3}
Mist	0.0003	0.1	0.25	21	57×10^{-3}
Drizzle	0.0042	0.2	0.75	36	93×10^{-3}
Shower	0.016	0.45	2.0	70	0.14
Rain	0.066	1.0	4.0	123	0.28
Heavy rain	0.25	1.5	5.0	130	0.83
Very heavy rain	0.66	2.1	6.0	138	1.8
Downpour	1.61	3.0	7.0	137	5.4

The *diameter of raindrops* generally fluctuates around 1 mm, the majority of drops having diameters between 0.2 and 0.6 mm, and the largest drops being as much as 6 mm in diameter. However, in violent cloudbursts the drop diameter (expressed by d_{50}) fluctuates between 2.0 and 2.5 mm; this refers mainly to *tropical rains* with a *high kinetic energy* (Hudson 1971). According to Best (1950), the median volume diameter $d_{50} = aI^{b}$, where i is the intensity of precipitation, and a and b are constants.

Table 27. Downpour characteristics

Average intensity [mm min^{-1}]	Duration [min]	Total amount of precipitation [mm]	Average intensity [mm min^{-1}]	Duration [min]	Total amount of precipitation [mm]
0.50	5	2.5	0.23	45	10.25
0.38	10	3.8	0.22	50	11.00
0.33	15	5.0	0.20	60	12.00
0.30	20	6.0	0.15	120	18.00
0.28	25	7.0	0.11	240	27.00
0.27	30	8.0	0.06	720	45.00
0.24	40	9.6	0.04	1,440	66.00

A *downpour* (according to Bern) is rainfall corresponding to the criteria given in Table 27.

At higher intensities of precipitation, the duration of rainfall tends to be shorter and the area affected smaller. Alekseev (1941) arrived at the following general formula for the calculation of the mean intensity of a downpour

$$i = \frac{A + B \lg N}{(1 + t)^{2/3}} \quad [\text{mm min}^{-1}] ,$$

where i is the greatest intensity of rain at time t in a period of N years, and A and B are geographical constants which are determined by climatic conditions and calculated from pluviometric data.

For Slovakia (CSSR), Dub (1955) derived the following formula for the mean intensity of a downpour

$$i = \frac{3{,}200}{(t + b)^{0.675}\,(150_p)^n} \quad [\text{l s}^{-1}\ \text{ha}^{-1}] ,$$

where p is the mean periodicity of the rainfall, n the exponent (ranging from 0.25 to 0.33) which is a function of the intensity of rain over the given locality, t the mean duration of the rain (minutes), and b the number approximating unity.

The size of raindrops increases with the velocity, and the erosive effect depends mainly on the *terminal velocity*. Of the many data available, those of Gunn and Kinzer (1949) are shown in Table 28; the measurements were made with electrooptic equipment.

The velocity of falling raindrops can be computed using any of several empirical equations. The well-known equation of Schmidt

$$v_k = 10^6\,\frac{0.787}{r^2} + \frac{.503}{\sqrt{r}} \quad [\text{cm s}^{-1}]$$

Table 28. Terminal velocity of water drops in stagnant air (pressure 101.3 kPa; temperature 20°C)

Diameter of drops [cm]	Terminal velocity [cm s^{-1}]	Weight [mg]	Diameter of drops [cm]	Terminal velocity [cm s^{-1}]	Weight [mg]
0.01	27	0.524	0,26	757	9,200
0.02	72	4.19	0.28	782	11,490
0.03	117	14.14	0.30	806	14,140
0.04	162	33.5	0.32	826	17,160
0.05	206	65.5	0.34	844	20,600
0.06	247	113.1	0.36	860	24,400
0.07	287	179.6	0.38	872	28,700
0.08	327	268	0.40	883	33,500
0.09	367	382	0.42	892	38,800
0.10	403	524	0.44	898	44,600
0.12	464	905	0.46	903	51,000
0.14	517	1,437	0.48	907	57,000
0.16	565	2,140	0.50	909	65,500
0.18	609	3,050	0.52	912	73,600
0.20	649	4,190	0.54	914	82,400
0.22	690	5,580	0.56	916	92,000
0.24	727	7,240	0.58	917	102,200

was simplified by Slastikhin (1964) to

$$v_k = 13\sqrt{d} \quad [\mathrm{m\ s^{-1}}] ,$$

where $r = d$ is the drop diameter [cm].

On the basis of the velocity and size of raindrops the *kinetic energy* of both *raindrops* and *rain,* respectively, can be computed, and from the kinetic energy values the amounts of soil particles dislodged by raindrops can be arrived at. According to Ellison (1952), splash erosion can be derived from the formula

$$s - v^{4.33} d^{1.07} i^{0.65},$$

where s is the amount of soil splashed in 30 minutes [g], v the raindrop velocity [feet s^{-1}], d the drop diameter [mm], and i the rainfall rate [inches h^{-1}].

Mirtskhulava (1970) obtained the kinetic energy of falling raindrops from the well-known formula

$$E_k = \frac{m_k v_c^2}{2} ,$$

where E_k is the kinetic energy, m_k the mass of the raindrops $-$ $\varrho = \pi d_m^3/G$, v_c the terminal velocity, and derived the highest velocity of falling drops permissible in artificial irrigation (i.e. the critical velocity of falling drops, v_{cd}, below which there is no appreciable soil damage) from the formula

$$v_{cd} = \frac{0.7MC}{\varrho},$$

where M is the coefficient of friction (approx. 0.3), C cohesion of soil when saturated with water (approx. 0.05 kg cm^{-1}), ϱ the density of water (1.02×10^6 kg s^2 cm^{-4}), and v_{cd} the critical drop velocity (in this case 1.0 m s^{-1}).

Finally, the amount of soil [t ha^{-1}] eroded by raindrops falling vertically is computed using the formula

$$q_D = \frac{0.13\gamma\, i\, v_c^2}{2g}\, t \left(\frac{d_k v_c}{d_{dk}\, v_{cd}} - 1 \right) 4 \sin \alpha' \qquad [\text{t ha}^{-1}],$$

where q_D is the amount of soil [t ha^{-1}] eroded by splashing, γ the density of soil saturated with water [t m^{-3}], i the intensity of precipitation [mm min^{-1}], v_c the terminal velocity of raindrops [m s^{-1}], t the duration of rainfall [min], d_k the mean diameter of the raindrops [mm], d_{dk} the diameter of raindrops [mm] of the critical size below which splash erosion does not occur, and α' the angle made by the line of vertical fall of the raindrops and the ground.

The amount of soil splashed by raindrops, as derived according to these equations, approximates the measurements obtained experimentally by Ellison (1952), as shown in Table 29.

Table 29. Amount of soil splashed by raindrops according to Ellison (1952)[1] and Mirtskhulava (1970)[2]

Drop diameter d_k = 3.5				Drop diameter d_k = 5.1 mm			
Rain intensity [mm min^{-1}]	Velocity [m s^{-1}]	Amout of soil[1] [g]	Amout of soil[2] [g]	Rain intensity [mm min^{-1}]	Velocity [m s^{-1}]	Amount of soil[1] [g]	Amount of soil[2] [g]
2.0	3.66	15.3	43	2.0	3.66	35.7	64
2.8	3.66	20.5	61	2.8	3.66	61.7	90
6.3	3.66	47.8	137	6.3	3.66	157	203
2.0	4.42	67.1	77	2.0	4.42	203	125
2.8	4.42	96.3	107	2.8	4.42	233	176
6.3	4.42	232	241	6.3	4.42	329	359
2.0	5.49	223	147	2.0	5.49	446	263
2.8	5.49	245	206	2.8	5.49	543	368
6.3	5.49	492	464	6.3	5.49	786	828

Example: Experimental plot 1.5×1.8 m = 2.7 m^2; $\gamma = 1.5$ t m^{-3}; $i = 2.8$ mm min^{-1} = 0.0028 m/60 s = 0.00004667 m s^{-1}; $v_c = 5.49$ m s^{-1};

$$v_{cd} = \frac{0.7MC}{\varrho} = \sqrt{\frac{0.7 \times 0.3 \times 0.05}{1.2 \times 10^{-6}}} = 1.0 \text{ m s}^{-1}$$

$t = 30$ min = 1,800 s; $d_k = 3.5$ mm = 0.0035 m; $d_{dk} = 0.2$ mm = 0.0002 m; sin α' = 0.1; $g = 9.81$; $q_D = 0.958$ t ha^{-1}, or 258 g per plot of area 2.7 m^2.

Performing the calculation, it is found that the amount of soil splashed by raindrops fluctuates between 0.6 and 3.07 t ha^{-1}.

Some authors quote much higher values. Stallings (1957) states that on bare soil more than 100 t acre^{-1}, i.e. 247 t ha^{-1}, may be splashed during a violent cloudburst. According to Osborn (1954), raindrops splashed 240 tons of soil per ha into the air during a downpour in which 50 mm of rain fell at a rate of 20 mm h^{-1}.

In assessing soil losses from erosion, account must be taken not only of the *amount of splashed soil,* but also of the *distance* over which soil is dispersed by raindrops during a downpour. On a horizontal surface and with vertical rainfall, there is no net transport because the drops are splashed equally in all direction. If the rain is falling obliquely or falling on a slope, there is net transport down the slope, or in the direction of the wind.

For an approximate analysis of this process it may be sufficient to examine the data of Ekern (1953), who established that the proportions of soil particles dispersed in opposite directions during rain splash can be estimated by adding or subtracting the percentage of the slope inclination to or from 50% of the total amount of soil being moved. For example, in the case of 240 t ha^{-1} total soil movement, 144 tons move down the slope for a 10% inclination, and 168 tons move down the slope for a 20% inclination. The net movement in the first instance amounts to 48 tons, and in the second, to 96 tons. According to Ellison, a slope of 10% would result in 75% of the splashed soil being transported down the slope and 25% up the slope. In the above-mentioned case this would mean that 60 tons were transferred up the slope and that 180 tons were moved down the slope; the net movement due to erosion would then be 120 t ha^{-1}.

Unfortunately, there are no reliable data concerning the distance travelled by the soil. According to data published by Ellison (1944), Mihara (1959), and other authors, soil particles are transferred over distances of 0.5 to 1.5 m, as calculated from the approximate equation

$$l = \frac{4(kv)^2}{g} \sin\alpha ,$$

where l is the distance of transfer [m], k the coefficient of resistance, α the angle of the slope, or the angle of deviation of the direction of drop fall from the vertical, and g the gravitational acceleration.

According to these considerations, soil splashed by rainfall can be transferred over distances ranging from negligible values up to a few metres. In the example considered above, for a soil movement about 120 t ha^{-1} and a transport distance of 1 m, the absolute soil loss in the upper part of the field would amount to 1.2 t ha^{-1}, and in the lower part there would be an accumulation of 1.2 t ha^{-1}.

Even if the splash distance were 10 m and the absolute soil loss was consequently ten times larger (12 t ha^{-1}), the losses would not greatly exceed the limit of tolerable *erosion, resembling somewhat the soil* shift due to ploughing, or other agricultural operations. It has been shown that during cultivation the movement of the soil is always greater downhill than uphill (Lammel 1958, 1960), so that the soil profile tends to be reduced on elevated sites, steep terrain, bends, upper borders of fields, etc. The steeper the slope, the deeper the soil cultivation, and the faster the cultivation operation – the larger is the net soil transport. If the shift of the soil due to turning the furrows in one 30 cm deep ploughing operation is considered alone, the total shift of soil amounts to 30,000 m^3 ha^{-1}, and the depletion in the upper part of the ploughed area (represented by one furrow with approximate dimensions 30×40 cm) is 12 m^3, or 18 t ha^{-1}. Even if a part of the soil is recovered as a result of the turning of the furrow up the slope, the total losses would show a relationship with slope inclination similar to that between total losses in splash erosion and slope inclination, reaching about 12 t ha^{-1} for tractor ploughing on an inclination of 10%.

Losses caused by *wash erosion* involve the entire eroded surface, although a transitional sedimentation occurs on the surface. When a statement of losses from erosion is given for a certain area, the value given refers to the amount of soil that has been fully removed from the area. A part of this soil may be deposited in the catchment area, but a large part flows away into the river system and is entirely lost from the land. As we have seen, a soil movement of 120 t ha^{-1} caused by raindrop splashing represents an absolute loss of 1.2 t ha^{-1}, whereas in wash erosion this loss over a transport distance of 100 m amounts to 12,000 t ha^{-1}, i.e. 100 times more.

In analyzing the process of splash erosion, the relative importance of this phenomenon in terms of the *difference* between the *kinetic energy of raindrops* and that of *flowing water* becomes clear. Hudson (1971) calculated that the kinetic energy of precipitation R, taking the velocity of raindrops as 8 m s^{-1}, is $1/2 \times$ mass $\times$ $(\text{velocity})^2 = 1/2 \times R \times 8^2 = 32\ R$, whereas the kinetic energy of runoff originating from 25% of the precipitation (assuming 75% infiltration) and moving with a velocity of 1 m s^{-1} $= 1/2 \times R/4 \times l^2 = R/8$. This means that the kinetic energy of rain in this instance is 256 times greater than the kinetic energy of the runoff. If all the precipitation appears as runoff, the difference is only 64-fold. In other instances, much larger differences of up to 100,000-fold have been quoted (Stallings 1957).

When the rain falls vertically on a horizontal surface, the ratio of the kinetic energy of rain to the kinetic energy of flowing water approaches infinity since the water cannot flow away; however, in this situation splash erosion will not occur. As slope inclination increases, the differences become smaller, and would theoretically disappear on a perpendicular surface.

This mechanical comparison is only partially correct, for two reasons. First, the main effect of precipitation as an erosive agent is the disaggregation of the soil and its transport over relatively short distances, both effects only becoming important above a certain drop velocity and drop size. In surface runoff, the transport of disaggregated and dispersed particles begins at a very low velocity, and the carrying capacity of the flow is an important factor. As the water velocity increases, the quantity of material transported by the flow increase according to the well-known relationship $Q = Av^6$ (Q is the mass volume or weight of material, A the cofficient of proportionality, and v the velocity of flow). Because the amount of water increases with distance down the slope, accumulating first in small, then increasingly larger rills, the water accelerates rapidly over short distances, and its kinetic energy and carrying capacity are thus markedly increased.

In making these considerations, the author does not wish to underemphasize the enormous destructive effect of raindrops on the soil, all the more so in regions with very intense and erosive precipitation. A vivid illustration of the erosive power of precipitation in these regions is given by Hudson (1971), according to whom erosive precipitation with an intensity exceeding 25 mm h^{-1} occurs in the following amounts:

Temperate climate -5% of the total rainfall erosive; for 750 mm annual rainfall, there are 37.5 mm of erosive rain.

Tropical climate $-$ 40% of the total rainfall erosive; for 1,500 mm annual rainfall, there are 600 mm of erosive rain.

The annual erosivity of precipitation, according to the calculations of Hudson, is $37.5 \times 24 = 900$ J m^{-2} in temperate regions, and $600 \times 24 = 14{,}400$ J m^{-2} in tropical regions. The extremes of precipitation erosivity on our planet are, of course, much wider, and therefore the erosive effect of raindrops is also expected to be very different in different regions.

4.2.2 Hail erosion

A still greater effect than that of raindrops is displayed by hail, which may attain large dimensions (up to the size of a hen's egg) and totally destroy the soil together with its cover of vegetation. This type of impact erosion results from the dissipation of kinetic energy several times greater than that of rainfall, and if *hailstorms* are accompanied by heavy rain, they cause catastrophic erosion. In these cases the considerable erosive effect is caused not only by the high kinetic energy, but also by the oblique angle of fall of the hail and raindrops. Whereas the hail beats down on the soil and clogs the pores, the rain-water rapidly washes the disaggregated material away. In such instance, one may speak of downpour-hail erosion or storm-hail erosion, this being the most destructive erosion phenomenon known.

During the author's 20 years of erosion research experience, he has recorded topsoil losses caused jointly by very violent hailstorms and a heavy downpour. One such occasion was 23rd May 1958 on the experimental plot of Lučatín and Hiadeľ (Low Tatras, CSSR). The downpour affected an area of about 18 km^2. Locally, the hail tore off nearly all the leaves from the trees, particularly the fruit-trees, damaged the roofs of houses and destroyed all types of agricultural crops. Soil losses of up to 1,000 $m^3\ ha^{-1}$ were recorded, the remainder of the soil being a skeleton which was relatively resistant to further erosion (Fig. 75). The effect of hail alone on the soil has not been investigated so far (see Sec. 4.2.5.2).

4.2.3 Rainwash erosion

In the previous section the author attempted to present raindrop erosion of the soil as a process which depends mainly on the impact and splash effects. The conclusion that this phase of precipitation erosion may, under certain conditions, have an essential and strong influence on the erosive degradation of the soil, even without the additional effects of surface runoff was reached. This is concordant with the point of view that the phenomenon of raindrop erosion is an independent process. But this view is not theoretically correct, except in situations in which either there is no surface runoff, or if precipitation does fall on the soil, it causes no erosion at all, either above the surface or beneath it. Such circumstances, of course, are hypothetical, since any movement of water brings about some erosive effect on the soil. Therefore it seems that more attention needs to be given in the future to the processes of rainwash, runoff, and surface-flow erosion, respectively.

Regarding the relationship between these processes, the question arises about whether the surface flow of rain-water acts in laminar form, or whether *splash erosion* is immediately followed by *rill erosion* without any intermediary sheet effect of the surface water. The answer is, and arguments can be put forward in support of this, that *sheet erosion* caused by surface water certainly occurs and takes several forms, as discussed in Chapter 2.

Fournier (1956) has established that on permeable soil, in places where the effect of washing is negligible, there is a loosening, dispersion, and grading of grains on microareas, owing to the influence of raindrops beating on the soil under downpour conditions; coarse-grained sand remains in situ, and finely grained colloidal components are splashed by raindrops over short distances, so that the soil takes on the appearance of a patchwork of plots measuring a few square centimetres. Sandy and clay spots alternate with these areas.

One form of sheet erosion caused by rain-water is that, in which there is a *washing of particles* which have become partially loosened, or which are clustered in aggregates of such a size that these can be carried away by water. The amounts and proportions of particles washed away are very different from the amount moved in the ordinary way, and therefore this washing process is also different in intensity. The largest movements of soil particles by washing occur during the thawing of frozen soil in heavy rain, the quantity increasing with the kinetic energy of the rain. A large proportion of the soil particles in aggregates is released by water action in any case, even without the disaggregating influence of rain. Consequently, the proportion of particles loosened by *frost, high temperature, raindrops, flowing water,* and *mechanically during soil cultivation,* etc., may vary greatly. During winter and spring cryogenic processes are predominantly responsible for the release of particles, but in periods of climatic and edaphic drought, hydrothermal processes and wind action are the predominant factors. Under conditions of heavy rainfall the kinetic energy of raindrops is of chief importance and in long periods of continuous rain, hydroedaphic and hygroedaphic changes predominate.

Generally, it may be stated, that the higher the intensity of the rainfall, the greater is the quantity of soil made susceptible to the process of washing on account of raindrop action. However, in the case of long continuous rainfall of low intensity most of the disaggregation of the soil takes place in the underwater environment. It has been established that during rainfall of the latter type, although the turbidity of the surface runoff is less, the total washing of debris from the catchment area under typical central European conditions is higher than that which occurs during downpours falling on only a small part of the catchment area. Also, the quality of eroded material is different, because during continuous rain, or where the precipitation is of low intensity, it is mainly the fine particles that are washed away, so that there is a stronger selective effect on the soil. The greater the intensity of the precipitation and subsequent surface runoff, the larger are the particles and aggregates that are gradually carried away; consequently the selective effect of erosion in the eroded part of the field is less, and during catastrophic downpours the entire arable soil, including stones is washed away.

The selected fraction of finer material is not negligible, as indicated by analyses of bed load from several rivers. Thus, for example, a total of 14,198 tons of bed load was carried away from the catchment area of the Sekčov river (eastern Slovakia, CSSR) between March and September 1956. 7,983 tons of this material were accounted for by sediments collected in the decantation vessel within 24 hours. The balance of 6,216 tons sedimented out of the river water in more than 24 hours, and therefore could not be properly established by the decantation method. The same ratio was found for the Danube in Bratislava section (CSSR);

with an average water flow of 2,050 $m^3 s^{-1}$ and a bed load flow of 235 kg s^{-1}, 7,983 thousand tons of material exceeding 0.002 mm in diameter were carried away, the total transport reaching 14,198 thousand tons. The ballance of 6,215 thousand tons consisted of very fine particles not recorded by the international decantation method. Although the ratio is by chance nearly the same in the Sekčov and Danube catchment areas and would probably be different for other rivers, nevertheless, the data show convincingly that the proportions of very fine particles in the bed load are very high, indicating the considerable selective effect of water coming from the catchment area where the soil is depleted of its finest particles. Similar errors in determining quantities of eroded soil are also made in experiments on runoff plots (as can be seen in Fig. 73).

Another significant factor in wash erosion is *chemical erosion,* in which mainly surface rain-water, and to some extent snow melt water also, washes away matter that has originated from *fertilizers* and various *biocides* (herbicides, fungicides, insecticides, pesticides, etc.). Such substances are now being applied in ever increasing doses with the result that they reappear in greater quantities in the hydrosphere polluting and contaminating the water environment. A large proportion of these substances is applied outside the growing period, and is therefore easily washed away by snow melt water or water derived from low intensity precipitation of long duration. It is estimated that in some regions up to 40% of this matter is carried into the rivers. This is also true of *industrial fumes* which increasingly pollute the soil surface, from where they are carried by surface flow into watercourses.

Besides the washing out of chemical additives, the *chemical leaching* of easily soluble substances including important plant nutrients must also be considered. This refers to the extraction of matter which is soluble in water or weak acids, the latter being formed from precipitation water enriched with dissolved gases, etc., or in larger measure from oxides and chemical contaminants deposited on the surface of the soil or on a cover of snow before the commencement of surface runoff; chemical contaminants may also be brought down from the atmosphere by precipitation. The degree of chemical leaching varies according to the chemical composition of soil, the nature of chemical complexes in the soil, the chemical composition of the rain-water, the duration of exposure of the soil to precipitation water, the temperature, and other conditions. Under some conditions leaching of the soil is considerably high and may represent the main form of erosive degradation.

Chemical erosion is highly intense and at present is on the increase as indicated by the proportions of substances finding their way from the fields into rivers by mechanical and chemical erosion, respectively. Owing to the chemical pollution of water mainly by organic matter from farm fields, a rapid *eutrophication* takes place in waterways with all the undesirable consequences of this for the ecosphere. So much irrefutable evidence and information is available in this respect that the author does not find it necessary to elaborate on this point. The subject of chemical erosion has been introduced for the sake of completeness; contamination of rain from the atmosphere and soil leaching occur, the latter being connected mainly with the effect of surface runoff. Leaching of the soil is extremely important and should receive greater attention not only from a theoretical, but also from a practical point of view.

Besides chemical erosion which is strongly associated with area, mechanical sheet erosion also occurs, as indicated by measurements of turbidity in the surface flow; the turbidity may be high, even when the water flow has not yet gathered into visible channels and there is no effect of raindrops. This problem is mentioned again in Sec. 4.2.5.

4.2.4 Snow thaw erosion

Whereas a great amount of attention is given to drop erosion, especially in recent literature, *snow thaw erosion* is rarely mentioned, even in authoritative reviews. Nevertheless, this form of precipitation erosion plays an important role in certain regions, especially where there is heavy snow precipitation and sudden thawing. The danger of soil erosion from melting snow is greater where snowdrifts have been formed, as for example on mountain ridges, leeward slopes, and in depressions and gullies, etc.

The main feature of snow thaw erosion is the *freezing of the soil* in the cold period, in which water is extracted from the soil aggregates to form small crystals around them. In adition to this, a considerable quantity of water rises from lower horizons into the freezing zone. The ice crystals as they form partially destroy the soil aggregates, so that when the thaw comes a mass of fine soil particles is released. Disaggregation and oversaturation increase, especially in the surface layers, during the *regelation* and retarted flow of snow water on the soil surface; the thawed soil takes on a muddy appearance and is inclined to flow, even in the absence of surface runoff.

Another effect of freezing which increases the erodibility of the soil during the spring is the greatly reduced infiltration rate of snow water into the deeper layers, so that when the soil thaws, starting at the surface, relatively intense soil erosion begins even though the first amounts of snow thaw are small. These erosion processes are accelerated when warm air masses accompanied by rain arrive. Since thawing tends to be more rapid on southern slopes, it is these southern aspects on which the greatest damage to the soil by snow thaw erosion occurs. It should be added that in some regions the processes of slope erosion are more complicated and of very great importance.

In spring the *soil protecting effect of vegetation* is poor and on agricultural arable land the soil is often almost completely bare, or covered only by small plants (e.g. winter cereals). But even on range land the vegetation is sparse, especially on shallow soil in arid regions, and southern slopes are more vulnerable to soil erosion for this reason also.

Interesting data on the erosive effects of snow water in the Valdai region (USSR) are presented by Lidov et al. (1973). They established that erosion by snow water begins on slope inclinations as gentle as 2–3°, the turbidity of the melt water increasing as its erosive activity increases with temperature:

Temperature [°C]	0–1	2	4	6	8	10
Turbidity [g l^{-1}]	1	3	7	15	25	50

At higher temperatures the quantity of water and its velocity, as well as the concentration and sizes of carried particles increase. At a velocity of 0.15 to 0.2 m s^{-1}, particles and aggregates with diameters up to 2 cm were set in motion; at a velocity of 0.3 m s^{-1}, the largest diameter of moving particles increased to 3 cm, and at a velocity of 0.6 m s^{-1}, it was 4–7 cm long. When the water gathers in rills, erosion quickly increases. The average soil loss on slopes with an inclination of 10° during snow thaw in spring was 2 mm, i.e. 30 ton ha^{-1}.

From among older investigations of erosion caused by snow water, a detailed description of the increase of erosion on slopes was given by Kornev (1937), who worked at the Novosil' Soil Conservation Experimental Station established in 1921, and made observations concerning the increase in turbidity of snow water as its kinetic energy became greater. At various distances from the dividing ridge, the turbidity was as follows:

Distance [m]	5	35	280	315	415	450
Turbidity [kg m^{-3}]	1.14	1.30	1.55	2.22	5.70	7.28

From the results of long-term observations he also established the dependence of the intensity of soil erosion on the intensity of precipitation and length of the slope, as expressed by the formula

$$E_0 = AS^{0.75}L^{1.5}i^{1.5},$$

where E_0 is the removal of eroded material [kg s^{-1}], A the coefficient of proportionality dependent upon other factors, S the slope inclination [%], L the length of slope measured from the divide [m], and i the intensity of precipitation [mm min^{-1}].

As well as the intensity of soil erosion, soil properties and the resulting total soil erosion were also affected. In the lower part of the slope, for example, the humus content was generally 2.0 to 2.5 times smaller than at the summit. The smaller influence of slope inclination and the greater influence of slope length resulted from the convex form of the slope. The same dependence has also been established between erosion and the rate of snow water runoff. This relationship is very close to that derived by Zingg (1940) in the USA

$$E = AS^{1.49}L^{1.6},$$

where E is the total removal (cubic feet), A the coefficient of proportionality dependent upon other factors, S the slope inclination (%), and L the slope length (feet).

The calculated quantities of removal material include, of course, all erosion losses caused by precipitation water. Thus the general importance of the influence of slope length on erosion processes is evident, this being explained by the increasing kinetic energy of the water as it flows downward.

In the spring of 1956, on the experimental plot *Radvaň near Banská Bystrica* (CSSR), a measurable degree of erosion caused by snow water was recorded only after a certain amount of water had accumulated and a certain velocity of flow was attained on the way down the slope. Therefore erosion damage occurred not so much on the upper part, but more on the lower part of the slope (Table 30).

Table 30. Soil losses in the neighbourhood of Radvaň (spring 1956)

Distance from the divide [m]	Ground angle of inclination	Soil loss		Area covered by rills [%]
		[m^3 ha^{-1}]	[kg m^{-1}] of rill	
40+	5°43′	—	—	
50+	7°39′	—	—	—
60+	8°47′	—	—	—
70	11°48′	1.65	1.39	0.81
80	11°47′	5.98	2.26	1.76
90	11°43′	4.15	2.68	1.17
100	8°53′	18.28	5.11	4.44
110	5°57′	40.64	7.14	8.44
120	5°48′	27.11	6.53	7.17
130	4°42′	31.23	8.12	6.64
140	6°13′	24.54	13.65	3.43
Mean 60—140 m	8°29′	19.19	5.86	4.23
Mean 0—140 m	6°21′	10.97	—	2.40

The most intense precipitation erosion occurred on the steepest part of the slope, where there was also a deterioration in soil properties (up to a 250-fold change in the value of d_{40}). The soil was permeable with a relatively high skeletal content.

In another locality near *Prešov* (CSSR) in the same year, on an area of less permeable soil, an average soil loss of 30.44 m^3 (45 t) ha^{-1} caused by snow water was recorded (total slope length 235 m; average inclination of slope 8°58′).

Table 31. Soil losses in the neighbourhood of Fintice (spring 1956)

Crop	Distance from the divide [m]	Ground inclination of section	Soil loss		Area covered by rills [%]
			[m^3 ha^{-1}]	[kg m^{-1}] of rill	
1	53	7°27′	—	—	—
2	97	10°35′	32.01	3.54	9.68
3	128	15°23′	27.71	2.57	8.41
4	128	15°23′	13.38	3.34	2.45
5	160	11°52′	29.06	3.02	7.94
6	207	8°29′	104.85	8.02	14.24
7	182	10°41′	39.33	3.48	7.87
8	235	8°58′	30.44	—	6.09

1 — clover and herbs, contour ploughing, 2 — winter crop, contour ploughing, lower part of field, 3 — winter crop, ploughing up and down slope, partial surface deposit, 4 — thin, low clover, ploughing up and down slope, 5 — winter crop contour ploughing and ploughing up and down slope, surface deposit, 6 — maize stubble and winter crop surface deposit, 7 — mean for the eroded part of the slope, 8 — mean for the whole slope.

The first visible signs of erosion occurred at a distance of 97 m from the divide, and the largest loss of about 150 t ha^{-1} was again observed on the lower part of the slope. The mean surface area damaged by rills was 6.09% of the total. Winter cereals and stubble fields represented the type of cultivation on the slope, except for the upper part, on which a sparse crop of clover was growing (Table 31). Similar results were obtained in the neighbourhood of Prešov on two other experimental plots. The average rate of infiltration over a period of 120 min varied from 0.025 to 0.125 mm min^{-1} for arable land, and from 3.0 to 6.0 mm min^{-1} for forested land.

Finally, on the less permeable soils of the neighbourhood of *Sobrance* (CSSR), a pronounced degree of erosion occurred at a relatively short distance from the divide and with only a very gentle slope inclination. An unprotected field lying on a moderately curved slope was selected for the first measurements of erosion losses (Table 32).

The data show that with an average slope inclination of 4°34′, and a slope length of 180 m, erosion losses reached 36.4 m^3 ha^{-1} (55 t ha^{-1}). Erosion losses in the lower part of the slope (angle of inclination 10–15°) reached 350 to 463 m^3 ha^{-1}. No selective influence of erosion on the granulation of the soil was observed in this loess loam. The averagre infiltration rate of water into these soils varies between 0.01 and 0.1 mm min^{-1} on arable land; when the soil is saturated, the infiltration rate may fall to less than 0.01 mm min^{-1}.

Data on erosion losses caused by snow water generally show that losses increase with soil permeability. The less permeable the soil, the lower the losses; where the losses are greater, they also appear on less steeply inclined parts of the slope as well

Table 32. Data on rill erosion of the soil on a strip 60 m wide and 180 m long (spring 1958, Krčava)

Distance from the divide [m]	Ground inclination of section	Soil loss		Area covered by rills [%]
		$[m^3\ ha^{-1}]$	$[dm^3\ m^{-1}]$ of rill	
10	2°16′	1.19	3.58	0.47
20	2°52′	7.96	11.95	1.78
30	3°23′	13.00	11.23	2.68
40	4°34′	28.43	17.06	4.37
60	4°17′	40.31	19.41	5.93
80	4°36′	45.87	24.09	5.50
100	5°25′	48.13	21.86	6.67
120	5°08′	46.18	43.85	3.88
140	4°52′	30.46	40.96	2.63
160	4°53′	40.09	60.40	3.33
180	5°30′	87.64	262.91	3.60
Mean	4°34′	35.43	40.91	3.98

as higher up on the slope. The selective effect of flowing water depends mainly on the mechanical texture of the eroded soil.

The relative importance of snow water erosion may be judged by analyzing bed load flow at various times of the year. As an illustration of conditions existing in central Europe, average monthly flows of bed load in the rivers of Slovakia (CSSR) are shown in Table 33.

The data show that in the Váh, Nitra, Uh and Laborec rivers the flow of bed load reaches a maximum in the spring months, i.e. at the time of snow thaw. A secondary summer maximum appears only in the case of Váh; since the drainage basin of this river is partly of alpine character, the flow appears as it does in the Danube, twice annually (bimodal flow). In the lower placed parts of the catchment areas, an increased bed load also appears in the winter months as a result of thawing. In general, for those rivers having their sources in the Carpathians (Váh, Nitra, Hron, Laborec, and Uh), the seasonal proportions of the total annual flow of bed load are 47.56% for the spring (March, April, May), 21.05% for the summer (June, July, August), 5.1% for the autumn, and 26.3% for the winter. This means that in Slovakia, 74% of the bed load flow occurs in the winter and spring, and 26% in the summer and autumn, a considerable part of the flow taking place during sudden thawing and long periods of rainfall. Thus for example, during flood conditions on the Váh lasting from 12th to 24th April 1956, 12.4% of the annual total water flow and 61% of the annual total bed load were carried by the river. Similarly, 19% of the annual flow of water and 44% of the annual bed load were recorded during the flooding of the Hron river between 13th February and 2nd March 1957. The

Table 33. Mean monthly discharge of bed load in various rivers [kg s^{-1}]

Month	River						
	Dunaj	Morava	Nitra	Váh	Hron	Laborec	Uh
1	46.14	2.19	2.10	4.54	2.22	3.47	7.74
2	189.57	3.92	4.68	17.80	20.18	22.25	15.35
3	395.67	5.51	7.16	51.60	42.85	13.63	13.90
4	212.02	5.86	7.46	68.45	20.92	14.75	24.68
5	272.14	3.36	1.60	18.45	8.55	7.45	12.95
6	398.78	1.71	0.57	18.55	4.20	3.33	8.42
7	734.92	7.15	2.65	60.80	7.59	1.37	5.54
8	288.38	2.14	1.16	12.95	8.48	1.24	2.33
9	86.00	0.80	0.29	3.17	1.48	1.29	3.12
10	82.57	1.41	0.44	5.32	2.55	2.95	3.99
11	29.43	0.65	0.29	4.15	0.67	2.08	2.87
12	79.77	1.88	1.83	28.80	6.97	11.75	24.21
Mean	234.71	3.05	2.51	24.70	10.48	7.03	10.43
Relative monthly maxima and minima [%]*							
Maximum	313.11	234.42	297.21	277.12	408.87	316.50	236.62
Minimum	12.53	21.31	11.55	12.83	6.39	17.63	22.33

*The monthly mean for all months of the year is taken as 100%.

turbidity of the water increased under these conditions to 4 kg m^{-3}, or more.

This phenomenon is not confined to the Western Carpathians, as is shown by data on bed load flow in regions of the USSR published by Lopatin (1952) and other authors. The data and maps of bed load flow show that in the lowland regions of the USSR, a predominant part of the annual bed load flow (70–80%) occurs during spates caused by snow-water. Only a small part of the flow is attributable to periods of high rainfall. Conversely, the rivers of the mountain massifs of south-eastern Europe, the Caucasus, and the Far East show a substantially higher bed load flow; erosion caused by rains of high intensity and long duration is much more prevalent there.

A detailed study of the erosion process under the influence of snow-water and rain-water leads to the conclusion that in the latter regions it is snow thaw erosion that is the more dangerous from the point of view of soil loss, since it involves the transport of larger quantities of loosened particles into the river system, which removes this material far away from its place of origin. The erosive influence of rain is limited to the area of rainfall, and depends on the duration of the rain and the protective cover provided by vegetation. The intensity of erosion during a downpour is usually relatively high, but smaller areas tend to be affected and the soil that is loosened in this process is not necessarily carried into the river system, and may only be moved over short distances. During downpours there may be a total destruction of the soil, a rapid growth of erosion rills, and a devastation of small catchment areas, especially by torrents of water. However, snow-water and rain together affect much larger areas, removing loosened soil to a greater extent, "cleaning the river-bed", and causing wide-spread river erosion.

In the next section, some consideration is given to rill erosion which develops after sheet erosion and is caused by the confluence of rain-water and snow-water as it flows down the slope. In the author's opinion rills are essentially small ditches (fossettes), and therefore this form of erosion is placed in the category of sheet erosion. However, no objection is voiced against treating drop erosion, sheet erosion, wash erosion, and rill erosion as independent forms, each of which may be of primary importance under different conditions.

4.2.5 Rill erosion

4.2.5.1 General

As we have already seen, soil erosion is largely influenced by the disaggregating effect of rain, as a result of which large amounts of material are released, although this material may only be transported for relatively short distances. On the other hand, the flow of water over the surface has a smaller effect on soil decomposition, but a larger transportation effect. Yet flowing water, especially on tilled land, can become the agent of transport of particles loosened mechanically, chemically, or by means other than the water flow itself, and therefore it is a phenomenon of great importance from the point of view of total soil losses.

The total erosive effect of flowing water suddenly increases when the confluence of surface waters takes place. The smaller the rate of infiltration of water into the soil, and the greater the precipitation and/or snow thaw, the sooner the surface waters gather into rills. As this gathering of water proceeds, the total amount of water remaining the same, the depth of the water increases, together with the velocity, kinetic energy and carrying capacity of the water. At high precipitation intensities there is greater clogging of pores, and the proportion of precipitation water making up the surface flow, and the numbers of particles separated from the soil by raindrops, both increase. The ratio of the amount of soil released by raindrops to the degree of confluence of the surface flow varies with time from the start of the rain and runoff, and also according to soil conditions, topography, climate, etc.

In order to illustrate this diversity in the erosive influence of precipitation, let us look at some examples of denuded slopes exposed during excavations of road cuttings. In all cases the soil was destroyed by violent downpours.

Figure 86 shows the slope of a road constructed on very *permeable soil* and *low resistance*. Althought the slope is very steep and very heavy rain has affected the soil, clear and regular rills have not been formed. The shapes of the depressions are irregular, being determined by the different resistances of the various components; the impact of raindrops is clearly visible, and depressions and microsuffosis forms have developed. Despite the steep incline, rill erosion is not typically represented.

Even more irregular forms arise on lateritic soils with a high content of iron nuggets (containing about 55% Fe, 1.2% Cr, 0.8% wolfram-vanadium). On the surface the *resistant* and less easily eroded components determine the form expressed. Consequently, the eroded surface is not levelled by raindrop action and surface runoff, and is divided instead into small pyramids and similar forms. Erosion channels are not formed, in spite of the steep slope and the very heavy rainfall (Fig. 87).

This series of examples includes one, in which the soil is covered by a layer of

Fig. 86. Slope erosion caused by heavy rain on a road excavation site on permeable material (Ghana). (Photo J. Jeník.)

magnesite dust, this encrustation protecting the soil from erosion. Contamination of the soil from industrial pollution produces different combinations of erosion processes. Figure 62 shows a plot in the first stages of pollution, with areas in which the vegetation has been destroyed by magnesite dust and the soil has been rapidly eroded during downpours. On the more resistant material, rill erosion prevails. Sheet erosion was the predominant form of erosion until a coarsely grained gravel layer was left on the soil surface. In Fig. 88 it can be seen that in the next stage of

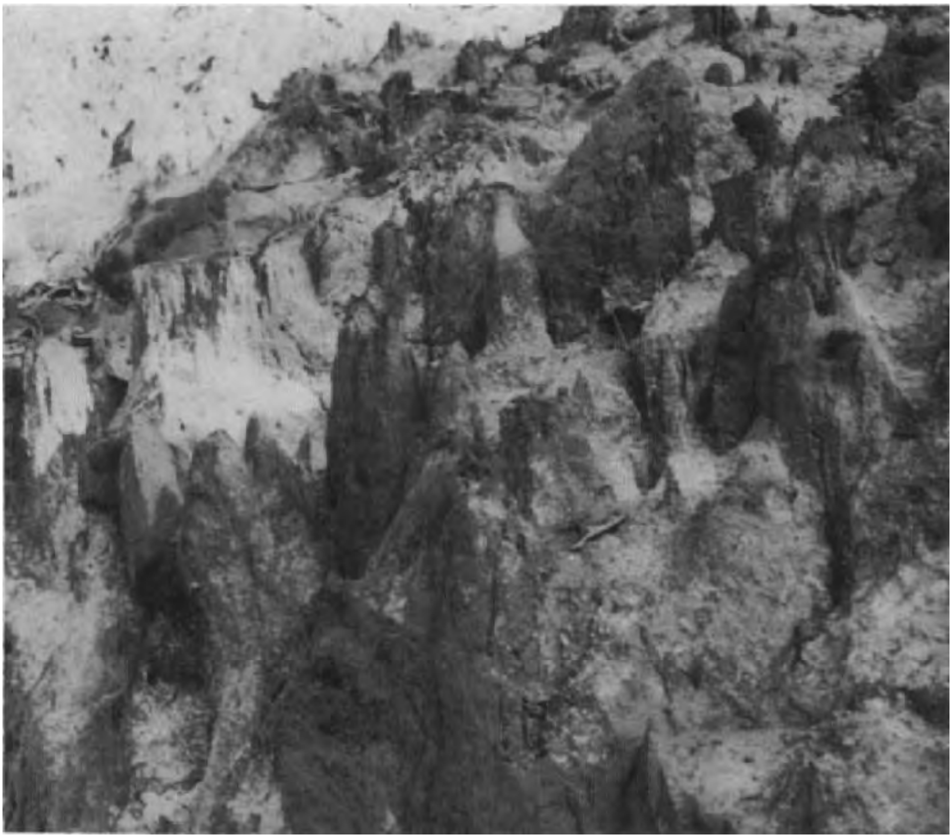


Fig. 87. Lateritic soils with iron nuggets eroded by severe rainstorms (Cuba). (Photo R. Leontovyč.)

intensive soil pollution the washing of the less resistant components of the magnesite crust proceeds, and micropyramids of the more resistant components are formed on the surface.

Erosion processes on *more homogeneous* and *less permeable soils* take a different course. Figure 89 shows the slope of a railway cutting which is eroded by sharp rills. In spite of the skeletal character of the soil, there has been intense rill formation as a consequence of heavy downpours. In this case, the erosive effect of the flowing water is much higher than that of the raindrops.

Another example, in which there is an even greater predominance of rill erosion over other forms, is given in Fig. 90, which shows the slope of a motorway built on impermeable material consisting of younger sediments which are susceptible to erosion. As can be seen, rill erosion has prevailed and affected the whole length of the slope, which means that precipitation water, as soon as it reaches the soil, flows away through the dense network of rills, virtually cutting the slope into thin plates

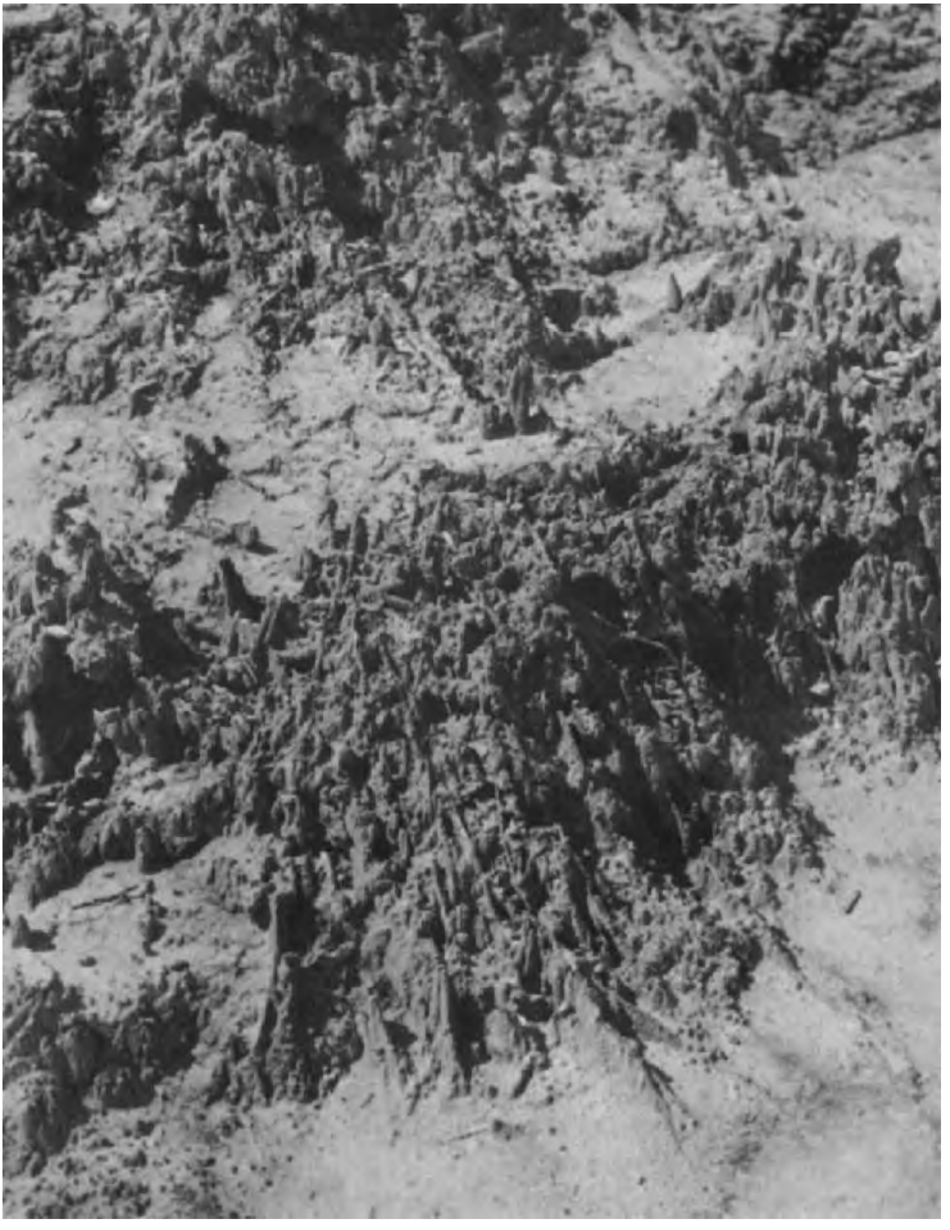


Fig. 88. Soil surface covered by magnesite dust. Micropyramids and other formations are the result of the erosion action (Czechoslovakia). (Photo A. Löffler.)

Fig. 89. Railway excavation slope consisting of loamy to sandy gravel material destroyed by rill erosion (Bulgaria). (Photo D. Zachar.)

and forming pointed pyramids in the upper reaches. The soilscape shown in this picture gives no indication, that either splash erosion or wash erosion have occurred, or that these forms may have been of some importance, and it would, of course, be erroneous to conclude that these forms have been absent.

It can also be seen in the previously mentioned figures that flowing water has a tendency to collect in channels, or other hollow forms. This process is also evident in the *slow corrosion of limestone* (Fig. 1), and still more evident in the *rapid chemical erosion of salt layers* (Fig. 2) where channelling is linked to the formation of micropyramids, lamellae, pipes, and other forms. In these cases, it appears as if there is no surface polishing by the water, which nevertheless still has a solvent effect. The dissolving process proceeds more rapidly, of course, where solutions of high salt concentration are quickly washed away by large flows of water.

Similar phenomena occur on steep slopes, even on impermeable loamy clay material. A very clear illustration of this is given in Fig. 31, which shows that on

Fig. 90. Slope fragmented by rill erosion in the cutting of a motorway through easily erodible sediments. (Photo A. Freininger.)

slopes of relatively homogeneous material furrowed by deep cuts, the walls of the cuts are modelled by rain-water mostly ir the form of rilling. Rill forms occur within a very short distance of the "divide" and rapidly carve into the bedrock.

On impermeable, or still heavier and more resistant material, rill erosion no longer creates pipes, but forms ridges which are separated by sharply cut rillets and gullies. The rillets are occasionally so narrow that they resemble cracks, and therefore one could speak of *crack erosion* – as the antecedent of tunnel erosion. Figure 24 shows this type of modelling on young, heavy sediments in the "Red Mountains" of Romania. On steep slopes composed of material of varying resistance, vertical openings are formed, and these soon develop into tunnel erosion or hollow erosion, separating the washed forms into isolated pipes, etc. Where the material is more homogeneous and the incline less steep, rilling prevails (Fig. 48).

If the material is more coarsely grained and less resistant, the geometry of the rills changes. Flowing water carries the soil along rapidly, and create triangular or trough-shaped forms with respect to the cross-section of the channel. In such cases, the lengths of the rills are greater, but the inter-rill lamellae are thinner, and the edges sharper (Fig. 32). The more coarsely grained and more permeable the material, the less pronounced is the channelling, until finally the rills are widely shaped, and resemble more the form of moderately undulating depressions, even on very steep parts of the eroded slope (Fig. 64).

On permeable, coarse-grained, non-resistant fluvioglacial deposits, on the other hand, shallow, rapidly growing rills develop with an immense production of silt. Here, the action of flowing water is the predominant force, often being associated with soil flow (aquasolifluction) and the conversion of gullies into ravines. The range of forms is again very varied (see Chapter 2). An example of rills developing into gullies and ravines in fluvioglacial material is shown in Fig. 91; as can be seen, the zone of surface erosion is narrow, and the development of shallow rills gives way after a short distance to the formation of gullies.

Should the *fluvioglacial* sediments be more resistant, well-expressed rills develop and pyramids are formed in more protected areas. The creation of earth pyramids is a typical feature, especially if these contain a lot of cementing substance (Fig. 56).

It may be stated, in general, that the more permeable the soil, the more distant are the rills from the head of the slope, and the less pronounced are the rills on the upper part of the slope. On shallow rendzina soil originating from carbonate (dolomitic limestone) bedrock, this rill distance varies from 1 to 10 m. Sheet erosion associated with the erosive action of raindrops is clearly expressed on the shallower soil and on the upper part of the slope (Figs. 55b, 63). These forms are already transitional towards gully erosion, but the contributions made by the various initial forms of precipitation erosion can still be observed. On very permeable, coarse-grained, stony material, erosion rills seldom occur (Fig. 50).

Fig. 91. Erosion of a slope consisting of fluvioglacial material of low resistance. Rapid growth of rills into gullies caused by the predominant action of channelled water (Galina-Bach basin in the Austrian Alps). (Photo D. Zachar.)

Fig. 92. Laterite soil affected by gully erosion; the initial raindrop, rainwash and rill erosion forms are no longer in evidence (Oriente Province, Sierra de Nipe, Cuba). (Photo V. Samek.)

Finally, the creation of linear erosion forms depends also on the *macrostructure of the soil profile and its substrata.* The forms of microcanyons developing on some types of lateritic soils are instructive in this respect; Fig. 92 shows a well-developed vertical soil macrostructure. In these cases, where there is a high resistance to erosion, the erosive effects of raindrops and surface runoff are relatively small, and vertical erosion occurring by virtue of vertical cracks through the soil becomes important. In this case the intensity of the erosion process depends mainly on the erosive action of flowing water. The structure of these soils can be seen in the open horizon in Fig. 93.

Many more examples could be given covering the whole range of situations between the complete predominance of rill erosion and the replacement of rill erosion entirely by other forms. This natural diversity occurs not only with respect to the processes of precipitation, runoff and climatic conditions, but also to the *properties of the soil;* thus under the same conditions of climate and topography, an immensely varied range of rill erosion forms and their associations with other erosion processes can be found. In one case, the erosive effect of precipitation may predominate, in another surface forms prevail, and in yet another, the confluence of surface flow from precipitation and snow water plays a major role.

Fig. 93. Horizont of laterite soil with expressive polygonal macrostructure exposed by erosion (Oriente Province, Cuba). (Photo V. Samek.)

According to the author's observations, rill erosion usually begins to appear in the lower part of the slope (as mentioned in Sec. 4.2.5). This is true especially when the source of the water is thawing snow or precipitation of low intensity. As soon as the intensity of the rainfall increases, the intensity and velocity of surface runoff both increase also, and consequently the proportion of the total erosion due to rills becomes greater, depending on the permeability of the soil. Thus at the outset of a period of rain, a certain amount of water is consumed in the moistening of the surface; after this, disaggregation and splashing of the soil, clogging of the pores, and a decrease in the rate of infiltration take place. Finally, most of the rain-water takes part in the erosive action of surface flow.

The question arises as to how the erosive action of raindrops is related to that of chanelled surface flow. To provide an answer is difficult, but since this is important for the understanding of the erosion process, the author wishes to discuss the subject in depth, using as a basis data on the erosive effect of thawed snow water acting in the absence of the effect of raindrops (Sec. 4.2.4). Perhaps it will not be out of place to give some examples of rill erosion in Czechoslovakia, i.e. examples occurring under the same conditions as those prevailing when the data on snow water erosion were collected.

Fig. 94. Part of an area of land ploughed across the slope and damaged by catastrophic erosion. The terrain features a slight depression, into which water flows from the higher up situated pastures. Sections of terraces protected by shrub vegetation were damaged eight times less severely (Lučatín near Banská Bystrica, Czechoslovakia). (Photo D. Zachar.)

4.2.5.2 The communities of Hiadeľ and Lučatín

The *first series of plots* on which measurements have been made of rills that formed during a violent downpour and hailstorm (Fig. 94), were situated in the communities of *Hiadeľ* and *Lučatín* near Banská Bystrica (Low Tatras region). The erosion in this case was caused by a cloudburst on 23rd May 1958 covering an area of about 18 km^2. This torrential rain wreacked catastrophic damage, with much local soil destruction. With regard to rill erosion, some data illustrating the importance of the erosive effect of flowing water are presented here.

The first measurements were made on fields *forming terraces* one above the other, and the rill volume was measured on a strip of land supporting a crop of oats; the strip was 11 m wide and above it, there was a 25 m wide strip of grass. The mean angle of inclination of the field was 17°30′. The crop of oats was not mature and covered about 60% of the ground surface. The rills were evenly distributed over the field – being rather shallow on the upper section, and deeper on the lower section. The removal of material was about the same in the upper and lower parts of the field, because a large amount of water with a small silt content flowed down from the upper part of the field with its grassy cover. Within the grass stand, rill erosion amounted to almost 3 $m^3\ ha^{-1}$, while in the lower part with oats covering an area of about 200 m^3, about 25% of the surface was damaged by rills (Table 34).

A different situation occurred on the second plot set out on a convex-concave slope with a potato crop and a rye *(Secale cereale)* crop. The field was tilled up and down the slope; the potato crop provided a somewhat inadequate ground cover and the rye crop covered the ground fully. Water was flowing down towards the field from a higher, 16 m wide strip of land with sparse grazing. Data taken from this plot are given in Table 35.

It can be seen from these data that on the potato field very intense rill erosion occurred, even in the upper parts of the field. With increasing distance down the slope, erosion increased as far as 50 m from the top, although the angle of

Table 34. Data on rill erosion of the soil near the community of Hiadeľ (slope angle of inclination 17°30′)

Parameter of rill erosion	Upper part of field	Lower part of field
Soil loss [$m^2\ ha^{-1}$]	206.83	197.07
Area damaged by rills [%]	27.66	24.17
Mean depth of rills [mm]	74.77	81.54
Soil loss per 1 m rill length [dm^3]	13.14	14.47

Table 35. Data on rill erosion of the soil on a potato and rye field, Hiadeľ in 1958

Strip	Distance [m]	Ground inclination	Potatoes Loss [$m^3\ ha^{-1}$]	Potatoes Damaged area [%]	Rye Loss [$m^3\ ha^{-1}$]	Rye Damaged area [%]
I	10	10°	209.00	26.31	—	—
II	20	15°	255.85	26.62	—	—
III	30	20°	282.85	33.85	12.77	3.85
IV	40	18°	318.15	31.85	17.08	5.62
V	50	16°	329.53	30.15	9.69	4.31
VI	60	12°	193.54	19.08	3.08	1.53
VII	70	10°	62.15	15.84	—	—
I—VII	70	14°26′	235.87	26.24	10.65	3.84

Fig. 95. View of a potato field very severely damaged by erosion. Neighbouring fields have rye crope *(Secale cereale)*. Above the fields there is a pasture from which water flows downwards (Lučatín near Banská Bystrica, Czechoslovakia). (Photo D. Zachar.)

inclination decreased by 4° between the 30 m and 50 m intervals, thus indicating that the erosion was a function of the kinetic energy of the surface water and its carrying capacity. A sudden decline in the rate of erosion occurred only after the water became saturated with silt, and for the same angle of inclination (10°), the rate of erosion was 147 m^3 ha^{-1} less in the lower part of the slope than the corresponding rate in the upper part. In the rye plot, a decline in the rate of erosion occurred only where there was a sudden decrease in the angle of slope. The average rill depth was 87 mm on the potato field, 29 mm on the rye field, and the corresponding rill volumes per 1 m length were 17 and 2 dm^3, respectively. Rills appeared in the rye field at a 16° incline, and disappeared at a 10° incline.

In another location, measurements showed that as a consequence of water flowing down from a 150 m wide pasture on to a potato field, the soil of this field was almost totally destroyed, and at a slope inclination of between 13 and 15°, 1,104 m^3 of soil per ha were removed, 60% of the surface area was damaged by rills, the mean rill depth was 184 mm, and the rill volume was 113 m^3 per 1 m length (Fig. 95).

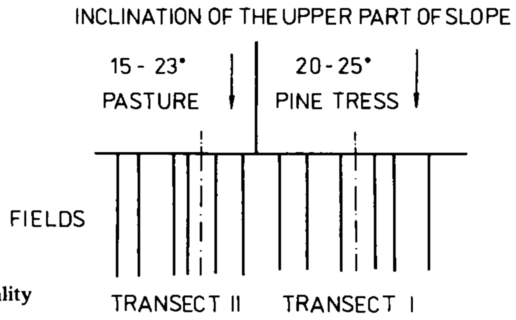


Fig. 96. Scheme of crop distribution on the locality Lučatín (Czechoslovakia).

An interesting pattern of rill erosion was observed on a concave slope in the community of Lučatín. On the upper reaches of the steeper part of the slope, there was a stand of pine trees and on a more gentle slope situated above fields, there was a pasture which showed low rates of infiltration. On the upper part of the slope, a 15 m wide strip of derelict pasture separated the forest from the fields (Fig. 96). In both areas the fields were oriented with the longer axis parallel to the line of greatest slope. Data on rill erosion on the upper and lower areas are given in Table 36 (transection I) and Table 37 (transection II), respectively.

The pattern of erosion on the upper plot was similar to that in the community of Hiadeľ, with the difference that lower down on this area the soil was less permeable, and consequently, erosion was still observed to occur at an inclination of 2°. As soon as water began to flow down from the higher lying pasture, erosion on the lower field suddenly increased considerably and totalled more than 1,000 m^3 ha^{-1} on a potato field (Fig. 94). But on a neighbouring wheat field losses were relatively low, indicating the protective effect of vegetation as a result of the

Table 36. Data on rill erosion of the soil on a potato field within a clover stand, Lučatín in 1958

Distance [m]	Ground inclination	Potatoes			Clover		
		Soil loss [m^3 ha^{-1}]	Damaged area [%]	Soil loss along rills [dm^3 m^{-1}]	Soil loss [m^3 ha^{-1}]	Damaged area [%]	Soil loss along rills [dm^3 m^{-1}]
10	20°	217.36	27.45	13.28	23.54	6.50	2.83
20	20°	218.06	30.25	13.65	22.96	6.17	2.76
30	18°	237.83	27.22	15.36	9.66	3.17	2.32
40	17°	324.43	31.67	19.47	2.67	1.58	1.60
50	15°	205.06	25.44	13.18	1.50	0.85	0.90
60	13°	211.17	26.44	14.66	—	—	—
70	12°	214.73	24.00	14.17	—	—	—
80	10°	219.55	23.32	17.96	—	—	—
90	8°	131.44	15.43	15.68	—	—	—
100	6°	120.22	13.00	13.53	—	—	—
110	4°	48.00	10.00	10.80	—	—	—
120	2°	31.44	5.56	9.43	—	—	—
0—120	12°	182.15	21.65	14.26	5.03	1.52	0.87

Table 37. Rill erosion on a potato and wheat field, Lučatín in 1958

Strip	Distance [m]	Ground inclination	Potatoes			Wheat		
			Soil loss [m^3 ha^{-1}]	Damaged area [%]	Soil loss per m of rills [dm^3]	Soil loss [m ha^{-1}]	Damaged area [%]	Soil loss per m of rills [dm^3]
I	5	14°	1,081.55	63.82	118.97	89.97	17.27	8.73
II	35	10°	986.82	49.09	169.91	8.85	2.67	3.65
III	65	8°	668.64	42.27	122.58	7.56	2.18	3.10
IV	95	6°	257.97	24.09	47.29	2.24	1.09	1.85
V	125	3°	120.45	8.73	44.10	—	—	—
I—V	5—125	8°	623.09	37.60	100.57	21.72	4.64	3.47

braking effect of the plant shoots above ground, and the mechanical binding of the soil by the plant roots; this view of soil protection by plants is rather different from that put forward in the theory of splash erosion, in which the soil is said to be protected from the kinetic energy of raindrops.

The author considers these results to be important aids in the assessment of splash and rill erosion under given conditions, because they cannot be simulated under laboratory conditions, nor can they be derived by theoretical analysis, no matter how well established.

For the sake of completeness it should be added that the soil on the plots was coarse-grained with a skeletal content between 5 and 25%, the mean grain diameter d_{40} at the soil surface having increased by up to 300 times.

In spite of this, the rate of infiltration on the pastures, established by cylindrical infiltrometers, did not exceed an average value of 0.5 mm min^{-1} over a period of 120 minutes, and varied between 0.1 and 0.2 mm min^{-1}, whereas in the neighbouring pine stand it varied between 6.5 and 9.1 mm min^{-1}.

Also on this plot some *furrowing strip erosion* occurred (Figs. 22, 82) which entirely removed the topsoil; this happened either in depressions where a large flow of water was collected from the surrounding area, or along fracture lines where there was a sudden increase in the angle of inclination of the slope. In the latter situations large amounts of chanelled water gained increasing velocity together with an extraordinary erosion force, although the sheet runoff was taking place over the surface without channelling (Fig. 97). In this case no sorting of material occurred in situ.

This form of erosion *(l'érosion en nappes ravinantes)* was discovered by Fournier (1956) in West Africa. Fournier pointed out that this form had been observed (a) in heavily eroded regions where the lower horizon contained clay and the upper layer was relatively thin, (b) in regions where the upper layers were easily and rapidly saturated, and (c) in situations where a hardened layer had developed near the surface. It should be added that, if under this layer there is a settled layer of resistant subsoil, the loosening of the topsoil by shallow ploughing (i.e. up to a maximum depth of 25 cm), is according to the author's experience of great importance. This form of erosion occurs only locally during violent downpours, and provides an example of the considerable erosive effect of sheet runoff as indicated also by the various surface erosion phenomena that are caused by flood water.

After a detailed survey of the territory, the author established that in the spring of 1957 and the autumn of 1958 erosion was greatest on those soils which were freshly ploughed. On unploughed areas, even though the slope was very steep, erosion was substantially less, varying from nil to 200 m^3 ha^{-1}. Similarly, on subsoil which was exposed by laminar erosion, further erosion was much weaker and erosion rills occurred only on roads and in those places on which water converged in enormous quantities (Fig. 98).

Fig. 97. Laminar sheet erosion of the soil; the unprotected topsoil is completely washed away. (Photo D. Zachar.)

Fig. 98. Road washed by a downpour in the Lučatín community (Czechoslovakia). The gully was 2 m deep in some places. (Photo D. Zachar.)

By making a detailed analysis of erosion processes (*Erózia pôdy,* Zachar 1970), the author observed that although the impact effect of raindrops may be considerable, especially when augmented by hail, it is nevertheless rill erosion, and not splash erosion, that is the predominaat force in soil erosion. This conclusion is typically illustrated in Fig. 99, which shows a part of the excavation carried out for a new road near the community of Lučatín. Only a part of the slope had been stabilized by fences, and when a downpour came, the surface water entirely destroyed the lower part of the slope.

Splash erosion played an important part both in the destruction of the soil surface and in the transport of soil over short distances; wash erosion was responsible for the flush of loosened soil into rills, but only during heavy sheet runoff was the effect of wash erosion in any way similar to that of rill erosion. The data show that the total loss of soil was associated mainly with rill dimensions, the

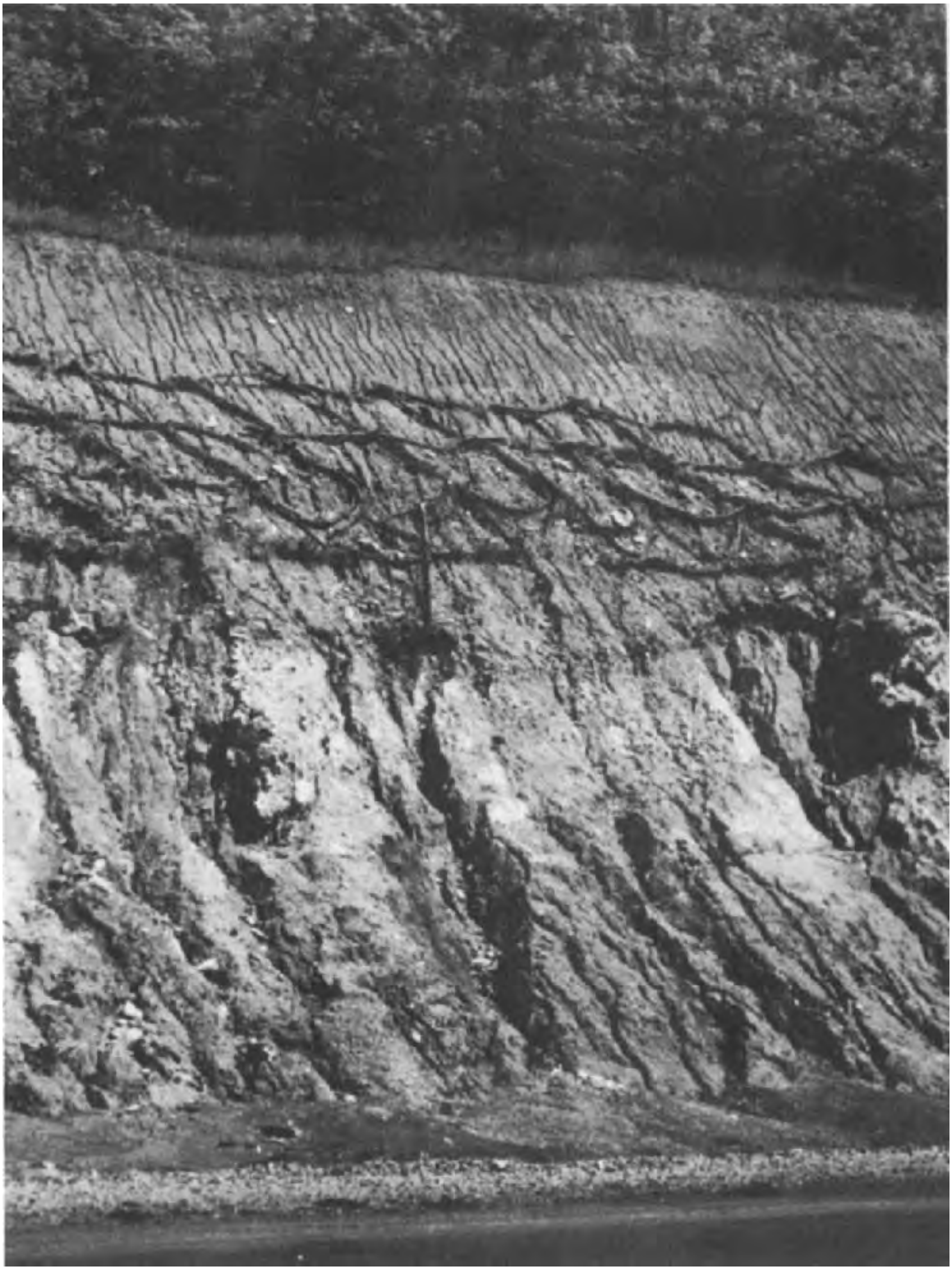


Fig. 99. Part of a slope in a road excavation damaged during a downpour. In the lower part of the excavation the pronounced effects of surface runoff can be seen (Lučatín near Banská Bystrica, Czechoslovakia). (Photo D. Zachar.)

depth of rills formed on arable land being strongly limited by the thickness of the topsoil. Only in furrows, where water converged and accumulated more strongly were soil losses substantially higher. The amount of splash and sheet erosion, as a proportion of the total erosion on this plot, was estimated to be in the range 5 to 30%. The lowest values were obtained on fields endangered by external sources of water. Wash erosion, in this case, had a very great selective effect, as indicated by the coarse mechanical composition of the soil after the heavy hailstorm.

4.2.5.3 The Hriňová dam

The *Hriňová dam* which serves as a municipal water supply was another location selected for the investigation of rill erosion. The dam is situated approximately in the same climatic region as the territory discussed in the previous section. The mean annual total precipitation is 797 mm, and the mean annual temperature is 7.5°C. A new road was constructed along the dam, and in the belt between the road and the water, the trees were removed together with the humus layer, the intention being to stabilize the soil by sowing grass. However, before this was done there was some very heavy rainfall, the precipitation for June, July, and August amounting to 326 mm. The summer rainstorms caused excessive rill erosion on the exposed soil and this was mapped by the author in September 1966. From the measurements that were made, only selected data concerning the rills are cited, in order to give an indication of the relationship between rill erosion and the relief.

Table 38. Data on rill erosion on experimental plot I near the Hriňová dam

Transection	Distance from the upper edge of the plot [m]	Ground inclination	Soil loss [m^3 ha^{-1}]	Area of rills [%]	Means	
					Length of rills [cm]	Density of rills [km ha^{-1}]
1	0	20°	130.4	13.3	285.7	3.5
2	10	20°	278.3	25.4	66.6	15.0
3	20	21°	448.0	23.2	111.1	9.0
4	30	21°	671.2	30.9	66.6	15.0
5	40	20°	630.2	41.3	51.3	19.5
Mean	40	20°	431.6	26.8	116.2	12.4

The *first transection* was established on a straight slope with an average inclination of 20°15′ and a length of 40 m. The data relating to the rills are given in Table 38. They show that on the straight slope, which consisted of relatively resistant loamy to sandy material, the volume of rills increased according to rill length and

Fig. 100. Rills formed on unprotected, bare soil of low permeability on moderate gradient (Hriňová dam, Czechoslovakia). (Photo D. Zachar.)

Fig. 101. As Fig. 100, on steep slope. (Photo D. Zachar.)

only the reverse effect of deposition in the region of the water level caused a moderate decline of erosive activity, as indicated by the reduced rill depth and the reduced vertical erosion. The average loss of 432 m^3 of soil per ha which occurred over the growing season can be considered as unusually high, having been a result of the confluence of the surface runoff from rainstorms. As can be seen in Figs. 100 and 101, the first two stages of rain erosion, namely splash and wash erosion, made only a small contribution to the total soil loss.

The *second transection* lay on a more gentle slope with an average angle of inclination of 13°30′, but with the same width of 40 m. From Table 39 it can be seen that as the angle of inclination decreased from 20 to 13°, the average soil loss declined from 432 to 229 m^3 ha^{-1}. However the influence of the latter soon diminished and the pattern of erosion then depended only on the increased kinetic energy of the surface runoff which had converged in the rills. A sudden decrease in the rate of erosion at a distance of 30 m was partly attributable to a resistant gravel layer, but mainly to the decline in the steepness of the slope.

Table 39. Data on rill erosion experimental plot II near the Hriňová dam

Transection	Distance from the upper edge of the plot [m]	Ground inclination	Soil loss [m^3 ha^{-1}]	Area of rills [%]	Means	
					Length of rills [cm]	Density of rills [km ha^{-1}]
1	0	13°	95.9	15.9	142.8	7.0
2	5	14°	35.5	7.9	333.3	3.0
3	10	14°	73.0	13.7	200.0	5.0
4	15	14°	367.7	29.6	83.3	12.0
5	20	14°	394.8	35.8	58.8	17.0
6	25	15.°	482.5	34.6	71.4	14.0
7	30	12°	167.5	22.8	55.5	18.0
8	35	14°	243.9	27.4	50.0	20.0
9	40	12°	201.9	22.2	58.8	17.0
Mean	40	13°30′	229.2	23.3	117.1	12.5

By comparing the two transections it becomes clear that with increasing steepness of the slope, rill dimensions increase also:

Size of rills [cm]	Transection	
	I	II
Average minimum	8.4	9.6
Average maximum	67.7	41.3
Average minimum	3.2	2.1
Average maximum	35.1	15.9

Table 40. Data on rill erosion on experimental plot III near the Hriňová dam

Transection	Distance from the upper edge of the plot [m]	Ground inclination	Soil loss [m^3 ha^{-1}]	Area of rills [%]	Means	
					Length of rills [cm]	Density of rills [km ha^{-1}]
1	0	10°	235.7	24.1	80.0	12.5
2	10	16°	309.8	25.8	90.9	11.0
3	20	17°	533.2	29.0	80.0	12.5
4	30	19.°	557.1	30.7	76.9	13.0
5	40	19°	821.1	31.2	117.6	8.5
6	50	15°	320.5	19.6	200.0	5.0
7	60	10°	287.7	16.8	166.6	6.0
8	70	11°	140.1	12.9	285.7	3.5
Mean	70	14°30′	400.6	23.8	137.2	9.0

Two further transections lay on less resistant material with a higher content of sand and gravel. In addition, the earth layer which had developed on the surface had been loosened by ploughing, and grass seed had been sown on it, unfortunately too late to stabilize the soil before the rainy season arrival.

Data form the *third transection* with its convex and concave profile are given in Table 40. Here also, the influx of water from the oblique road bed enhanced erosion, which totalled 821 m^3 ha^{-1} on the curved part of the slope. With declining steepness of the slope, the soil losses rapidly decreased, although the average loss of 400 m^3 ha^{-1} for an average inclination of 14°30′ was still relatively high. When comparing soil losses at a distance of 40 m down the slope, the following values were obtained for the four transections:

	Angle of inclination	Erosion loss [m^3 ha^{-1}]
Transection I	20°	432
Transection II	13°	229
Transection III	16°	491
Transection IV	16°	339

In the third transection there was a decline in the erosive activity of surface water, which was explained by the greater amount of silt in the water.

The *fourth transection* also had a convex and concave profile, and crossed a bedrock of similar type to that under the third transection (Table 41). A reduction in the rate of soil removal occurred, owing to the fact that scattered tufts of

Table 41. Data on rill erosion on experimental plot IV near the Hriňová dam

Transection	Distance from the upper edge of the plot [m]	Ground inclination	Soil loss [m^3 ha^{-1}]	Area of rills [%]	Means	
					Length of rills [cm]	Density of rills [km ha^{-1}]
1	0	15°	205.7	22.6	100.0	10.5
2	10	15°	312.0	21.7	90.9	11.0
3	20	17°	240.1	23.1	83.3	12.0
4	30	18°	266.8	23.5	105.2	9.5
5	40	17°	671.4	32.4	111.1	9.0
6	50	12°	420.5	24.2	133.3	7.5
7	60	6°	497.7	21.8	142.8	7.0
8	70	8°	397.2	27.4	222.2	4.5
9	80	6°	166.4	22.9	200.0	5.0
10	90	11°	437.4	27.0	142.8	7.0
Mean	90	12°30′	361.5	24.7	133.2	8.25

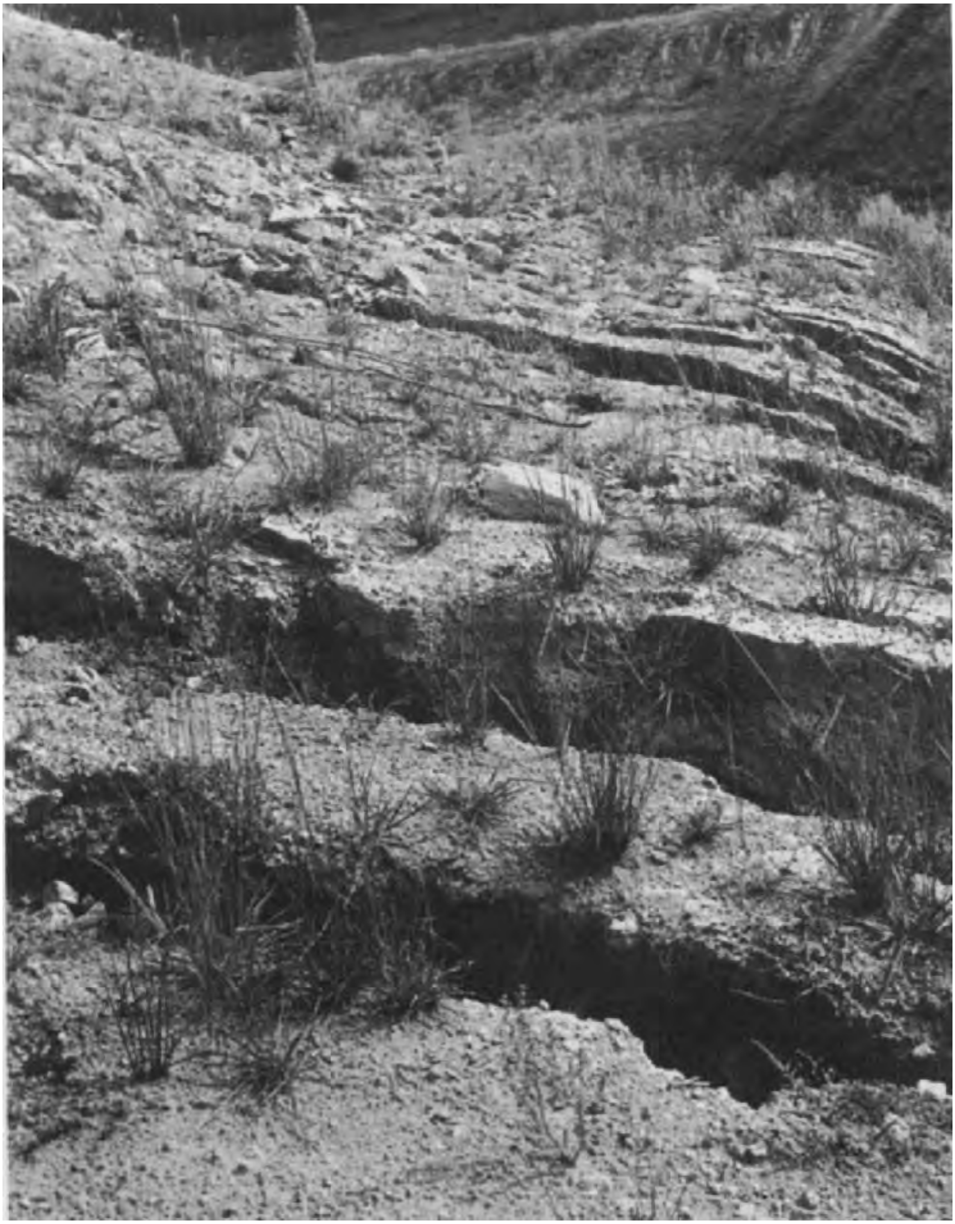


Fig. 102. Area of coarse-grained to skeleton-like material affected by rill erosion (Hriňová dam, Czechoslovakia). (Photo D. Zachar.)

grass had survived and slowed down erosion in some places (Fig. 102). An unexpected result from this transection was the increase in loss recorded in the last measuring position; this was explained by the observation, that on the upper part of the slope where the angle of inclination had dropped to 6°, the water deposited some silt, and as soon as the angle of inclination increased again, erosion activity also increased. This phenomenon was observed by the author in the Hiadeľ and Lučatín communities also.

These processes are described here in detail in order to illustrate how, under particular conditions, the intensity of rill erosion depends very largely on the kinetic energy and transporting capacity of flowing surface water rather than on the impact of raindrops, the effect of which is far smaller than that of surface water.

It is interesting to note the effect of *drop* and *sheet erosion on soil texture;* in the case of the loamy soil (transections I and II), the change was a decrease in the first granular fraction (particles with a diameter of up to 0.01 mm) from 53 to 39%, but the d_{50} value only increased twofold in the shallow upper layer. The fine earth content did not change. On more coarse-grained soil (transections II and IV), the changes in texture were more pronounced. The first fraction decreased from 46 to 9%, the fine earth content dropped from 80.1 to 30.4%, and the d_{50} value increased from 0.05 to 2.6 mm – a 52-fold (and locally as much as a 200-fold) increase.

In this connection, the author would like to point out that the coarser the soil, the greater is the selective effect of both raindrops and sheet runoff, but then a "protective layer" of small gravel and stones forms more quickly and this prevents further splashing and wash erosion; in such cases conspicuous erosion only occurs if surface water converges to from channels. Referring to the soil surface and the form of the rills in Figs. 100 and 101, it can be seen that under entirely identical downpour conditions, different erosion forms occurred according to the various local soil properties. On loamy soil the surface features were rounded off and rill volumes were difficult to measure, whereas on coarse-grained soil the rill edges were sharp and gravel and stones were much more in evidence on the bottoms of the rills. The losses caused by rill erosion were enormous in both cases.

4.2.5.4 The viticultural region of the Little Carpathians

Further understanding of the relative significance of raindrop, wash and rill erosion may be derived from a study of the *vine-growing region* of the Little Carpathians. The plot chosen by the author was situated in the vineyards of the Myslenice community near Bratislava. This is a region of low annual precipitation, although noted for its frequently occurring summer downpours and a rather skeletal soil. Without going into details, the author would like to describe the

Fig. 103. A newly established vineyard in Myslenice near Bratislava (Czechoslovakia) damaged by erosion. During a downpour, deposits accumulated on the foot of the erosion rill. (Photo D. Zachar.)

outcome of a downpour which occurred on 14th June 1966 with total rainfall of about 75 mm (Meteorological Station in Limbach). Soil losses were as follows: in a newly established vineyard with an 8 to 10° slope – 300–500 m^3 ha^{-1}; in a vineyeard established 5–7 years previously – only 50–120 m^3 ha^{-1} (i.e. 5–6 times less removal of soil). Thus erosion was greater where the soil was loosened and where no "protective" skeleton layer had yet been formed. On bare soil erosion rills developed on a gentle inclination of 2° and had a volume of about 30 m^3. In places where water converged in larger quantities, soil removal exceeding 1,000 m^3 ha^{-1} was measured (Fig. 103). In scattered locations the rills were found to be from 30 to 50 cm deep, and on convex areas of the slope practically all the topsoil was removed.

In terms of erosion, the greatest damage was caused by cultivation rills which ran obliquely across the slope collecting large amounts of water. Within the critical

Fig. 104. Deposits on the lower part of the slope in a newly established vineyard (Myslenice near Bratislava, Czechoslovakia). (Photo D. Zachar.)

distance the crests between rills broke and the water created deeper rills. A section of the lower part of the slope with oblique cultivation rills is shown in Fig. 29. It can also be seen that raindrop splashing has displaced soil into the rills, whereas on wider strips on moderately steep inclines the soil was displaced a short distance by raindrop action (centre of Fig. 29). In this case when the downpour yielded, the rills became choked, and in the lower part of the slope the depth of deposits reached 30 cm, up to a maximum of 55 cm (Fig. 104).

The effect of raindrop action on the narrow crests between cultivation rills was clearly visible. Freshly loosened soil was literally gnawed away by raindrops, and so in this case, a large amount of material was displaced into the erosion rills by splash erosion, although the proportion of material moved in this way relative to the amount removed by rill erosion was small. Moreover, the furrows running obliquely along the slope were broken up by the surface water (Fig. 105).

Fig. 105. Detail of a furrow ploughed obliquely on the slope. On the surface of the furrow microformations caused by raindrops and rills tearing into the furrow can be seen (Myslenice near Bratislava, Czechoslovakia). (Photo D. Zachar.)

On an earth road in the same neighbourhood which was not stabilized but which had a much more resistant surface layer than the loosened soil of the vineyard, the effect of splash erosion was less and soil losses were caused mainly by the erosive activity of water flowing in the tracks (Fig. 106).

Consequently, rill erosion made the largest contribution to erosion losses in this case also. Splash erosion was of importance mostly on loosened, freshly cultivated soil in fields with cultivation rills and narrow strips of furrows.

4.2.5.5 The community of Kendice near Prešov

Another example which is useful for considering the effect of precipitation erosion comes from the community of *Kendice* near Prešov (eastern Slovakia, Czechoslovakia) situated in the same climatic region as that referred to in the examples of snowmelt erosion. Again the effects of a spring downpour with hailstones (on 14th April 1964) were observed. Data were collected by Midriak (1965), and here the author presents selected results which complement the picture of rill erosion and consolidate the conclusions already made.

Midriak estimated that downpours of this intensity had a probable periodicity of 50 years and that on this occasion an area of 17–20 km^2 was severely affected, as in the Hiadeľ–Lučatín region. Soil losses were measured on a concave slope, on which agricultural crops were grown. Of the 430 ha of one crop-growing estate, 91 ha (i.e. 21%) were hit by the catastrophy; 59 ha (13.7%) were damaged by erosion and 32 ha (7.4%) were affected by sedimentation. The soil was loamy, of low permeability and had a fine earth content of 98–99%. The rill volumes are given in Table 42.

Table 42. Data on rill erosion on agricultural land in Kendice

Transection	Distance from the divide [m]	Ground inclination	Soil loss [m^3 ha^{-1}]	Damaged area [%]	Crop
I	25	18°	212.4	33.3	Oats
II	85	15°30′	317.7	54.5	Oats
III	135	12°30′	513.9	65.0	Spring barley
IV	185	9°	245.2	78.2	Spring barley
Mean	—	13°45′	322.3	57.7	—

◀ **Fig. 106.** Rills caused by the deepening of tracks on a field road running parallel to the steepest axis of the slope. The narrow strips between the tracks are eroded by raindrops. The type of soil is the same as that shown in Fig. 104. (Photo D. Zachar.)

Table 43. Dimensions and density of rills on eroded land (measured in horizontal transection)

Transection	Distance from the divide [m]	Ground inclination	Mean distance [cm] (density) [km ha^{-1}] of rills	Rill width [cm]			Rill depth [cm]		
				min.	max.	Mean	min.	max.	Mean
I	25	18°	50 (16.87)	5	60	19.7	1	19	6.5
II	85	15°30′	21 (21.79)	8	75	25.0	1	13	5.7
III	135	12°30′	32 (19.40)	8	89	33.5	1	22	7.6
IV	185	9°	29 (25.00)	5	51	19.3	2	11	5.5
Mean	—	13°45′	33 (20.76)	6.5	68.7	24.4	1.2	16.2	6.3

The average soil loss at a slope angle of 14° was about the same as in the previously described cases, i.e. 322 m^3 ha^{-1}, but the rill-damaged area was larger – 58%. It is interesting to note in the context of the present topic that at greater distances down the slope soil losses continued to increase, even beyond a distance of 135 m from the slope ridge, although the steepness of the slope declined from 18 to 12°; at a distance of 185 m and a 9° angle of inclination, soil losses were still larger than on the upper part of the slope (235 against 214 m^3 ha^{-1}). At a distance of 25 m down the slope, and with a steepness of 18°, the degree of erosion reached 212 m^3 ha^{-1}. Only when the water was saturated with silt and the incline was less than 2° did soil accumulation begin to prevail over erosion. Total soil losses in the Kendice community as a result of this downpour were estimated at 25,000 tons.

Data on rill dimensions and rill density are given in Table 43. Compared with the previous examples, the rills were shallower (average depth 6.3 cm) and wider (average width 24 cm), probably as a result of previous humification of the soil by rain which fell on 11th April 1964, and the particular properties of the local soil. Spring cereals (oats and barley) were sown on the slope, and therefore the soil was disturbed by the spring sowing operations, mainly in the surface layers which were not adequately stabilized by the crop later on. Changes in the soil properties of the slope due to erosion were less than those occurring in skeleton soils. The grain diameter d_{40} increased in the upper soil layer from 0.02 to 0.05, i.e. 2.5 times.

4.2.5.6 Loess region in the Hlohovec township

Finally, some of the author's own observations on the effects of heavy rainfall in the *loess soil* region near *Hlohovec* (southwestern Slovakia, Czechoslovakia) are briefly mentioned. In this case as before, the outcome of a violent downpour with hailstones was investigated. This occurred on 3rd June 1958 between 4.30 and 5.45 p.m., and affected an area of about 9 km^2. The soil was impermeable, heavy, and varied in type from a loamy to a clay soil. Damage was caused only on those fields which were largely unprotected by vegetation, and once again, more damage occurred where the soil was freshly cultivated. From the data collected, only the averages for crop groups are listed here:

Crop	Angle of inclination	Soil loss [m^3 ha^{-1}]	Rill area [% of total]
Tended root crops	15°	363	31.2
Older vineyards	15°	123	13.9
Maize (ploughing up and down the slope)	9—10°	309—396	30.9—31.4

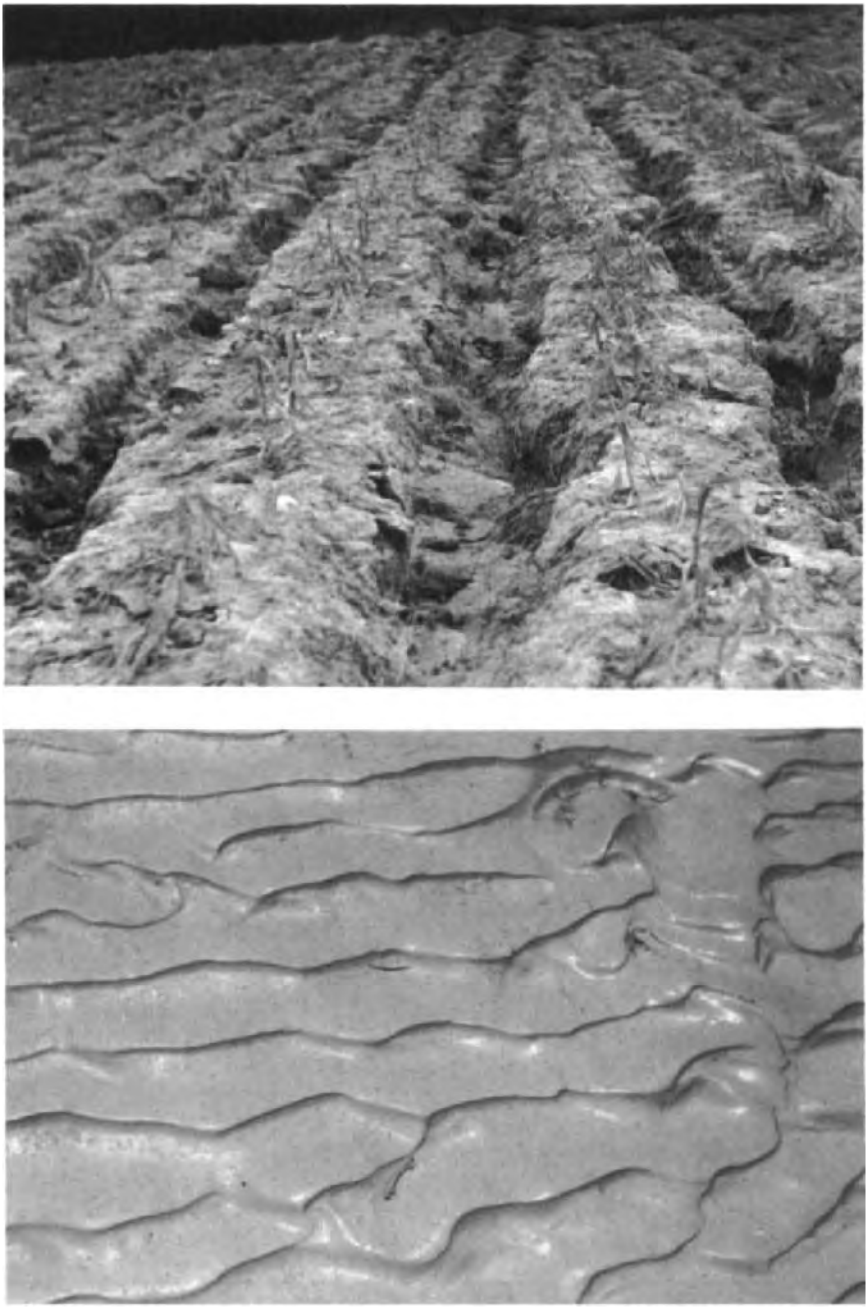


Fig. 107. Cultivation rills deepened by heavy rain on a maize field (a), and finely grained rippled mud deposits of loess loam on the lower part of a field (b) (Hlohovec, Czechoslovakia). (Photo D. Zachar.)

In this case, the soil loss increases about 3 times on account of the loosening effect of cultivation, and about 6 times as a result of the longitudinal cultivation rills occurrence on the slope, these losses being recorded under the same conditions of slope steepness and length. Figure 107 shows that soil losses were again caused mainly by rill erosion, although a considerable part of the soil was transferred into the rills by the splashing of raindrop impact and by wash water. Whereas in all the previously described cases there was a considerable selective effect in the washing of eroded soil and deposits, in this case the selection was weak.

4.2.5.7 Other data from Czechoslovakia

Besides these data from Czechoslovakia, many other examples could be given of the large erosion losses caused by summer downpours.

In this respect it is particularly worth mentioning the results of Demek and Seichterová (1962), who investigated soil losses caused by rill erosion in *central Moravia* (the basin of the Oskava river) after downpours on 22nd May 1960 and 24th June 1961. In 1960 in the locality of Horné Libiny, they found that rills began to develop on tilled land at an angle of inclination of 6° with a concomitant soil loss of 8 m^3 ha^{-1}; at 74 m distance down the slope, the rate of removal increased to 1,400 m^3 ha^{-1}. On another slope, rills began to develop on a 2 to 3° gradient producing an erosion rate of 22 m^3 ha^{-1}, but at distances of between 40 and 110 m from the top of the slope and with the gradient increasing to 9°, soil removal rose to 712.5 m^3 ha^{-1}. In the locality of Mostkov, water originating from a meadow with a 22° slope was found to give rise to erosion losses of 236.2 m^3 ha^{-1} on a lower situated field with a 7° slope and a width of 270 m; in the upper part of this field the topsoil was washed away in strips 1.8 m wide, exemplifying a transition stage between rill erosion and laminar erosion. In the lowest part of the field deposition occurred while the angle of inclination was still 5°.

On 24th June 1961 a downpour lasting for two hours was experienced around the community of Mirotinky. The slope on which the erosion was studied was covered by soil varying from loam to a sandy soil with a higher skeletal content. The above-mentioned investigators established that on a potato field on the lowest part of the slope, the rills had a volume of 1,102 m^3 ha^{-1} where the slope was 15°, and 792 m^3 ha^{-1} where the slope was 10°. This was the result of a convergent runoff.

Similar values pertaining to rill erosion were obtained by Štelcl (1962) and Czudek (1962) in the region of *northern Moravia.* Czudek also observed the occurrence of laminar erosion.

In *southern Moravia,* total soil losses in a loess region were estimated by Mařan (1958), who measured the volume of deposits in the Alluvium and was also successful in estimating the ages of willow-trees. He found that precipitation erosion had carried away 304,755 m^3 of deposits from a 374 ha area in 48 years, i.e. 17.6 m^3 ha^{-1} $year^{-1}$. Mařan estimated that the average loss had been 20 m^3 ha^{-1}, which in the author's opinion is a low figure, since in loess regions sedimentation tends to be small on the lower parts of slopes. Nevertheless, this represents a high level of soil erosion according to the author's classification (Table 3), and in fact the above figures represent for average slopes of 10 to 13% (5°43′ to 7°24′) inclination in Czechoslovakia, where the soil is often damaged by rill erosion.

Relatively small erosion losses were found to occur on sample plots measuring 19.8 × 6.0 m, gradient 24°, in the *České Středohoří Mountains* (Holý 1964). On unprotected plots the losses totalled 25.2 tons, i.e. 16.5 m^3 ha^{-1}, 85.5% of this being attributable to three downpours, the remainder to 23 recorded rainstorms. The figures bear testimony to the considerable resistance of soil originating from createaceous rocks of the Bohemian Massif.

As can be seen from the results obtained on the territory of Czechoslovakia, soil losses caused by rill erosion during heavy rain are greater than those caused by snow thaw water. However, the selective effect of the former is greater only on soils of high skeletal content, and this is mainly a result of splashing and sheet erosion. Losses caused during cloudbursts usually affect smaller areas and occur more sporadically than losses caused by the thawing of snow. Violent downpours have a tendency to deepen and break up the relief as rills are formed and the bottoms of river-beds are cut deeper, whereas long-lasting rain, on the contrary, washes the terrain and soil surface, rounds off any unevenness, and hastens further transport of loosened particles. These processes vary greatly around the globe, and an understanding of the relationships among erosion phenomena, their causative factors and conditions of occurrence in each region is important from the point of view of soil conservation.

4.2.5.8 Other investigations

As far as studies outside Czechoslovakia are concerned, let us consider some results that were obtained in the *USSR,* where great attention has been given to this form of erosion, only few references being available in the English language literature. The observations were mostly made by Sobolev using the volumetric method, and the data (Table 44) are taken from his paper on rill erosion (Sobolev 1948). The data convincingly demonstrate the important role of surface runoff in rill erosion.

Interesting observation were made by Mizerov (1966) of soil erosion in the *Far East* and *Sakhalin Island* caused by snow thaw water and summer and autumn downpours. Rain-water runoff here represents about 72% of the total annual precipitation, and summer an autumn floods occur frequently. Mizerov measured the total soil losses over a period of a number of years on land recently brought under cultivation. This enabled him to obtain data on actual erosion (by the

Table 44. Data on rill erosion according to Sobolev (1948)

Place and date	Crop	Slope inclination	Length of slope [m]	Soil loss [m^3 ha^{-1}]
Orlov region, 25th July 1940 downpour	Fallow	0°	0	0
	Fallow	2°	100	98
	Fallow	4°	285	228
	Fallow	4°	340	82
	Meadow, rye	4°	360	0
Oka basin, snowmelt and spring rain, measured 15th June 1940	Rye	0.5°	0	<0.5
	Rye	1°	200	5
	Rye	2°	350	13
	Rye	3°	400	13
	Rye	3°	500	25
	Rye	2°	650	23
	Meadow	8°	670	0
Bashkir Autonomous SSR, measured Sep. 1942	Rye	2°	100	<0.5
	Rye	4°	200	13
	Rye	8°	300	13
	Rye	13°	320	41
	Rye	12°	340	48
	Rye	9°	390	39
	Rye	5°	468	28
	Rye	3°	528	19

Table 45. Soil losses in the Soviet Far East

Part of slope	Profile of slope	Slope inclination	Loss (−), or deposit (+) [cm]	Agricultural crop	Harvest [cwt ha^{-1}]
Upper	Concave	4°	−3, −4	Oats	30.5
Middle		3°	−14	Oats	27.5
Lower		2°	+7	Oats	58.8
		3°	+1	Oats	41.3
Upper	Moderately undulating	1°	−2	Barley	8.1
		2°	−6, −7	Barley	9.8
Middle		3°	−3, −4	Barley	7.5
		2°	−2	Barley	9.3
Lower		1°	+2, +3	Barley	33.8
		3°	−7, −8	Barley	7.3
Upper	Convex	2°	0	Barley	39.5
		2°	−2	Barley	53.7
		8°	−15	Barley	33.4

Table 46. Rill and sheet erosion in the Tien Shan Mountains. Annual soil losses

Precipitation [mm]		Slope inclin-ation	Agricultural crop	Soil loss [t ha^{-1}]			Depth of soil lost [mm]
Rain	Snow			Rill erosion	Sheet erosion	Total	
			Light chestnut soils				
319	116	23°	Winter cereals	160	130	290	22.3
			Bare fallow	350	230	590	57.0
			Dark chestnut soils				
361	134	30°	Winter cereals	110	105	215	18.2
			Unweeded fallow	120	70	190	14.6
			Bare fallow	250	180	430	43.0

historiocomparative method discussed in Chapter 2). From the comprehensive results obtained, some of the data on total soil losses recorded during the 7 years from 1945 to 1951 are presented in Table 45.

Although soil losses and the pattern of erosion both depend on the method of ploughing, and other factors, soil losses generally increase with increasing both slope length and its steepness. Thus a steep angle of inclination multiplies the danger from erosion on a long slope, but with increasing steepness the soil is threatened on all parts of the slope, even though total losses over the entire slope are not very high. The method of ploughing and the ease with which surface runoff can occur are of decisive importance here. On the upper parts of slopes losses are most in evidence, whereas on the central and lower parts deposition may occur. With respect to total losses, convex slopes ending directly in the hydrographic network are the most susceptible. Accumulated flows of water have a strong erosive effect, and soil loosened by erosion may be carried directly into watercourses. It would be difficult to understand these processes without first acknowledging the powerful erosive force of flowing water. This topic is further discussed in the section on topographic factors.

Finally, Mikhaĭlov (1949) has published data on sheet and rill erosion occurring in the *Tien Shan Mountains* of Central Asia. Although precipitation levels were relatively low, erosion losses were very large (Table 46). According to Mikhaĭlov, erosion is caused mainly by snow melt water in those regions where the subsoil is still frozen at the time of the thaw.

On fields of winter cereals an unexpectedly large proportion of total soil losses is caused by sheet erosion and the proportion of the total erosion accounted for by rill erosion is relatively high, so that much of the soil loss occurs in the spring. The author's findings have shown that the contribution made by sheet erosion is smaller than that made by rill erosion and does not exceed 30% of the total losses. Holý (1964) found that rills only appeared during two rainstorms out of 25. In lighter rain with a 14.5% wash, rills did not develop; the occurrence of rills during very heavy downpours was not established. If the ratio of sheet erosion to rill erosion is the same as that found on the author's plot in the Low Tatras, the ratio of losses caused by sheet and rill erosion, respectively, can be expected to be about 45 : 55.

Unfortunately, the relative soil losses caused by splash, wash and rill erosion have not been investigated in detail, and therefore the above approximation must suffice, provided it is understood that the ratio will change in favour of rill erosion in arid regions, and in favour of wash erosion in humid regions. The greater the occurrence of violent downpours, the greater is the activity of splash erosion, especially on permeable and semipermeable, non-resistant soils. The proportions of the various forms of sheet erosion are strongly influenced by ploughing, soil cultivation, runoff intensity, and soil surface compaction by cattle, etc. An important observation often made is that both the intensity and the form of erosion depend on the time which has elapsed since the start of the activity. This is confirmed by measurements obtained by Burakovskaya (1963); she found that on freshly ploughed fields soil losses were 5 to 8 times larger than on other fields, the form of erosion changing also.

Research activity in this direction needs to be intensified, because the selection of appropriate soil conservation measures depends on the availability of information on the different possible causes of erosion losses and the relative importance of these. The problem seems to be more complicated than was first supposed and requires more methodical approach including a revaluation of information already obtained.

4.2.5.9 Morphometric data on rills and the pattern of rill growth

For the sake of completeness some aspects of the relationship between the volume of losses caused by *rill erosion* and the density and area of rills are discussed.

By interpretation from measurements it seems that the more intense the erosion, the larger is the rill volume. On the upper part of a straight slope the rills are mostly small and they gradually become deeper unless the water is saturated with silt, in which case the rill depth decreases. A state of equilibrium with respect to rill dimensions usually occurs at a certain distance beyond the steepest part of the slope (see Table 47).

The data show that the area damaged by rills may vary greatly, from nil to 60%, and is proportional to the flow of channelled water and the rate of vertical erosion.

Table 47. Data on rills occurring on various plots

Plot, and slope inclination	Year	Erosion factor	Depth of soil lost [mm]	Mean rill depth [cm]	Mean rill width [cm]	Area of rills [%]	Density of rills [km ha^{-1}]
Radvaň, 8°	1958	Snowmelt	1.9	45.2	45.0	4.2	0.9
Závadka, 9°	1958	Snowmelt	6.4	42.1	71.8	15.2	2.1
Kendice, 14°	1964	Downpour	32.2	63.0	244	57.7	22.0
Hiadeľ, 14°	1958	Downpour	110.4	18.4	167	60.1	41.5
Hriňová II, 13°	1966	Rain-water	22.9	206	73	23.3	12.5
Hriňová III, 20°	1966	Rain-water	43.2	244	126	26.8	12.4
Hriňová III, 14°	1966	Rain-water	40.1	126	288	23.8	9.0
Hriňová IV, 12°	1966	Rain-water	36.2	112	331	24.7	8.3

In rill erosion caused by thawing snow vertical erosion may be limited by the depth of unfrozen soil, whereas in rain-caused rill erosion the depth of topsoil or surface cultivation may be limiting. On unploughed homogeneous soil the rill density and depth/width ratio of rills are determined mainly by the permeability and resistance of the substratum. Rill dimensions increase with time and with increasing distance down the slope, while concomitantly the rill density declines, and the juvenile forms turn into gullies.

The erosion losses brought about by precipitation water flowing in rills arise mainly from the erosive effect of flowing water, although during downpours a large amount of soil is transferred into the rills by the splashing process. This amount as a proportion of the total soil erosion may vary greatly, as has been shown. In any case, the soil surface tends to be dissected by the development of small channels, and new microerosion bases arise which both increase the splashing effect, and accelerate the erosive action of flowing water. Lidov et al. (1973) found that in the spring thaw the quantities of flowing water increased together with the velocity of flow, thus enlarging the rills. The following relationship between flow rates in rills and the velocity flow was established by Lidov:

Water flow rate [l s^{-1}]	0.5—0.6	2.0—3.0	7.0—9.0
Water velocity [m s^{-1}]	0.4	0.7	1.0

With increasing water flow in the rills both their depth and their volume grew, and in so doing the volume of soil loosened by erosion increased also. According to Lidov et al. (1973), these values were related (for a gradient of 3°) as follows:

Water flow rate in rills [l s^{-1}]	0.1	0.2—0.5	1—2	3
Rill depth [cm]	1	5—10	20—50	50 (85)
Soil loss per 10 m of rills [m^3]	0.002	0.08	2.4	4.2

With increasing rill depth there is a continuing accumulation of water in the rills, and the velocity of flow, kinetic energy, and carrying force all increase further. Kostyakov (1938) has deduced that the confluence of water in rills, for a given level of runoff, increases soil erosion as much as fourfold compared with sheet runoff.

Based on these theoretical considerations together with information on the pattern of precipitation erosion, a number of equations have been set up for the calculation of the amounts of soil removed by precipitation water. These equations generally focus on one of three considerations; one group of equations is centred on the kinetic energy of the rainstorm and its splashing effect, another group focusses on the erosion and washing effects of surface flow (originating from both rain and snow water), and the third form of equation is based on the erosive effect

Fig. 108. Soil damaged by sheet and rill erosion overlying the soft subsoil of the Rimava basin-shaped valley (Cerová Hills, Czechoslovakia). (Photo P. Plesník.)

of runoff channelled into rills. Each of these equations concerns only one phase of surface erosion, namely splash, wash, or rill erosion; a universal equation taking into account all the diverse erosion processes has as yet to be formulated, though some existing equations are taken to be universal.

In the assessment of soil erodedness, however, the effect of ploughing on soil displacement, referred to as mechanical erosion by some authors (Yatsukhno 1976, and others), has not been properly considered. This is a man-made form of erosion of which the author suggests the term *arable erosion* (Latin *aratio* – ploughing). It can be calculated from the formula

$$E_a = \frac{h \Delta_x \gamma \cdot 10^4}{l} ,$$

where E_a is the aratial erosion [t ha^{-1}], h the depth of ploughing, or other agricultural operation [m], Δ_x the distance of displacement due to ploughing [m], γ the specific weight of soil [t m^{-3}], and l the slope length [m].

As already mentioned, the ET index of this form of erosion is small, and yet the effects on the upper and steeper parts of slopes can be considerable, especially where the ploughing has been done quickly.

The overall effect of all these processes on tilled land is manifested in even sheet form, because any unevenness caused by erosion disappears during soil cultivation, whereas on unploughed land the *rills* deepen and turn into *gullies*. The aggregate effect of these influences on the soil surface is the ultimate exposure of the lower soil horizons or even of the bedrock, where fertility is generally much lower (Fig. 108).

Before discussing linear forms of precipitation erosion, the factors and conditions governing all the forms of surface erosion discussed hitherto are summarized.

4.2.6 Factors and conditions governing surface erosion

In the foregoing sections it was seen that the active factors in *precipitation erosion* were the precipitation itself and the flow of precipitation water down the slope; conditions affecting the erosion were the properties of the soil, the nature of the relief and vegetation, the type of soil management, and erosion control measures.

4.2.6.1 Precipitation, climate and runoff

With respect to precitipation, an important distinction is made between liquid precipitation, especially *rain*, and solid precipitation, especially *snow*. The most important aspects of rainfall are its total quantity, and its intensity; the same can be said of the thawing of snow to produce runoff. The erosive effect of rain is enhanced by the disaggregating and splashing effect of raindrops and hailstones, and likewise, the influence of snow water is increased by the disaggregating effect of frost and the reduced permeability of frozen underlying strata. Besides depending on the disaggregating effect of raindrops and frost, the total amount of eroded soil also depends on the erosive action and transporting capacity of surface flow. Without surface runoff, the amount of soil erosion caused by precipitation is relatively small. Therefore a critical factor that determines the erosive effect of precipitation water is the permeability of the soil, which indirectly influences total soil losses and the pattern of erosion processes on slopes. While the erosive activity of raindrops is determined by the kinetic energy of the raindrops, the erosive action and transporting capacity of surface flow depends on its quantity, velocity and degree of confluence.

While the erosive effect of raindrops depends on the size of the soil grains for a given type of soil and on the velocity of falling raindrops (which is a function of their size), the erosive effect of surface flow depends on the critical velocity of the water and its carrying capacity, which varies according to the soil grains being carried. The erosive action of rain-water increases with increasing size of the raindrops, since larger drops have the effect of reducing soil permeability. Ambrov (1954) established that an increase in raindrop diameter from 0.5 to 1.5–1.8 mm is associated with a 2.1-fold decrease in the permeability of chernozem soil, and that there is a 3.6-fold decrease in permeability when the raindrop diameter reaches 2.4–2.9 mm. This means that as the intensity of precipitation increases, the contribution made to the overall erosion by surface runoff increases faster than that made by the impact of the rain on the soil.

Zaslavskiĭ (1966) reported that during the period 1944–1953, surface runoff on the loess soils of the Tien Shuy erosion control station developed in only 81 rainstorms out of 709 (Table 48), and observed that with increasing rain intensity the proportion of surface runoff erosion grew larger, and consequently erosion losses increased.

Table 48. Total amount of precipitation in a period of 12 years and number of rainstorms with and without the occurrence of surface runoff. Region Tang-shan, period 1945—1956

Amount of precipitation for each rainstorm [mm]	Total number of rainstorms in 12 years	Number of rainstorms with surface runoff	Percentage rainstorms with surface runoff
<5	367	3	0.8
5—10	142	12	8.4
10—20	118	25	21.2
20—30	45	18	40.0
30—40	22	12	54.5
40—50	7	5	71.4
50—100	6	5	83.3
>100	1	1	100
Total	709	81	11.4

Regardless of the relative extents of the various phases of the erosion process, it may be stated that *rain intensity* is the most important factor governing soil erosion caused by rain. Well-founded data on the relationship between rainfall intensity and erosion are given in Table 49.

The general rule is that the more permeable the soil, the smaller is the erosive effect of rain, and vice versa. On comparatively impermeable soils, soil wash occurs provided that the total amount of rainfall is large, even if not of high intensity.

Interesting data on the relationships between rain intensity, runoff, and erosion losses were obtained from the many pluviosimulation experiments carried out by

Table 49. Effect of precipitation intensity and corresponding number of precipitations on erosion losses (according to Fournier 1972)

Precipitation intensity calculated from 5-minute intervals of rain [mm h^{-1}]	Number of precipitations	Erosion loss [t ha^{-1}]
0 — 25.4	40	3.75
25.4— 50.8	61	5.95
50.8— 76.2	40	11.78
76.2—101.6	19	11.44
101.6—127.0	13	34.24
127.0—152.4	4	36.32
152.4—177.8	5	38.72
228.6—254.0	1	47.93

Mařan (1958). In the case of artificial rain, too, it was observed that a 8.3-fold increase in precipitation intensity was associated with a 201-fold increase in runoff, and a 4,214-fold increase in soil erosion losses. The rate of soil removal was found to increase over the duration of the precipitation, this being consistent with the observed reduction in infiltration and soil disaggregation at the same time.

However, Kazakov (1940) found during artificial rain experiments that the greatest turbidity of the water occurred at the beginning of the experiments when most of the already loosened particles were washed from the soil surface. After the turbidity had declined to a constant level, the rate of erosion increased at a steady precipitation intensity because the proportion of water appearing as surface runoff increased together with its erosive effect. After stabilization of the runoff coefficient, the rate of erosion also became stable. As soon as the intensity of erosion increased, a rapid growth of erosion losses was observed.

The same relationships between precipitation, precipitation intensity and erosion are also evident from the data of Zaslavskiĭ (1966) (Table 50).

As the precipitation intensity increased from 0.15 to 0.30 mm min^{-1}, erosion losses were four times greater.

Table 50. Effect of rain intensity on erosion losses at the Suige experimental station in 1956

Date of rain occurrence	Total rainfall [mm]	Duration [min]	Average intensity [mm min^{-1}]	Volume of runoff [m^3 ha^{-1}]	Erosion loss [t ha^{-1}]
3rd July	40.4	805	0.05	6.1	0.2
22nd July	44.7	292	0.15	106.2	34.8
8th Aug.	45.0	150	0.30	231.7	141.8

The erosive effect of rain also increases during a succession of downpours. It has been shown that the first rain builds up the soil moisture content, disaggregates soil clumps by impact or by dissolution, diminishes soil permeability, and to some extent models the nanorelief and the microrelief. Thus erosion losses increase in successive downpours, although the intensity and amount of rainfall may decrease. The frequency of heavy rain is of great importance in the study of erosion losses under particular climatic conditions, such as those of the Mediterranean.

Assessments of the erosive effect of rain may be based either on the kinetic energy of the rain (to which the erosive effect of raindrops is related in a specific way according to each geographical region), or on the so-called *critical rainfall* at which damaging erosion starts on the bare soil of a given relief. According to author's classification, the harmful erosion intensity expands from 150 to 15,000 kg ha^{-1} year^{-1}; average harmful erosion intensity being 750 kg ha^{-1} year^{-1}. It has been established that where the critical precipitation results in losses of this degree, acute erosion is taking place, and consequently the slope gradient and slope length at which the precipitation causes this level of damage may also be referred to as critical.

According to results obtained by Nishikata (1955), Ambokadze (1957), Kuron and Jung (1958), Kuron (1958), and others, a precipitation rate of 0.1 to 0.5 mm min^{-1} producing 5 to 10 mm total rainfall may be considered as harmful; rainstorm of this intensity may bring about erosion on medium-resistant soil. In some regions, of course, these figures vary, as indicate data on surface runoff and erosion losses on various soil types. Using the results of studies on periodic rainstorms of high intensity and total rainfall, the erosive effect of all rainstorms may be computed and applying mathematical formulae, the probable erosion damage of rain could be even predicted. It is supposed that so-called *non-erosive precipitations* represent only a small part of total erosion, and therefore may be omitted from the calculation.

Erosion losses may also be assessed from slope gradient and slope length, the effects of which depend on all relevant factors relating to the soil and the topography.

The interrelationships between *precipitation intensity* and *erosion* has been expressed in different ways by different authors, according to the method applied and the particular soil under consideration.

Neal (1938) using 3.6 × 1.1 m monoliths established that the following relationship is valid for gradients ranging from 0 to 14.4° and precipitation intensities between 0.38 and 1.67 mm min^{-1}.

$$E = S^{0.8} i^{1.2}, \text{ and } E = KS^{0.8} M i^{1.2},$$

where E is the soil wash [t ha^{-1}], S the monolith surface gradient [%], K the coefficient determined by other factors, M the total precipitation [mm], and i the precipitation intensity.

In this form, the relationship is valid only after saturation of the soil has taken place. Under these conditions the depth of soil removed by erosion is proportional to the duration of the precipitation t, and thus the total soil loss is

$$E = 0.4S^{0.8}ti^{2.2},$$

where E is the soil loss (feet per 1/1,000 acre), S the slope gradient [%], t the duration of the precipitation [h], and i the precipitation intensity (in. h^{-1}).

Kostyakov (1938) correctly takes into account only the non-infiltrating precipitation water, and derives the formula as follows

$$E = A(i - k)^{1.5}S^{0.75}L^{1.5},$$

where E is the rate of runoff of eroded particles [kg s^{-1}], i the precipitation intensity [mm min^{-1}], k the rate of infiltration [mm min^{-1}], S the slope gradient [%], and L the slope length [m].

According to Musgrave (1947) erosion, E, can be defined by

$$E = f(AZ_{30}^{1.75}),$$

and according to the previously mentioned formula of Wischmeier and Smith (1965), by

$$E = f(AZ_{30}^{1.85}),$$

where Z_{30} is the highest precipitation intensity occurring in a 30-minute period of precipitation multiplied by 1 hour.

A survey of the literature shows that exceptional importance is attached to all measurements related to precipitation intensity, which is a vital factor in any calculation of erosion losses.

Besides the intensity and duration of precipitation, other climatic factors are also of great importance; these factors may determine (a) the descent angle of raindrops (wind speed), (b) the properties of the soil (frost, temperature variations, wind, rates of evaporation), and (c) the growth of vegetation and the extent of its protective effect (conditions affecting plant growth rate, canopy closure, ground cover, etc.). These conditions may increase or decrease the influence of precipitation to a very considerable degree.

In the thawing of snow, the rate of thawing and production of surface runoff are the chief factors from the point of view of erosion.

4.2.6.2 Soil

The most important soil properties are *soil permeability* and the *resistance of the soil to erosion.* Soil permeability determines the quantity of surface flow. The resistance of the soil to erosion is referred to in the literature as erodibility (see Chapter 3).

Soil with a sufficiently high *permeability* to absorb precipitation of maximum intensity (5 mm min^{-1}) is only seldom affected by sheet erosion, and is therefore damaged only by splash erosion. However, soils of this type show only little resistance to erosion and any confluence of surface water easily carves rills which attain a considerable size during heavy downpours. If the permeability is very high, damage caused by intrasoil erosion occurs.

Low permeability soil and impermeable soils, on the other hand, are more resistant, but then much greater surface runoff develops on soils of this type. Erosion increases if soil permeability is reduced artificially, or if surface layers are loosened as a result of *soil cultivation.* However if infiltration is increased by soil cultivation above a critical level, erosion decreases. Zaslavskiĭ (1966) gives data obtained from pluviosimulation experiments on a light-loam to sandy chernozem soil lying on a 5° gradient; total precipitation was 60 mm and the rate of precipitation was 2 mm min^{-1} (Table 51).

Soil permeability, as well as resistance to erosion may be increased by improving the *soil structure,* especially if the proportion of water stable aggregates is increased.

In additon to other soil properties, soil erodibility can be determined by *earth texture* (i.e. *granulometric composition*), by the *active surface area of particles,* and by the *homogeneity of granulation,* etc. The coarser the fine earth texture, the smaller the active surface area of the fine earth particles, and the more homogeneous the granulation, the smaller is the resistance of the soil to erosion. Since all these properties are altered by the selective action of erosion and by the transport of particles loosened by erosion, soil which has already been transported is less resistant to erosion. This also holds true for slope deluvia and slope foot deluvia which are two to three times less resistant to erosion (Gussak 1937).

Resistance to erosion is diminished both by the *soil skeleton* which constitute the framework of the soil and forms a *protective layer,* and by *colloid activity, hydrophily,* and *soil plasticity.*

The ratio of R_2O_3 to SiO_2 is particularly important with respect to the resistance of aggregates, including their stability in water, and other properties (see Chapter 3). Yet the importance of this indicator tends to have been overestimated in the past.

As a result of the computer processing of 10 indicators for various types of soil and earth, Mirtskhulava (1970) established that resistance against wash water 6 cm deep increases with the content of particles of less than 0.05 mm size; resistance

Table 51. Infiltration runoff and soil wash for various depths of cultivation

Depth of cultivation [cm]	Mean rate of infiltration [mm min^{-1}] 0—10 [min]	0—30 [min]	Total infiltration in 30 min [mm]	Period before start of runoff [min]	Runoff [l]	Soil wash [kg]
0	1.5	1.0	29.1	2.5	30.9	2.87
6	2.0	1.4	42.0	10.3	18.9	1.86
12	2.0	1.6	46.8	10.0	13.2	0.45
25	2.0	1.7	50.1	10.0	10.0	0.19
35	2.0	1.7	50.7	10.2	9.3	0.15

Table 52. Effect of soil moisture level on water runoff and soil removal from bare plots at the Albacher Hof, Erndtebrück and Marburg experimental stations

Date	Condition of soil	Rainfall [mm]	Intensity of rainfall [mm min^{-1}]	Water runoff [%]	Soil removal [kg ha^{-1}]
	Albacher Hof, Experimental station I				
18th July 1954	Medium moist	13.0	0.03	4.5	38.1
10th Aug. 1951	Very wet	11.0	0.02	8.5	330.0
27th Aug. 1955	Dry	34.0	1.13	40.5	15,625.0
31st Aug. 1955	Very wet	13.5	0.40	65.2	8,750.0
	Erndtebrück, Experimental station II				
1st July 1953	Medium moist	15.0	0.15	5.20	69.0
7th June 1953	Very wet	10.6	0.15	20.9	931.2
	Marburg, Experimental station III				
9th June 1955	Medium moist	2.5	0.25	42.4	10.4
2nd Aug. 1955	Very wet	3.0	0.30	48.9	537.0
18th July 1955	Dry	20.0	0.80	84.5	10,180.0

also increases with the content of particles of less than 0.001 mm, with increasing soil plasticity, and with an increasing angle of internal friction. However resistance decreases with increasing porosity and maximum molecular water capacity.

Another important factor is the instantaneous soil moisture content, which plays a vital role during successive rainstorms or during downpours which occur in the dry season. It has generally been established that the higher the soil moisture, the lower is the resistance of the soil to erosion, these properties being associated both with infiltration and with the resistance of soil aggregates. Data obtained by Jung (1956) from runoff plots may serve to illustrate this (Table 52).

Table 53. Rill erosion on soils of different particle sizes (spring 1958)

Locality	Diameter of grain d_{50} [mm]	Rate of inclination [mm min^{-1}]	Mean slope inclination	Soil loss [m^3 ha^{-1}]
Krčava	0.01	0.01	4°34′	35.4
Závadka	0.05	0.52	9°12′	63.7
Pohorelá	0.8	2.09	18°30′	20.0

The importance of the initial soil moisture content diminishes in the case of intense, high rainfall. Also, on very dry soil with a disaggregated surface erosion losses are larger, owing to the fact that under the initially wetted surface an air cushion develops which prevents further penetration of water; this occurs mainly in arid regions. On saline soil after drying, water enters the cracks and causes intrasoil erosion. Very intense erosion may occur even on a gentle gradient on account of the swelling of surface particles and the peptization of aggregates.

It may be interesting to look at three examples obtained from the author's own measurements, in which snow thaw erosion resulted from an almost uniform rate of thawing (Závadka, Pohorelá) (Table 53).

On the Krčava plot, although the snow cover was the least, erosion losses were proportionally the highest. After adjusting for the quantity of snow thaw water, it was found that soil erodibility on the Krčava plot was about 15 times higher than soil erodibility in the Pohorelá region. If a wider range of soil types, including the most extreme types, were to be taken into consideration, the erodibility of soils would be found to vary by up to a hundredfold.

Even without considering the soil type extremes, very much larger variations in soil erodibility than those discussed in the literature may be expected over large areas containing different pedolithic structures. After all, some skeletal soils in mountain areas are hardly eroded on 20 to 25° slopes, whereas loess loam and soil developed on loess loam may be damaged on slopes with gradients of 2 to 3°, and this difference is much larger if a gradient function with an exponent of 1.35 to 1.5 is being considered (see Sec. 4.2.6.2).

Wischmeier et al. (1958) and Smith and Wischmeier (1962) assessed soil erodibility in terms of the depth of removed soil divided by the index of rainfall-mediated erosion. These parameters were based on measurements made on denuded soil on experimental plots of constant dimensions (22 m length, 9% inclination). Soil erodibility assessed in this way is doubtless the nearest to reality, yet it depends on the dimensions of the experimental plots, and to a certain extent is distorted by the amount of surface runoff.

In the course of the author's research, soil erodibility was assessed from the magnitudes of the critical inclination and slope length, at which acute erosion commenced on denuded soil, or at which erosion of a certain intensity occurred under standard precipitation conditions (according to the method used).

4.2.6.3 Relief

The *relief* of the terrain is of fundamental importance in determining levels of precipitation erosion. Included among the factor of the relief are: slope inclination and length, slope form, modelling of the relief, slope aspect, and – affecting erosion indirectly – elevation above sea level.

Slope inclination

As the slope becomes steeper, the runoff coefficient increases, the kinetic energy and carrying capacity of surface flow become greater, soil stability and slope stability decrease, splashing erosion increases, and the possibility of soil displacement in a downhill direction during ploughing is greater. Thus the likelihood of soil erosion increases with the growing steepness of the slope.

This may also be seen from the basic equation for the acceleration of a falling body, $v = \sqrt{2gh}$, where v is the final velocity, g the acceleration due to gravity (9.8 m s^{-1}), and h the height of fall. Accordingly, if the height of fall from top to bottom of the slope is increased fourfold, the velocity of flowing water, regardless of friction, increases twofold. If the water velocity doubles, the kinetic energy (being proportional to the square of the velocity) increases fourfold, but the volume of particles carried away (given by the relationship $Q = Av^6$) increases 64 times. Likewise, if the mass flow of water doubles, the kinetic energy (KE = $mv^2/2$) doubles, too.

It follows from these very general considerations that with a doubling of the steepness of the slope, the kinetic energy of runoff increases in proportion, but the carrying capacity becomes 32 times greater, which means that with increasing slope inclination transport capacity outstrips erosion. Therefore, any influence which causes soil particles or aggregates to be released augments erosion losses. From this point of view the disaggregation effect of raindrops, hailstones and frost, the operations involving the shallow cultivation of the soil, the mechanical destruction of the soil surface by cattle, rapid temperature changes, etc. are of great importance. The impact effect of raindrops is the most important of all.

Another general conclusion is that any agricultural operation that diminishes the permeability and roughness of the soil surface results in an increase in the volume of surface flow, its velocity, and consequently also its erosive activity. Soil permeability diminishes during heavy rain as a result of the blocking of pores, the creation of an air cushion, and the saturation of the soil surface layer, all of which leads to a greater degree of runoff during rainstorm. Also the subsoil is generally much less permeable than the topsoil, and therefore the shallower the topsoil, the sooner it becomes saturated with water, and the sooner surface runoff develops with the ensuing erosion during heavy, successive rainstorms. With increasing inclination of the slope, the proportion of surface runoff increases, and coarseness of the soil diminishes in importance, for a given range of soil properties.

A rapid increase in the velocity and quantity of water occurs if small channels develop with reduced friction of water flow, the water thus acquiring increased kinetic energy and a still greater carrying capacity. Under these circumstances soil particles and aggregates of substantially larger size may be carried away from the rills. It may be generally stated that during a rainstorm the coarseness of the soil surface diminishes and small channels develop, in which the erosive activity and transport capacity of surface flow rapidly increase up to certain limits. As water approaches saturation with silt, its erosive activity declines, and a temporary sedimentation of particles may often be observed when the gradient decreases. The pattern set by the processes of erosion, transport, and sedimentation depends on the pattern of rainfall and on soil conditions, the more important of the latter being permeability, resistance to erosion, and the coarseness of the surface. The steeper the ground surface, the shorter is the distance over which runoff gathers in volume to the extent that erosion of the soil can begin. This, again, holds true within certain limits. However, the overall effect largely depends on the intensity of precipitation, or the intensity of snow thaw, both of which have a stronger influence on erosion and the carrying capacity of surface flow than the direct effect of raindrops and their kinetic energy. Thus after a certain slope inclination and slope length are exceeded, total erosion losses depend mostly on the erosive activity of surface flow, although raindrop erosion may also be important, its contribution to the total erosion is varying according to circumstances.

Some now well-known results were published by Bennett (1939) in his first monograph (Table 54). The precise effect of ground inclination was found to depend on the properties of the soil, nevertheless when the gradient increased twofold, soil losses increased threefold, and since the proportion of water going into the surface runoff increased only by a small amount, this indicated an increase in the erosive activity of the runoff. From another paper by Bennett (1955), it may be seen that on less permeable, yet easily erodible soil supporting a maize crop, the soil removal increased from 158.8 to 243.7 t ha^{-1} when the ground inclination increased from 8 to 20%, whereas on more permeable soil supporting a cotton crop, the soil loss increased from 50.1 to 136.8 t ha^{-1} when the slope inclination increased from 8.7 to 16.5%.

According to field measurements recorded by Gadzhiev (1962), erosion losses increased most within a range of slope angles from 10 to 25° (Table 55).

From the author's results from comparable plots on the Hriňová dam, the following relations were established:

1. slope inclination 13°30′, soil loss 229 m^3 ha^{-1};
2. slope inclination 20°15′, soil loss 432 m^3 ha^{-1}.

For 1.5-fold increase in slope inclination, erosion losses increased 1.9 times.

More precise data have been obtained from specific studies of the relationship

Table 54. Effect of slope gradient on annual soil and water losses under clean tillage *

Soil and location	Period, inclusive	Rainfall [inches]	Length of slope [feet]	Slope gradient [%]	Crop	Soil loss [t acre^{-1}]	Water loss [% precipitation]
Muskingum silt loam, Ohio	1934—1936	36.46	72.6	8	Corn	60.0	30.0
				12		73.2	42.0
Houston black clay, Texas	1933—1936	34.90	72.6	2	Corn	10.6	13.4
	1933—1936	35.47		4		30.4	16.6
Kirvin fine sandy loam, Texas	1931—1936	40.82	72.6	8.7	Cotton	27.9	20.9
	1933—1936	43.00		16.5		72.0	14.6
Shelby loams, Missouri	1918—1931	40.37	90.75	3.7	Corn	19.7	29.4
	1931—1935	34.79	72.6	8		68.8	28.3

* Measurement at soil water conservation experiment stations, Soil Conservation Service.

Table 55. Soil losses caused by erosion on slopes of different steepness in Azerbaijan

Slope inclination					
	Range	10—15°	16—25°	26—30°	31—40°
	Mean	12.5°	20°	27.5°	35°
Soil loss [t ha^{-1}]		61.4	148.5	195	240.5
Increase in losses with increase in slope [t ha^{-1}]		61.4	87.9	46.5	45.5

between erosion losses and slope inclination. Thus, Gussak (1937) working with 1.0 × 0.4 × 0.5 m monoliths with surface inclinations of 5, 10, 20 and 30°, established the following relation between erosion *(E)* and slope inclination *(S)* for resistant soils

$$E = f(S^{0.4}).$$

Neal (1938), also using monoliths, arrived at a relationship $E = f(S^{0.8})$. The biggest increase in the rate of removal was recorded when the inclination of the soil monolith was increased from 3.6 to 7.2° (1.86 times), while the smallest increase occurred over a change of inclination from 1.8 to 3.6° (1.42 times). The relationships between slope inclination and soil erosion arising from Neal's experiments are shown in Fig. 109.

Zingg (1940) evaluated data from 8 experimental plots 2.43 m long, 1.22 m wide, and with gradients of 4.8 and 12%. The above relationship was found to be characterized by $E = f(S^{1.4})$, the rate of removal increasing 2.61-fold when the gradient was doubled, although runoff increased only 14.5% at the same time.

Similar analyses were also undertaken by Musgrave (1947), who evaluated the results from 17 experimental stations in the USA. Plot dimensions were 6 × 2.76 feet, and on each station between 150 and 300 precipitation periods were studied. The relationship between E and S was found to be similar to that arrived at by Zingg: $E = f(S^{1.35})$.

Smith and Wischmeier (1962) evaluated the results obtained from many observations on experimental stations, and, taking into consideration the relationships derived by Hays, Zingg and other authors, they derived the following equation

$$A = 0.43 + 0.30S + 0.043S^2,$$

where A is the soil loss (m^3 ha^{-1}), and S the slope inclination [%].

Hudson and Jackson (1959) established that under extreme tropical conditions the exponent of slope inclination is approximately 2 ($E = f(I)^2$).

Results obtained by pluviosimulation methods are similar to those obtained by deluometric methods on experimental plots with natural precipitation, and are apparently valid also for thawed snow water.

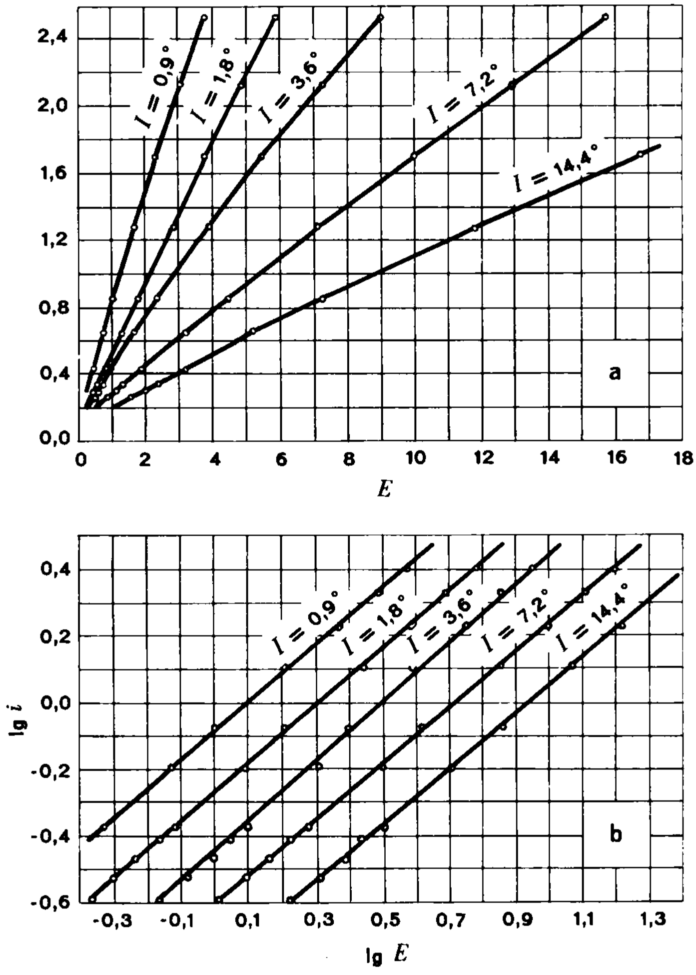


Fig. 109. Relationships between precipitation intensity (i, mm min^{-1}), the upper surface gradient of the monolith (I, degrees of inclination), and soil wash (E, t ha^{-1}) (according to Neal, 1938).

An important indicator of probable erosion is the so-called critical slope inclination at which harmful erosion occurs on unprotected soil (Table 3). In seasonal erosion phenomena caused by very heavy precipitation (e.g., rate of rainfall averaging 0.5 mm min^{-1}, duration 45 to 60 min), or rapid snow thaw, the inclination of the ground at which the development of rills starts is considered to be critical. In linear erosion this is equivalent to the inclination at which gullies start to develop. Critical inclinations established in the course of the author's research are given in Table 56.

Table 56. Lower limits of the critical inclination for precipitation erosion obtained from various plots

Locality	Critical inclination (degrees) for:			
	sheet erosion	slope rills	road erosion	valley rills
Popričný (loess loam)	1	2.5	1	1
Šarišská vrchovina, Čerhovské pohorie Mts, Neogene flysch (loess loam)	2	3	1.5	1
Kremnické pohorie Mts (carbonate rocks)	3—4	7	3	1
Slovenský kras Mts (foot of slopes)	2—5	5	—	1.5
Low Tatras (crystallinicum)	5—8	8—12	5	3
Brezová, Tematín Hills (mostly dolomitic rocks)	8—10	17	—	6

Although the values given may not accurately express the prevailing conditions over entire geographical regions, they nevertheless show that the *critical inclination* varies within a considerable range. The extremes of this variation are from 1 to 8° for sheet erosion, from 2.5 to 16° for slope rills, and from 1 to 6° for rills at the foot of the slope.

In Germany, Kuron (1941) considers an inclination of 4.5° to be critical. The following gradients are also considered to represent a threat from erosion: 2° for loess, 5° for sandy soil, 7° for more resistant soil, and 2° or even less for valleys. Schultze working in Thuringia, established that the critical inclination varies between 1 and 7° for fields, between 5 and 10° for roads, and between 20 and 30° for meadows and forests. According to Schultze, there is a high probability of erosion on fields when the critical inclination is exceeded by about 5°. In a survey of dangerous erosion levels, he lists the critical inclinations for 14 soil substratum types, including the limits of tolerable steepness for fields, which lie within the range 5 to 10°. It seems that this range includes within it the limit of slope inclinations above which the absence of a protective vegetation cover is dangerous for soil stability under central European conditions; thus on slopes of this steepness on tilled land, erosion control measures are called for.

If other indicators of erosion are used, the angle of inclination at which eroded soil begins to appear may then be considered as critical. It can be assumed that the latter occurs when soil losses outweigh soil formation, in which case the soil must be protected against further losses. No great harm is done if the average annual losses on bare land range within 0.5 to 5 m^3 ha^{-1} – according to land use. Expected losses on steeper slopes may be computed using the relationships mentioned above, taking into account, of course, the slope length, and other aspects of the relief.

If for various critical slope inclinations the same magnitude of soil loss is assumed (e.g. 5 m^3 ha^{-1} for acute erosion), and if the gradients in the investigated territory are assumed to range from 4 to 28%, then

$$A = 0.43 + 0.30S + 0.043S^2.$$

Values for K factor (soil factor) are obtained, as shown in Table 57. The equation for the calculation of losses from slope inclination may be verified using empirical data obtained from the region in question.

These data are in close agreement with the author's measurements of soil removal, and determinations of critical slope inclination (Sec. 4.2.6.2). The method may be used to advantage for making estimations of the magnitude of soil erosion losses, provided that the critical angle of inclination and the average losses due to

Table 57. Example of data for the calculation of the K factor according to critical slope inclination, and of soil losses according to the equations of Wischmeier et al. (1958)

Category of soil	Slope gradient [%]		Soil factor	Soil loss	Relative erodibility against category
	Range	Mean			
I	3—5	4	K_1	2.32	5.04
II	6—10	8	K_2	5.58	2.10
III	11—15	13	K_3	11.70	1.00
IV	16—20	18	K_4	19.76	0.59
V	21—25	23	K_5	30.08	0.39
VI	26—30	28	K_6	42.54	0.28

Table 58. Example of data for the calculation of soil losses in different soil categories by the critical slope inclination for a loss of 5 m^3 ha^{-1}

Slope inclination [%]	Soil factor					
	K_1	K_2	K_3	K_4	K_5	K_6
4	5.0	2.1	1.0	0.6	0.4	0.3
8	12.0	5.0	2.4	1.4	1.0	0.7
13	25.0	10.5	5.0	3.0	2.0	1.4
18	42.6	17.7	8.5	5.0	3.3	2.3
23	64.9	27.0	12.9	7.6	5.0	3.5
28	91.7	38.4	18.5	10.7	7.0	5.0

harmful or permissible erosion are known, the relationships between soil loss and slope inclination being valid, albeit approximate, for both rain erosion and snow thaw erosion, regardless of the cause of the soil losses. More information would, of course, be desirable with respect to these relationships.

The calculation of the magnitude of erosion for various soil categories with different resistances is given in Table 58.

Slope length

Slope length is important mainly with respect to the increase in the flow of water on slopes and the degree of confluence. As the quantity of water and its degree of confluence grow, the velocity and transporting capacity change. Unfortunately, little attention has yet been given to these matters in erosion research, mainly because the mathematical relationships between erosion losses and slope length have been based on data from relatively short runoff plots.

Some well-known results obtained by Bennett (1939) show that soil losses do not always increase in proportion to the slope length. Situations have been known, in which erosion losses decreased with increasing length of the slope. This may have been due to the use of inadequate methods which failes to account for the erosive transporting effect; it was found at the same time, that with increasing distance down the slope the proportion of surface runoff had declined, or changed little. Data selected from the publications by Bennett (1939) as well as from other sources are given in Table 59.

The data show that erosion losses tend to increase with distance down the slope. The largest increments were: over the first 20 m on the Krčava plot (6.6 times), over the first 20 m on the Radvaň plot (4.5 times), over the first 10 m on the Hriňová plot (2.1 times), over the first 100 m on the Rokytovce plot (4.0 times), and over the first 32 m on the San Chuan plot (7.2 times). On other plots erosion losses grew up to distances of 11 and 14 m, respectively. In general, with the growing length of the slope the multiple of erosion intensity decreases, although the absolute differences have an increasing tendency.

The empirico-mathematical relationships based on data obtained from experimental plots under different conditions are diverse. Kornev (1937) established that the total discharge of washed material is given by the relation: $E_1 = f(L^{1.5})$ kg s^{-1}, and erosion losses by $E = f(L^{0.5})$ t ha^{-1}. Zingg (1940) derived the following relationships, (a) for total soil loss: $E_1 = f(L^{1.6})$, (b) soil loss per surface area unit: $E = f(L^{0.6})$. Musgrave (1947) arrived at a slope length exponent of 0.5 for the same relationship.

Wischmeier and others used statistical data from 532 plots which included measurements of two or more slope lengths. Data were collected during natural rainstorms on 15 experimental plots. The resulting mean value for the exponent of slope length varied between 0 and 0.9, depending on the effect of slope length on

Table 59. Effect of slope length on soil removal

Place of observation	Author	Slope length [m]	Slope inclination	Soil removal [t ha^{-1}]	Remarks
Czechoslovakia, Krčava near Sobrance	Zachar (1970)	10	2°16′	1.2	Loess loam, spring 1958, tilled land
		20	2°55′	8.0	
		40	4°34′	28.4	
		80	4°36′	45.9	
		180	5°30′	87.6	
Czechoslovakia, Radvaň near Banská Bystrica	Zachar (1970)	10	11°48′	1.7	Loamy soil, spring 1958, tilled land
		20	11°47′	6.0	
		40	8°53′	18.3	
		80	6°13′	24.5	
Czechoslovakia, Hriňová dam	Zachar (1970)	5	20°	130.4	Dare loam-sand soil on the dam bank, 1966
		10	20°	278.3	
		20	21°	448.0	
		40	20°	630.2	
Czechoslovakia, Rakytovce, eastern Slovakia	Kozlík (1958)	25	5°	7.5	Removal caused by downpour on potato field, 1954
		50	5°	5.0	
		100	7°	20.0	
		200	8°	78.0	
		350	8°	148.0	
Cuba, San Chuan	Grazaia et al. (1936)	8	4°	1.2	Loam-sand soil, fixed place of measurement, tobacco
		16	4°	2.3	
		32	4°	16.6	
People's Rep. of China, Sujde experimental station	Zaslavskiĭ (1966)	14	26°	152.5	Fixed place of observation, 1956, *Setaria viridis* L. crop
		20	26°	227.8	
USA, Oklahoma, Vernon fine sandy loam	Bennett (1939)	11.0	7.7 %	42.5	Fine sandy loam, fixed place of observation, 1931—1936, cotton
		22.1	7.7 %	55.6	
		44.3	7.7 %	95.3	
USA, Wisconsin, Clinton silt loam	Bennett (1939)	11.0	16 %	159.0	Sand-loam soil, fixed place of observation, 1933—1936, maize
		22.1	16 %	248.0	
		44.3	16 %	286.0	

the erosive action of the runoff. This provided confirmation of the author's theory of the erosive transport effect of precipitation water. With respect to those places in which the runoff decreased with distance down the slope, the value of the exponent approached zero, but where it increased significantly, the exponent approached

0.9. On plots where the runoff neither increased nor decreased, the exponent varied between 0.27 and 0.6. For practical purposes a value of 0.5 ± 0.1 is recommended for the exponent of slope distance, since this fully concurs with the conclusions of previous research, and with data obtained by the author in the field. It is also consistent with the author's concept of the erosive transport effect of precipitation water.

In the author's research programme, repeated attempts were made to establish the critical slope length at which acute erosion begins. The critical slope length is related to the critical inclination, and acute, harmful erosion occurs as the result of a certain combination of inclination and slope length. The lower the critical inclination, the larger (in most cases) is the critical length. On very steep slopes the critical slope length approaches zero, which means that the soil in such a situation cannot be left unprotected. Whereas the critical inclination is relatively easily established from the inclination at which sediments begin to form, the critical length can be measured only where erosion is taking place, or it may be computed from the critical water velocity and the critical transport capacity of the water.

The derivation of critical slope length from the characteristics of the surface runoff is based on Chézy's equation

$$v_x = cRI \quad [\mathrm{m\ s^{-1}}] \,,$$

where v_x is the velocity of the water at distance x from the divide, I the slope gradient, c the velocity coefficient (which is a function of surface roughness), R the hydraulic radius; $R = F/O = ly/(1 + 2y)$, F the area of cross-section of flow, O the wetted circumference of cross-section, and y the depth of runoff.

Neglecting the small term $2y$ in the denominator, $R \doteq y$, and thus

$$v_x = c\sqrt{yI} \quad [\mathrm{m\ s^{-1}}] \,.$$

According to Bazin, the velocity coefficient may be computed from the relation

$$c = \frac{87\sqrt{y}}{\gamma + \sqrt{y}} \doteq \frac{87}{\gamma} \sqrt{y} = m\sqrt{y}.$$

The value of m $(m = 87/\gamma)$ is determined from Table 60, or from other data on the roughness of the soil surface (Cablík and Jůva 1963). For a ploughed field, the least favourable values are: $\gamma = 3.5$, $m = 24.8$, $O = 1$, except in the case of ploughing up and down the slope.

Cablík and Jůva (1963) have established that the critical slope length L $(L = x,$

Table 60. Values of the coefficient of surface runoff according to Cherkasov (1948)

Condition of soil surface	Coefficient γ	$m = \frac{87}{\gamma}$
Field ploughed up and down slope	2.0	43.50
Ploughed field with levelled surface	3.5	24.85
Field with reed growth	4.0	21.75
Field with moss growth	5.6—6.0	17.40—14.50
Meadow with low grass growth	6.0—8.0	14.50—10.88
Rough soil with many molehills	8.0—15.0	10.88— 5.80

the *safe zone*) can be calculated from the *extreme water velocity*, v_k, according to the relation

$$L_1 = \frac{v_k^2}{moi} \frac{1}{\sqrt{I}} \qquad [\mathrm{m}] ,$$

where L_1 is the critical slope length [m], o the runoff coefficient $(i - k = oi)$, i the rainfall intensity [mm min^{-1}], k the infiltration rate [mm min^{-1}], and v_k the extreme water velocity [m s^{-1}].

According to Velikanov (1948), $v_k = 3.14 \sqrt{15d + 0.006}$ m s^{-1}, where d is the mean grain size. Thus for very small particles $v_k = 0.24$ m s^{-1} (d is the diameter of grain).

In terms of the carrying capacity, U_k, in kg m^{-2}, the following critical slope length (L_2) (or safe zone of the field) was derived

$$L_2 = \frac{mU_k^2}{10^{-6}oi} \frac{1}{I^{3/2}} \qquad [\mathrm{m}] ,$$

where L_2 is the critical slope length, U_k the maximum carrying capacity (according to Strele (1950); $U_k = 1.1$ kg m^{-2} for loamy particles, 0.3 kg m^{-2} for sand, and 2 kg m^{-2} for sod).

For the conditions $v_k = 0.24$ m s^{-1}, $U_k = 1.1$ kg m^{-2}, $\gamma = 3.5$, $o = 1$, $i = 0.58$ mm min^{-1}, duration of rain $t = 45$ to 60 min, or mean rainfall intensity $i = 97 \times 10^{-7}$ m s^{-1}, the L_2 equation gives lower values than the L_1 equation.

Both equations have the general form

$$L = \frac{m^z}{oiI^n} \mathrm{f}(v) \qquad [\mathrm{m}] ,$$

where L is the critical slope length, m and o are expressions of soil properties as above, i is the rainfall intensity, I the slope inclination, $v = V_K^2$, or $v = U_K^2\, 10^6$, and z and n are exponents which must be determined by field experiments.

As soon as confluence begins, the critical slope length decreases according to the value of the rill coefficient $\gamma > 1$(varying between 1 and 2), so that the L_l value changes according to the relation

$$L_l = \frac{1}{\gamma^2}\frac{v_k^2}{moi}\frac{1}{\sqrt{I}} .$$

Slope conformation

The combined variations in length and inclination of the slope determine its *conformation* which is an important factor governing the erosion pattern and erosion losses. The profile may be straight, convex, concave, concavo-convex, or undulating.

On a straight slope, soil erosion depends mainly on the gradient and the length according to the relationships mentioned earlier. The shorter the slope, the smaller is the total quantity of eroded soil, but the eroded soil as a proportion of the total quantity of soil is relatively larger, and vice versa. On long slopes heavy erosion mainly occurs during violent downpours, but the total losses depend on the relationship between the duration of runoff and the overall displacement of the eroded soil.

If, for example, the duration of surface runoff is 5 minutes, the velocity of the runoff is 0.5 m s^{-1}, the distance of water and silt movement down the slope is 150 m (300 s × 0.5 m s^{-1}), and the average quantity of displaced soil is 100 m^3 ha^{-1}, then the soil loss is 100 m^3 ha^{-1} on a slope 150 m long, 50 m^3 ha^{-1} on a slope 300 m long, and 10 m^3 ha^{-1} on a slope 1,500 m long. On very short slopes affected by heavy rain of short duration, the losses are nearly equal to the amount of soil that is loosened by raindrops and surface water. During downpours of long duration and during the thawing of lying snow, the losses (per unit volume of eroded soil) increase more rapidly with increasing slope length than during downpours of short duration.

Consequently, the more broken the relief and the shorter the slope, the larger are the relative soil losses and the transport of silt into the rivers. Under such circumstances the lower parts of slopes tend to become convex, thus causing the water which converges on that section to be more erosive and to transport almost all the eroded soil into watercourses.

The converse occurs on *concave* slopes. The soil removed from the upper parts of the slope is deposited at its foot and only a small portion of the eroded soil is carried into watercourses, although on both types of slope the effect of erosion on the soil may be the same, producing wasteland as a result. The more permeable the soil, the smaller is the displacement effect and the more prevalent is the intrasoil washing. Average soil losses on concave slopes are always smaller than those on convex slopes for a given height and distance between top and bottom.

The smallest soil losses occur on undulating terrain where erosion and deposition processes take place, and where erosion usually has a levelling effect. Erosion takes a similar course on terraced slopes.

In computing probable erosion on complex slopes, the profile of the slope may be accounted for by limiting calculation to the area which lies between the point of critical slope inclination and the upper border of the slope foot deluvium.

In calculating the intensity of soil erosion and the transport of erosion products into the hydrographic network, slope conformation can be accounted for by using the following correction factors:

Slope profile S_f	Correction factors
Concave slope	0.75
Straight slope	1.00
Convex slope	1.25

In very broken terrain with convex slopes and a high density of rills, the correction factor may be as high as 1.5.

Slope aspect

The effect of slope aspect operates through the different degrees of insolation occurring on sunny versus shaded slopes. With the higher temperatures attained on sunny slopes, the rate of decomposition of organic matter, the rate of evapotranspiration, the thawing rate of snow, the degree of soil salt concentration (particularly in arid regions), and other processes all increase. The aspect affects erosion processes mainly on desiccated soil.

Detailed measurements have shown that conditions on slopes directly attributable to slope aspect have a considerable effect on the intensity of soil erosion and land spoilage.

For example, Schubert (1928) made measurements over a period of 17 years (1907–1923) in Potsdam, and found that the total daily solar radiation in June was 30% lower on a northern slope when compared with a southern slope. Krauss (1911) discovered that on lime bedrock in the surroundings of Karlstadt (50° northern latitude), the average soil temperature at a depth of 2 cm was higher on a southern slope than on a northern slope, the differences being 5.9°C in the spring, 8.1°C in the summer, and 3.2°C in the autumn.

The author has made records of soil temperature from 1956 to 1958 in the crystalline region of the Low Tatras where the mean annual temperature is 5.6°C and the average annual rainfall is 815 mm. It was found that conditions on a southern slope differed from those on a northern slope as follows: the soil temperature was higher during the summer months by about 6°C, the average soil moisture content at 100 cm depth was about half of that on a northern slope (19.57% against 40.58%), and the rate of water infiltration into the soil was about four times slower than on a northern slope.

Similar differences were also found on carbonate bedrock in the Kremnické pohorie (mountains in central Slovakia) where the mean annual temperature is 8°C, and the average annual rainfall is 853 mm. The overall average soil surface temperature on southern slopes was 4.5 to 5°C higher, and the average maximum soil surface temperature 11 to 12°C higher than on northern slopes. A forest cover diminished soil temperature by 22°C. Average soil moisture contents were 11.23 and 22.99%, and average rates of infiltration were 0.19 and 0.41 mm min^{-1} on a southern slope and northern slope, respectively.

Still larger differences were found in soils over dolomitic and limestone bedrock, especially with regard to the warming up of the slopes (differences of 8 to 10°C), and soil moisture content (2- to 3-fold differences). These differences in the hydrothermic régime coupled with the effects of slope aspect on the vegetation cover and the resistance of the soil to erosion account for the fact that most wasteland in the temperate zone occurs on sunny slopes.

Detailed research carried out by Midriak (1965) in the Tematín Hills (the dolomitic limestone area of the SSR) has shown that of the total length of erosion rills occurring on an area of 163 ha of wasteland, 82% occurred on sunny aspects and 18% developed on shady slopes. Similarly, in the flysch region of the Ondava Hills (SSR), 82.2% of a total rill length of 71,565 m were found on sunlit slopes, the remainder occurring on more shaded inclines. The most eroded areas tend to be on southwest facing slopes, while northern slopes are the least affected. Differences between the latter range from 10- to 24-fold.

The effects of slope aspect on sheet erosion are not so pronounced, Zemlyanitskiĭ (1937) observed that soil wash on south facing slopes in Central Asia were 2.5 times greater than on northern slopes. Similar differences were found in the Tula region of the USSR by Sil'vestrov (1949). Lidov and Setunskaya (1959) found that in the hill country along the Volga, the stream erosion on south facing slopes was 1.5 to 2.0 times greater than on north facing slopes. Smaller differences were found to occur on arable land where tilling altered some of the properties of the surface layers of the soil, and the protective effects of vegetation were reduced.

Similar differences in the intensity of soil erosion on slopes of different aspect were observed by Ibragimov (1972); it was found that on southern slopes in the Dashkezan district of the Azerbaijan SSR, the rate of soil wash was almost 3 times higher than that on northern slopes. Thus on slopes with a gradient of 10 to 12°, annual soil removal was 73 m^3 ha^{-1} on north facing slopes, 113 m^3 ha^{-1} on east facing slopes, 135 m^3 ha^{-1} on west facing slopes, and 205 m^3 ha^{-1} on southern slopes.

Table 61. Coefficient of erosion danger for the Lithuanian SSR

Angle of slope inclination	Coefficient of inclination	Coefficient of the mechanical structure of the soil				Coefficient of slope aspect*	
		Sandy	Sandy loam	Loam and clay	Humus content in % <1.3	South	East and West
1°	1.0	0.1	0.0	0.1	0.1	0.0	0.0
3°	1.1	0.1	0.0	0.1	0.1	0.0	0.0
5°	1.3	0.15	0.0	0.2	0.1	0.1	0.0
6°	1.4	0.15	0.05	0.2	0.1	0.15	0.05
8°	1.5	0.25	0.1	0.25	0.15	0.2	0.1
10°	1.7	0.3	0.1	0.3	0.2	0.25	0.15
12°	1.9	0.35	0.15	0.3	0.2	0.3	0.15
14°	2.1	0.4	0.15	0.3	0.2	0.3	0.15
15°	2.2	0.4	0.15	0.3	0.25	0.35	0.2
20°	2.7	0.6	0.25	0.4	0.3	0.45	0.3

* There was no tilled land on northern slopes.

A more accurate assessment of the significance of slope aspects in soil erosion was carried out by Rachinskaz (1972) on the basis of experimental research undertaken in the Lithuanian SSR. In spite of the cold climate and the low potential energy the relief, precipitation erosion damages 7% of the territory, mostly the arable land. According to Rachinskaz, as slope inclination increases, the importance of soil properties in determining erosion levels decreases, while the importance of slope aspect increases (Table 61). The level of erosion intensity can be expressed in terms of the sum of coefficients. Thus, for example, for a gradient of 8° with a heavy soil and humus content being less than 1.3% and with an eastern aspect, the level of erosion intensity can be computed as: $1.5 + 0.25 + 0.15 + 0.1 = 2.0$. Rachinskaz divided the territory into areas belonging to different categories, and appropriate erosion control measures and methods of soil utilization were applied to each category.

The effect of the relief, in particular the effect of slope aspect on extensively cultivated land in central Slovakia is shown in Fig. 110. Northern slopes are eroded substantially less than southern slopes. Gully erosion was caused mainly by incorrectly situated and badly maintained field tracks. The effect of the relief is pronounced where the bedrock consists of dolomite. Figure 111 shows an area in which the intensity of sheet erosion depends mainly on the gradient, the aspect, and the influx of water from adjacent land. As far as bedrock material is concerned, the slope aspect is of greatest importance on limestone. Figure 112 shows the lower parts of the slopes of the Silica plateau – a plateau which is almost entirely laid bare as a consequence of soil erosion. The northern slope of this plateau is covered by forests and erosion in discernible forms does not occur there.

Fig. 112. Lower parts of slopes of the Silica plateau (Czechoslovakia) showing the transition to alluvia. In the central part of the picture wasteland originally used for the raising of agricultural crops can be seen. (Aerial photo.)

◀ **Fig. 110.** Territory damaged by erosion in the neighbourhood of Banská Bystrica (Czechoslovakia). On the right of the picture there are northern slopes with well-protected soil, and on the left severely eroded soils can be seen. There is also pronounced road erosion. (Aerial photo.)

Fig. 111. Southwest and southeast facing slopes of severely eroded soil on dolomitic bedrock in the Tematín Hills (Czechoslovakia). (Photo D. Zachar.)

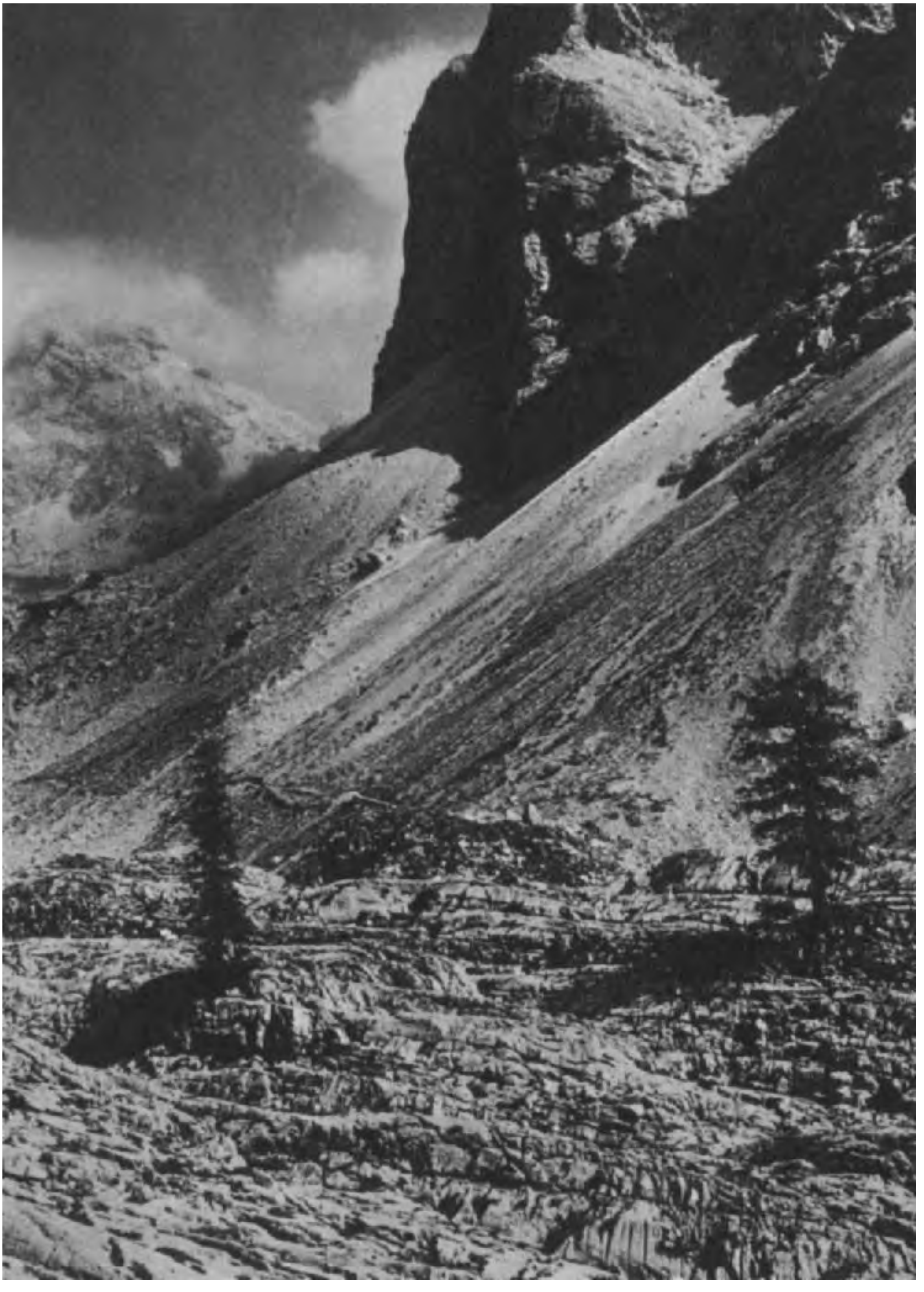


Fig. 113. Denudation and karst development of limestones in the region of the upper timberline of the valley of the Triglav Lakes (Julian Alps). (Photo P. Plesník.)

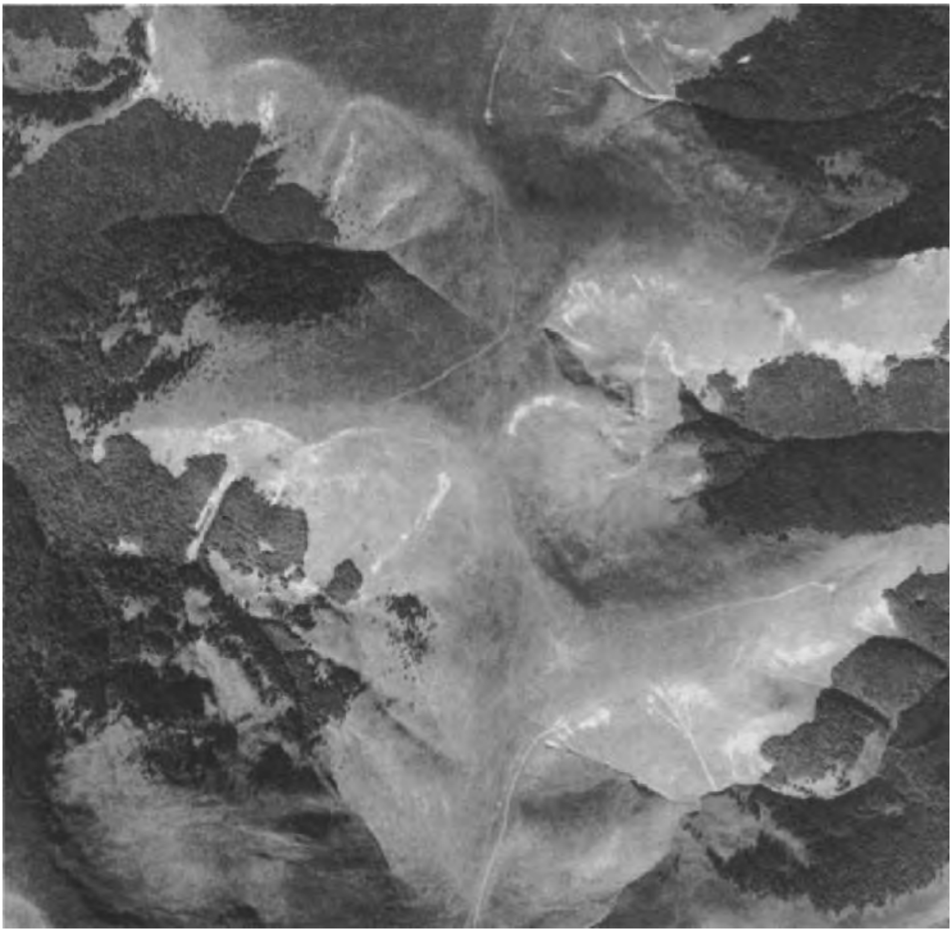


Fig. 114. Snow-melt and rain-wash erosion on deforested slopes of the Veľká Fatra Mountains (Czechoslovakia). (Aerial photo.)

Elevation above the sea level – geographical location

Both these factors have an indirect effect on erosion by their influence on physical conditions.

With increasing *elevation* and at *higher geographical latitudes,* the temperature generally decreases and the amount of precipitation increases. However the intensity of precipitation increases with elevation and decreases at higher latitudes. The combined influence of temperature, precipitation, wind, potential energy of the relief, and surface structure affect the erosive activity of both precipitation and wind. These, together with other geomorphic factors create a varied pattern of destruction phenomena.

The aggressivity of an erosion factor such as water or wind is of vital importance and depends on the available kinetic energy and the degree of protection afforded by vegetation, the latter depending in turn, on the available supply of water and solar energy.

In Czechoslovakia, a height of 200 to 250 m above the sea level may be regarded as a level at which important changes take place. Below this elevation, wind erosion and locally weak sheet erosion prevail. In the belt between 200–250 and 600 m, wind erosion occurs only sporadically on loess, while sheet erosion prevails; most of the gullies and wastelands are to be found here. In the 600 to 1,400 m belt, erosion is slight in spite of the high potential energy of the relief, and the soil is well protected by forests. In areas above 1,400 m (1,200 to 1,600 m, according to the mountain range) the palette of erosion is highly variegated, and sheet, gully and wind erosion together with various combinations of frost and snow phenomena occur here.

However when the vegetation is removed, soil erosion quickly increases to a level of high intensity and this may cause a rapid removal of soil and the exposure of the bedrock where the soil is shallow and permeable. Figure 113 shows a state of complete devastation of land in the Julian Alps of Yugoslavia after the destruction of the forest vegetation. Figure 114 shows the revivification of erosion processes in the region of the upper timberline in the Veľká Fatra Mountains (CSSR). The intensification of erosion is mainly caused by thawed snow water on the leeward side of ridges where large quantities of snow accumulate.

4.2.6.4 Vegetation

Vegetation is of vital importance as a protection for the soil against precipitation erosion. It protects the soil against the action of falling raindrops, increases the degree of infiltration of water into the soil, maintains the roughness of the soil surface, reduces the speed of surface runoff, binds the soil mechanically, diminishes microclimatic fluctuations in the uppermost layers of the soil, and improves the physical, chemical and biological properties of the soil. Where there are favourable conditions for the formation of a continuous vegetation cover, the extent of erosion is well below harmful levels. Where the conditions are less favourable for the growth of natural vegetation, the protective effect of a vegetation cover can very easily be lost as a result of man's interference.

This is also true with respect to protection of the soil by cultivated forms of vegetation. In this case the soil is protected only slightly, if at all, during certain parts of the growth cycle. Thus plant growth provides many different forms and degrees of soil protection, and the protection given to the soil against precipitation

Table 62. Removal of soil under different types of vegetation cover

Locality in the CSSR, erosion agent, author	Crop	Angle of slope inclination	Soil removal [m³ ha⁻¹]	Paired comparisons	
				[%]	[multiples]
Fintice, spring wash, 1956	Winter cereals	15°	27.7	100.0	1.0
Zachar (1970)	Thin clover	15°	14.4	52.0	1.9
Csadné, spring wash, 1965	Field with rolled surface	6°20′	91.0	100.0	1.0
Midriak (1965)	Field with broad furrow	9°40′	42.0	46.2	2.2
Hiadeľ, downpour on 23rd May 1958	Potatoes	14°	235.9	100.0	1.0
Zachar (1970)	Rye	14°	10.4	4.5	22.0
Lučatín downpour on 23rd May 1958	Potatoes (forest above the field)	12°	182.2	100.0	1.0
Zachar (1970)	Clover	12°	5.0	2.7	36.4
Lučatín, downpour on 23rd May 1958	Potatoes (pasture above the field)	8°12′	623.1	100.0	1.0
Zachar (1970)	Wheat	8°12′	21.7	3.5	28.8
Hlohovec, downpour on 3rd June 1958	Potatoes, maize	15°	363.0	100.0	1.0
	Old vineyard	15°	123.0	31.0	3.0
Veľké Žarnoseky, growth period	Fallow	24°	125.9*	100.0	1.0
1959—1963	Soil protected by vegetation				
Holý (1964)	partially	24°	19.7*	15.6	6.3
	fully	24°	0.2*	0.16	629.5

* Data are given in ton per ha for the period 1959—1963.

erosion by one particular crop may vary under different conditions. This means that one indicator is not adequate for the assessment of the soil conservation values of different crops. Nevertheless, the soil conserving effect of vegetation may be classified into a number of categories. Let us first examine some data on erosion losses which have occurred either, during downpours or over longer periods of time (Table 62).

Table 62 shows that during spring soil wash resulting from the thawing of snow, differences between bare and partially protected soil are relatively small. Other observations have revealed the following relative degrees of spring wash occurring under different cropping conditions:

Tilled land with rolled surface	100%
Tilled land with rough furrow	40–50%
Winter cereals	25–30%
Thin clover, (a) one-year-old stand	12–15%
(b) several years old	< 1%

As the growth of a crop progresses, its capacity to protect the soil increases, and at the time of first heavy rains at the end of April and in May, winter cereals afford adequate protection even against the more violent downpours. However root and tuber crops, as well as specialized crops only begin to protect the soil effectively in the second half of June and early July. Data obtained in Czechoslovakia indicate the following relative degrees of protection afforded by various crops in May and June:

Potatoes, maize	100%
Spring cereals	70–80%
Older vineyards	30–40%
Winter cereals	3–5%
Grass stands	< 1%

The arrangement of crops on the slope is also important from the point of view of soil protection. The low grass growth on pastures protects the soil against erosion, but during downpours a lot of runoff is created, and therefore lower lying crops are damaged more severely. It has been found that soil erosion on land situated below pastures may be 5 to 20 times greater than on land adjacent to a forest or a shelterbelt. Differences in erosion on protected and unprotected land may be as high as 630-fold.

Measurements made by Gerlach (1976) in the Tatra Mountains revealed that rates of soil removal from forests, pastures (meadows), and fields, respectively, occurred in the ratio 1 : 25 : 30,000 on the upper part of the slope, and 1 : 1,066 : 15,666 on the lower part of the slope. This means that permanent grass

Table 63. Rates of removal of soil protected by various types of vegetation

Place, source	Crop	Slope inclination	Rate of soil removal [t ha^{-1} year^{-1}]	Comparisons with fallow [%]	Comparisons with fallow [multiple]
Southern	Fallow	10 %	148.3	100.0	1.0
Piedmont	Cotton	10 %	69.9	47.1	2.1
Bennett (1955)	Crop rotation	10 %	32.0	21.6	4.6
	Grass	10 %	0.7	0.5	211.9
	Forest	10 %	0.004	0.0003	37,075.0
Wisconsin,	Fallow	16 %	427.8	100.0	1.0
Bennett (1955)	Maize	16 %	250.2	54.8	1.7
	Maize in rotation	16 %	62.3	14.5	6.9
	Grasses	16 %	0.2	0.05	2,139.0
Cuba	Fallow, ploughed deep 10 cm	4 %	36.0	100.0	1.0
Forns (1957)	Fallow, ploughed deep 20 cm	4 %	31.0	86.1	1.1
	Maize	4 %	16.0	44.4	2.3
	Rice	4 %	6.0	16.7	6.0
	Grass	4 %	0.25	0.7	144.0
GFR	Fallow	11 %	15.6	100.0	1.0
Kuron (1947)	Oat stubble	11 %	0.45	1.8	34.7
	Thin clover	11 %	0.12	0.7	130.0
Romania,	Fallow	12°	16.5	100.0	1.0
Arvam (1956)	Maize	12°	13.3	80.6	1.2
	Potatoes	12°	7.9	47.9	2.1
	Spring wheat	12°	0.9	5.5	18.3
	Winter wheat	12°	0.8	4.8	20.6
	First year grass	12°	0.4	2.4	41.3
	Second year grass	12°	0.2	1.2	82.5
Azarbaijan,	Grass stand				
Gadzhiev (1962)	canopy 0.1—0.2	—	552	100.0	1.0
	0.3	—	330	59.8	1.7
	0.4	—	174	31.5	3.2
	0.5	—	142.5	25.8	3.9
	0.6	—	102	18.5	5.4

and forest reduced soil wash 15 to 1,200 times, and 15 to 30,000 times, respectively, compared with field crops. In the forest the annual removal was found to be 3×10^{-4} m^3, i.e. a negligible amount.

Under central European conditions agricultural soil is least protected against precipitation erosion outside the growing period, especially where thawing occurs bringing snow water with it one or two months after sowing or planting, which in any case tend to coincide with the beginning of the rainy period; specialized crops, such as vines and fruit trees do not give much protection to the soil at any time of year.

Also the foreign literature contains a large amount of data on the different effects of vegetation from the point of view of erosion control (Table 63).

Crops which develop late in the season and crops that require hoeing or other attention afford only a little protection against erosion, and reduce the erosion losses, which would occur if the land were barren, by 10 to 50%. The degree of erosion control achieved by crops increases two- or threefold where a crop rotation is followed. The soil conserving effect increases proportionally with plant density and the cover of the canopy; a fully closed grass stand protects the soil completely, reducing erosion to a harmless level.

From various sources of data the following scale of the relative soil conservation effects of different crops may be set up:

Fallow, or barren land	100%
Orchards with managed soil	80–90%
Sugar beet, grain maize	85%
Root and tuber crops, specialized crops	50–80%
Spring cereals	30–50%
Winter cereals	5–35%
One-year-old grass stands	1–5%
Older grass stands	0.5%
Forest	0.01%

It should be noted that substantial deviations from the above limits may occur depending on the climate and the condition of the crop. In general, forests and permanent pastures reduce erosion below the tolerable limit. The effect of a particular crop at any stage of its growth may be determined in most cases from the degree of soil cover afforded by the foliage or the crop canopy. The effect of root systems which have remained in the soil after harvesting and which are of advantage in crop rotation situations must also be taken into account.

The overall soil conserving effect of all forms of vegetation is currently being calculated from the area covered, and the degree of protection given by various crops and types of natural stand.

4.2.6.5 Agricultural measures and logging

Soil cultivation plays an important part in the reduction of erosion mainly on account of the effect on surface roughness, soil permeability, soil resistance against destruction caused by raindrops and surface runoff, freezing of the soil, and the mobilization of nutrients and water for plant growth. It is generally accepted that

Table 64. Rates of soil removal with different types of soil management and logging operations control

Place, source	Method of soil cultivation	Rate of removal [t ha^{-1} year^{-1}]	Comparisons with control conditions [%]	[multiple]
Bashkiria,	Ploughing up and down slope	457	100.0	1.0
Sobolev and	Ditto + autumn ploughing	331	72.4	1.4
Sadovnikov (1956)	Ploughing across slope	74	16.2	6.2
	Ditto + autumn ploughing	43	9.4	10.6
Crimea,	Unterraced slope (25—28°)	69.0	100.0	1.0
Velichko (1962)	Terraced slope	15.0	21.7	4.6
Yowa (9 % slope)	Fallow, with fertilizer			
Bennett (1955)	0 t ha^{-1}	130.97	100.0	1.0
	17 t ha^{-1}	103.19	78.8	1.3
	35 t ha^{-1}	81.98	62.6	1.6
	Maize, with fertilizer			
	0 t ha^{-1}	49.43	100.0	1.0
	17 t ha^{-1}	20.69	41.9	2.4
	35 t ha^{-1}	10.59	21.4	4.7
Zakarpatskaya,	Gravitational timber	357.0	100.0	1.0
Polyakov (1962)	skidding			
	Skidding by tractor	270	75.6	1.3
	Skidding by cable	36.0	10.1	9.9

soil cultivation, *fertilizing, irrigation,* and *crop distribution* according to *rotation practice* are basic soil *conservation measures* applied on *agricultural land* by means of which erosion on land of low and medium erodibility may be reduced to a harmless level. In addition, also mulching of the ground and its reinforcement by incrustation are important factors in the reduction of erosion. Table 64 shows the effects of different measures in combating erosion losses.

Thus in the arid region of the Bashkir Autonomous SSR soil erosion was reduced as much as 6.2-fold by ploughing along contour lines and as much as 1.4- to 1.7-fold by introduction of autumn ploughing. Agricultural measures gave improvements in infiltration and the water regime of the soil, and increased the roughness of the soil surface.

An important degree of soil protection was achieved by terracing on steep slopes that were poorly protected by vegetation; erosion losses being reduced by as much as 4.6 times. The soil was coarse-grained with a tendency for mudflow.

Important differences were achieved by fertilizing, so that soil loss on fallow land was 1.6 times smaller as a result, and losses on a maize field were as much as 4.7 times smaller; the overall reduction in erosion losses achieved by growing maize and fertilizing the soil amounted to a 12.37-fold drop, and erosion diminished from a very serious level to being slight or even permissible.

Table 65. Coefficient of contour ploughing and alternation of crops in strips

Inclination [%]	Contour ploughing	Width of strips [m]	Alternation of crops
2—7	0.5	30—35	0.25
7—12	0.6	25—30	0.3
12—18	0.8	20—25	0.4
18—25	0.9	15—20	0.45

The survey includes information on soil erosion caused by logging. Total soil losses caused by skidding of timber over the ground (exploitation erosion), by the resulting precipitation erosion, by both of these factors together, by cableway skidding, by the consequent precipitation erosion, and by both of the latter together, were 192, 357, 549, 31.5, 36.0, and 67.5 t ha^{-1}, respectively. The greatest reduction in soil losses due to precipitation erosion achieved by using different logging methods was a 9.9-fold decrease and there was a 8.1-fold reduction with respect to overall erosion losses. Erosion soil losses were proportional to the fractional area of soil destroyed by logging. Although erosion

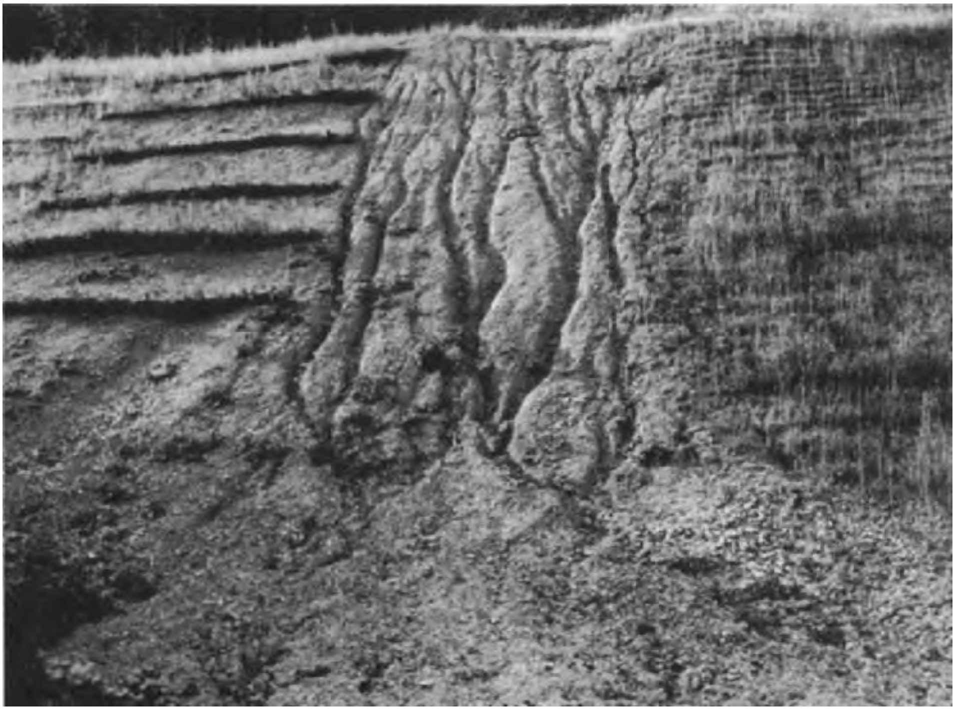


Fig. 115. Slope stabilization trial in Slovenia (Yugoslavia). (Photo F. Rainer.)

subsequent to logging diminishes over the years, losses nevertheless continue to be high on shallow forest soils.

According to experience gained with respect to tilled land in the USA, the relationships between slope gradient and the effect of ploughing along the contour lines on the one hand, and the effect of strip rotation of crops on the other, are characterized by coefficients 0.5 to 0.9 and 0.25 to 0.45, respectively (Table 65).

For gradients exceeding 24% the effect of contour line ploughing is minimal and also the width of the fields becomes much less important. Consequently, other measures are required under these circumstances.

It is important to have information on the soil conserving effects of various protective measures on the most exposed parts of slopes, and where the rocks show little resistance to weathering. Figure 115 shows an experiment in the soil conservation using "living" fences, cordon planting, and the sowing of grass in various combinations. In the centre of the figure there is the unprotected control plot.

4.2.6.6 Complex assessment of sheet precipitation erosion

The successful combating of sheet precipitation erosion requires that the influence of all factors and conditions be expressed in an integrated form. Such an expression may be used for different purposes, including determination of the level of accuracy that is required. Expressions of aggregate influence are most often used in determining the intensity of soil erosion, i.e. for establishing the rate of decrease in the thickness of the soil profile and the soil losses that occur within a certain time interval, usually within a year. According to the average intensity of soil erosion, it is possible to estimate (a) *the threat imposed on the soil by erosion,* (b) *the distribution of erosion* over a particular region, (c) *the potential soil erosion* in the absence of a vegetation cover, (d) the degree of *actual erosion,* and (e) the *expected rate of soil erosion* which may occur as a consequence of changes in *soil utilization* and *soil protection.*

Finally, calculations of the intensity of soil erosion form an important part of the interpretation both of *experimental findings* and of long-term *empirical observations of soil erosion,* and are essential for the selection and application of effective conservation measures.

Therefore, much attention in the world literature has been given to the determination of erosion intensity, and many empirical formulae have been developed. In order to obtain accurate results from these formulae, sufficiently accurate data are necessary, otherwise deviations from reality of up to tenfold may result and the calculations are then less reliable than an expert estimation thus defeating their object of improving the understanding of erosion processes and failing in optimizing the value of erosion control measures.

In this respect much appreciation is owed to the efforts of Zingg (1940), Musgrave (1947), Wischmeier and Smith (1965), and other authors, who have searched for a universal equation, valid for all the possible geographical conditions occurring on our planet, that would assist in the determination of erosion losses. These equations with their various adaptations are in use in different countries, and are widely regarded as being the only method for determining soil losses. The complications involved in the investigation of soil erosion have already been indicated in the methodological section of this work, and in the following discussion attention is drawn to the snags inherent in generalizing information which is to be inserted in a universal equation (the example given in Sec. 3.3.14 is used).

First of all it should be noted that the erosion intensity is being determined for arable land with an assumed management régime, and without channelling of the runoff. On arable land, the course of runoff and erosion is different from that on a grazed surface, or on meadow land or cleared forest land. Even on the same surface the properties of tilled land change from one year to the next.

Furthermore, the universal equation is derived, taking rain erosivity as the main factor in the precipitation erosion of the soil, and is based on the kinetic energy of erosive rain. In order to determine correctly the relationships between rain erosivity and soil erodibility, usable results need to be obtained for regions where downpours of short duration are common. During downpours of longer duration with low intensity and high wash, this equation gives low values. In addition, this method of calculation does not apply to regions, in which there is snow thaw erosion exclusively, and it is only partially valid for mixed regions where the soil is eroded both by rains and by thawed snow water.

In the CSSR, the latter mixed type of erosion occurs over most of the territory. In countries situated to the north and east, the soil is threatened mostly by thawed snow water. The various types of precipitation erosion affect areas as indicated in Table 66 (Zaslavskiĭ 1977).

These data show that the universal equation in the proposed form would be valid for about 10% of the territory in the USSR. In the CSSR about 65% of a total of 4 million tons of silt, and in the SSR about 74% of the silt flows off during the

Table 66. Surface area of soil endangered by various types of precipitation erosion in the USSR

Erosion caused by	Surface area [km^2]	Proportion of total area [%]
Snowmelt	8.5×10^6	56.7
Snowmelt and rain	4.9×10^6	32.7
Rain-water	1.6×10^6	10.6
Total	15.0×10^6	100.0

winter and spring period. Not only may the calculated value deviate from the real value by more than ± 65%, but there is also a basic difference in principle, namely the difference between the calculation of erosion on the basis of all types of erosion and the same calculation based on only one type of erosion, in particular on the type that represents smaller share of the total. Unfortunately, no practical method has as yet been developed for determining the intensity of soil erosion caused by snow thaw, and therefore in those regions where this form of erosion is important, the determination of the total soil losses is more complicated.

Total soil losses in sheet erosion

For practical purposes where sheet erosion is concerned, the author recommends the use of a method which takes into account the proportions of the erosion losses due to rain and snow thawed water, respectively. Total soil losses caused by precipitation erosion can be expressed by the relation

$$E_t = E_i + E_r + E_s + E_{ch},$$

where E_t is the total erosion, E_i the impact erosion (raindrop splash and hail impact erosion), E_r the runoff erosion (rainwash and rill erosion), E_s the snow thaw erosion, and E_{ch} the chemical erosion.

As well as these forms and types of soil erosion, on tilled land there is also aration erosion, E_a, caused by the displacement of soil on the slope by ploughing. Although this type of erosion does not form a part of precipitation erosion, it is mentioned in this chapter for practical reasons. By introducing aration erosion, the equation for the calculation of total erosion in regions predominantly affected by precipitation erosion then takes the form

$$E_t = E_i + E_r + E_s + E_{ch} + E_a.$$

The procedure for obtaining quantitative values should take account of all ascertainable types and forms of sheet erosion.

Impact or splash erosion

Impact (splash) erosion, as mentioned earlier in Sec. 4.2.1 and 4.2.2, is of importance mainly with respect to soil losses on the upper parts of slopes, on ridges, on elevated tracts of land where the soil permeability is high and there is little resistance to erosion. The recording of soil splash is also important where erosion occurs on narrow terraces, on ridges on hill farms, etc.

From results obtained by Ellison (1952), Mirtskhulava (1970), Hudson (1971), and others, it may be supposed that raindrop action on a 10% gradient during a period of rain lasting 10.7 min with an intensity of 2.8 mm min^{-1} will cause soil displacement of approximately 100 g m^{-2} [t ha^{-1}] over a distance of 1.5 m. The displacement of this amount of soil requires an amount of energy of about 10 MJ ha^{-1}. Hensch (1970) has established in the course of research in Morocco that a soil displacement of 100 t ha^{-1} dissipates approximately 100 MJ ha^{-1} of the energy of the water running over the surface.

At a rough estimation, from 0.6 to 3.0 t ha^{-1} of soil is displaced on a 10% gradient by precipitation delivering 10 MJ ha^{-1}. The lower limit refers to resistant soil, and the upper limit to easily erodible soil. The degree of impact erosion caused by raindrop action varies with slope inclination according to the relation $E_2 = f(I^{1.5})$; erosion caused by hail is correspondingly greater. Unfortunately, there are no data in the literature referring to the latter type of soil destruction.

It must be stressed, however, that the total quantity of soil splashed by rainfall is substantially higher, and is estimated to represent tens or even hundreds of tons per ha. If it is supposed that the kinetic energy of so-called erosive precipitation with a large splash effect varies from about 1 kJ m^{-2} in the temperate region to 15 kJ m^{-2} in the tropical region (i.e. from 10 to 150 MJ ha^{-1}), the average soil loss on a 10% gradient may be expected to lie in the range 1 to 15 t ha^{-1} $year^{-1}$.

These data, although greatly lacking in precision, nevertheless show that on slope ridges and similar places erosion may equal or even surpass tolerable levels. As a consequence, soils under these conditions tend to be shallow and are damaged by the selective influence of splash erosion. Therefore in such cases, levels of precipitation erosion need to be assessed separately, and impact erosion must be included within the total erosion balance.

Aration erosion

A similar effect is also produced by *aration erosion* (E_a) which increases with increasing slope inclination and increasing velocity of the agricultural implement. Even if the ploughing follows the contours and the furrows are thrown both against and down the slope, soil is still displaced down the gradient. The intensity of aration erosion can be computed from the relation

$$E_a = \frac{h\Delta_x \gamma \, . \, 10^4}{100} \, .$$

More details are given in Sec. 4.2.5. The equation is valid for a surface area of 1 ha measuring 100 × 100 m. If it is required to know the intensity of aration erosion on a narrow strip adjacent to the ridge, the value of the denominator must be reduced and the value for the rate of soil ploughing must be set relatively higher. Example: For a soil depth of 0.3 m, a soil displacement of 0.3 m down the slope, a specific weight of the soil of 1.5 t m^{-3}, and a 10 m width of the ploughed field, the intensity of aration erosion will be: $E_a = 0.3 \times 0.3 \times 1.5 \times 10{,}000 \div 10 = 135$ t ha^{-1}. The weight of only one furrow measuring $0.3 \times 0.3 \times 100$ m will be 13.5 t – a value exceeding tolerable erosion. Thus the soil is continually induced to slide down from the upper parts of the slope, where the soil profile becomes thinner and the bedrock becomes exposed.

For the sake of comparison, it may be noted that tractor ploughing involves the dissipation of about 300 MJ ha^{-1} for the displacement (for 30 cm ploughing depth) of 3,000 m^3 soil per ha in one operation, i.e. 15 times more than the lower limit set for catastrophic erosion (200 t ha^{-1}), and 12.5 times more than the greatest amount of soil that is splashed during heavy downpours (240 t ha^{-1}). Such a level of interference with the soil mantle requires an assessment of possible unfavourable consequences for the fertility of the upper and convex parts of the slope.

Rainwash erosion

This is the form of erosion which has received the greatest attention in the world literature, and empirical relationships for its calculation have been the most sought after. The use of these, of course, depends to a large degree on the reliability of the input data. In the CSSR, the method of Frewert [adapted by Zdražil (1965), Stehlík (1970, 1975) and Míchal (1973)] has been used for mapping and large scale regional planning. Several studies of the occurrence of potential and *actual erosion* in the CSSR have been undertaken, using the Frewert–Zdražil method. In addition to this method, that of Wischmeier and Smith (1965) has also been used and adapted for conditions in Czechoslovakia by Pretl (1970), as well as being used for the assessment of erosion inhibition in forests (Papánek 1971).

Frewert-Zdražil method, as modified by *Stehlík,* is based on the equation

$$x = DGPS,$$

where x is the potential soil erosion [mm year^{-1}], D the climatic factor, G the petrological factor, P the soil factor, and S the slope factor.

By means of this relationship the potential erosion of soil unprotected by vegetation, and the effectiveness of control measures can be assessed. The actual erosion can be computed from potential erosion if x is multiplied by the slope length, L, and the proportion of crops providing little protection against erosion, O, as follows

$$x = DGPSLO.$$

The climatic factor D in the equation is expressed in terms of the precipitation the duration of which is at least $\sqrt{5t}$, consequently the intensity is $i = \sqrt{5/t}$ in mm min^{-1}. Processing of precipitation data in the CSSR gives the following values for the climatic factor D, for precipitations of 10 to 60 minutes duration with the following frequencies:

Average frequency of rain	0.8	1.0	1.2	1.4	1.6	1.8	2.0	2.2	2.4	2.6
Climatic factor, D	0.26	0.32	0.38	0.45	0.51	0.57	0.63	0.70	0.76	0.82

The value of the climatic factor in the CSSR varies between 0.26 and 0.82, the lower values applying to the lowlands and the higher values to mountain regions.

The distribution of the various rock groups according to resistance and permeability was studied in detail (Table 67).

Table 67. Values of the rock coefficient, G, in relation to rock characteristics

Permeability of rock	Granulation of weathered debris	G
Low	Fine	1.5—1.3
Slight	Sandy loam	1.3—1.1
Moderate	Loamy sand	1.1—0.9
High	Coarse sand to stony	0.9—0.7

The values of the geological factor, G, vary between 0.7 and 1.5.

The soil factor should express the erodibility of the soil. The authors of the equation derived this factor from the proportion of coarse clay ($\varnothing < 0.1$ mm) and the humus content for the various soil types (Table 68).

Table 68. Values of the soil coefficient, P, for different soil characteristics

Type of soil	Content of clay (<0.01 mm) [%]	Content of humus		
		<2 %	2—3 %	>3 %
Sandy	<10	1.4	1.1	1.0
Loamy sand to sandy loam	10—30	1.5	1.25	1.75
Loamy	30—45	1.25	1.0	0.8
Clay / loam	45—60	1.4	1.15	0.9
Clay	>60	1.5	1.25	1.0

The extreme values of the soil factor calculated in this way show only a twofold difference. Real values of soil erodibility show much greater differences; Hensch (1970) gives of up to 2,400-fold erodibility differences for various rocks and soils in Morocco.

The slope factor is expressed in terms of the gradient according to the relationship: $S = 0.24 + 0.106p + 0.0028p^2$, where S is the slope factor, p the slope gradient (%), and S has the following values:

Slope gradient [%]	5	7	9	12	15	20	30	40	50
Slope factor, S	0.35	0.65	1.0	1.45	2.0	3.0	5.35	8.61	12.02

The figures show that the outcome of the calculation of erosion depends most of all on the slope gradient; if the gradient increases tenfold (i.e. from 5 to 50%) the slope factor becomes 34.3 times greater.

Stehlík used the shortened form of the equation, $x = DGPS$, to calculate the overall potential erosion for soil in the CSR. The maximum value obtained on an unimportant surface was 14.45 mm year^{-1}. Over the larger part of the territory, the intensity of erosion varied from 1.0 to 5.0 mm year^{-1}, i.e. from 10 to 50 m^3 ha^{-1} year^{-1}.

Midriak (1977) based a study of forest regions in the CSSR on the more general form of the equation $x = DGPSLO$, where L is the slope length factor and O the vegetation factor (the proportions of low resistance vegetation and forest in the region).

Values for the slope length factor were derived as follows:

Slope length [m]	20	50	100	150	200	250	>300
Slope length factor, L	1.0	1.6	2.5	3.2	3.8	4.3	5.0

Thus if slope length increases twofold, erosion increases approximately 1.5 times.

The value of the vegetation factor, O, was obtained as follows:

Vegetation of low resistance to erosion [%]	0	5	10	20	30	40	50	60	80	100
Forest cover [%]	100	95	90	80	70	60	50	40	20	0
Vegetation factor, O	0.20	0.25	0.3	0.4	1.0	1.22	2.0	2.5	3.2	4.0

The method is thus based on a model in which erosion under a full forest cover diminishes to 20% of the level occurring under a forest cover of 70%, and in which erosion increases fourfold if the forest cover is completely removed. There is a twentyfold difference in the levels of erosion occurring under 100% forest cover and crops of low resistance to erosion, respectively.

Midriak made calculations for an average slope length of 300 m and a 100% forest cover on forested land. Analyzing the potential erosion of the soil on 597 forest working units in the CSSR covering a total area of 4,436,528 ha, he established that the mean potential soil removal was 1.54 mm year^{-1} (0.77 mm for the CSR and 2.63 mm for the SSR).

These data refer not to potential, but to actual erosion, the degree of erosion attributed to forested territory being rather high. The method is well-suited for assessing relative levels of soil erosion in regional planning. When comparing region with region, the data are consistent with the author's findings (Zachar 1970) derived from silt flow measurements and data on soil erosion for the whole soil mantle, as can be seen from a comparison of erosion intensities expressed in m^3 ha^{-1} $year^{-1}$

Author	CSR	SSR	CSSR
Zachar (1970)	2.5	7.5	5.0
Midriak (1977)	7.7	26.3	15.4

In the first case actual erosion of the entire soil mantle was estimated, and in the second case reduced level of potential erosion for the forest mantle of the country was arrived at.

Another method of calculating rainwash erosion was proposed by Wischmeier and Smith (1965) and is currently in use in many countries as a universal method. This method is well-known under the name of the *Universal Soil Loss Equation* (USLE), and therefore some brief comments and a few examples of results obtained will suffice here. The original form of the equation *($A = RKLSCP$)* is given in Sec. 3.3.15. It is used for the determination of both actual and expected soil erosion in different farming and erosion control situations, mainly on ploughed soil.

The USLE is based on the assessment of soil erosion losses according to *rain erosivity* (factor R), calculated from the kinetic energy of a 30-minute period of rain, EI_{30}. Equal values of this factor are shown on the map as isoerodents, which range in practice from negligible values of about 50 up to 600, exceptionally even higher. A problem arises when assessing erosive and non-erosive rains in regions where experimental measurements are not available and climatic conditions differ from those in the USA.

The second basic factor in the USLE is K, which expresses the erodibility of the soil. This factor is calculated from measured values and expresses the quantity of eroded soil arising from the dissipation of a constant amount of rain energy (constant rain erosivity) on a 9% gradient of 22.6 m length. The K factor varies for different soils ranging from negligible values (0.03) for the most resistant soil types up to 0.69 for soil types most susceptible to erosion.

By multiplying rain erosivity *(R)* by soil erodibility *(K)*, the intensity of soil erosion *(A_e)* on the 22.6 m long "etalon" plot with its 9% gradient is established. Accordingly, the extreme values of A_e vary within the limits $A_{e\ min} = 50 \times 0.03 = 1.5$, and $A_{e\ max} = 600 \times 0.69 = 414$ t $acre^{-1}$ $year^{-1}$. After conversion to the international system (1 American ton $-$ 0.9071853 t, 1 acre = 0.404678 ha), values ranging from 3.3 to 920 t ha^{-1} $year^{-1}$, i.e. approximately 0.22 to 61.3 mm $year^{-1}$ are obtained. The difference between the extremes of the range may occasionally be greater.

In the CSSR, the rain factor takes lower values (e.g. 30) for regions with the lowest annual precipitation (450 mm), and for the wettest regions which will still support agricultural crops this factor increases to 50 or 60. The rain factor, R, was calculated according to $R = 0.068H$, where H is the annual precipitation expressed in mm. Wischmeier (1970) calculates the precipitation energy from the formula $E = 10.3 + 89 \log_{10} I$, where E is the energy of 1 cm of rain in t m^{-1} ha^{-1}, and I the intensity of the rain in cm h^{-1}.

The *soil factor, K,* has a relatively wide range of values for conditions in the CSSR and varies from 0.1 to 1.0 t ha^{-1} $year^{-1}$. The range of A_e, the "etalon erosion", is determined by the extreme values of R and K, and lies between 3.0 and 30 t ha^{-1} $year^{-1}$; easily erodible soils are found only in regions characterized by a low rain factor. For soils of different degrees of erodibility the values of the critical slope given in Table 69 have corresponding values of K.

Table 69. The soil factor, K, and the critical inclination, S_k, for soils in Czechoslovakia (see also Table 57)

Soil grade	Erodibility of soil	Critical slope inclination, S_k [%]	K
I	Very high	< 5	1.00
II	High	10	0.75
III	Moderate	15	0.50
IV	Low	20	0.25
V	Very low	25	0.10
VI	Intrasoil erosion	>30	>0.04

The *critical inclination of the slope, S_k,* is the angle of inclination at which *acute soil erosion* occurs, which under the prevailing conditions of the CSR averages approximately 0.5 mm $year^{-1}$, i.e. 7.5 t ha^{-1} $year^{-1}$. This is approximately ten times the level of erosion which can be compensated for by soil formation. If these increased losses are offset by soil fertilization and accelerated soil formation by intensive cultivation of agricultural crops, acute erosion may then by considered as the same, under these conditions, as admissible erosion.

According to this view of soil loss caused by rainwash erosion, the "etalon erosion", A_e, for soils in the CSSR with minimal resistance to erosion would be $A_e = RK =$ (from 30 to 50) $\times$ 1.0 = 30–35 t ha^{-1} $year^{-1}$. These A_e values for a 9% gradient and a slope length of 22.6 m are probably among the highest in the CSSR. However under conditions of exceptionally heavy rainfall, the value of A_e may become as much as ten times greater. The lowest A_e values should not be less than 3.0 t ha^{-1} $year^{-1}$.

From the established "etalon erosion", which in some way represents the potential soil erosion on the etalon plot (according to USLE), the total potential erosion, A_p, may be derived from the formula $A_p = A_e\ LS$, where LS is a slope length and slope inclination factor. The factor LS may be computed from the empirical expression

$$LS = \frac{l}{100}\,(1.36 + 0.97s + 0.1385s^2)\,,$$

$$L = \left(\frac{l}{22.13}\right)^P,$$

where l is the slope length expressed in metres, the exponent P varies from 0.3 for gradients of up to 10%, to 0.6 for gradients exceeding 10%. Values of L are given in Table 70.

Table 70. Values of the slope length factor, L

Slope length [m]	5	10	15	20	30	40	50	60	80	100	150
L	0.48	0.68	0.82	0.95	1.17	1.35	1.52	1.66	1.91	2.13	2.61
	200	250	300	350	400	450	500	600	700	800	900
	3.02	3.38	3.69	3.99	4.27	4.52	4.77	5.22	5.64	6.04	6.39

$$S = \frac{0.43 + 0.3s + 0.043s^2}{6.613}\,,$$

where s is the slope gradient in %; values for S are given in Table 71. The factor LS is given in Fig. 116.

Table 71. Values of the slope steepness factor, S

Slope inclination [%]	2	3	4	5	6	7	8	9	10	11	12
S	0.18	0.26	0.35	0.45	0.57	0.70	0.84	1.0	1.17	1.35	1.55
	13	14	15	16	18	20	22	24	26	28	30
	1.75	1.97	2.21	2.46	2.99	3.57	4.21	4.90	5.64	6.43	7.28

Calculation of ***potential soil erosion***, A_p, from $A_p = RK\ LS$ gives the highest rate of soil removal in the absence of soil protection by vegetation and erosion control measures. The value thus computed may still be adjusted for slope aspect or for any other relevent factors. Example: let us consider a region with rain factor $R = 30$, a medium resistant soil with soil factor $K = 0.5$, a slope of gradient $S = 10\%$ and length $L = 100$ m, $LS = 2.5$. Under these conditions the potential erosion A_p is

$$A_p = RK\ LS = 30 \times 0.5 \times 2.5 = 37.5 \text{ t ha}^{-1} \text{ year}^{-1}.$$

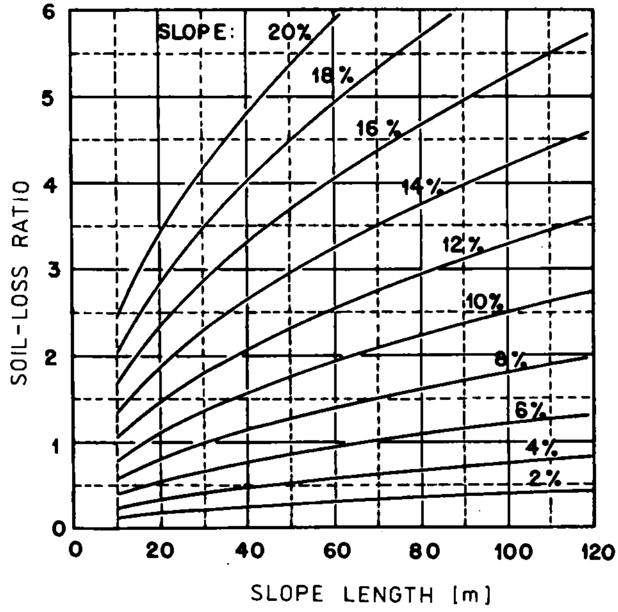


Fig. 116. Slope factor chart. Values of *LS* (in the erosion equation) are read on the vertical scale.

In order to prevent undesirable losses, this potential erosion must be reduced to the admissible or *tolerable level*, A_t, which generally amounts to 7.5 t ha^{-1} year^{-1}. This means that in the above example, the potential erosion needs to be decreased by 30.0 t ha^{-1} year^{-1}.

The planner or manager may select a suitable method of soil conservation from available cultivation schemes, taking into account the appropriate selection of crops, crop rotation, fertilizing, interruption of the slope gradient, terracing, or other measures. Since the reduction required is small in absolute terms, the desired results may be achieved by soil cultivation, fertilizing, crop distribution, and rotation.

Selected values of C for the main crops are given in Table 72.

Example: if in the region described above, erosion losses are reduced (a) by introducing transverse ploughing as 30% of all ploughing (coefficient $P_a = 0.7$), (b) by increasing fertilizer applications by 20% (coefficient $P_f = 0.8$), and (c) by introducing 65% specialized crops into a rotation with cereals (coefficient $C_c = 0.35$), then A_p, the potential erosion, will decrease to the desired level of erosion: $A_a = A_p C P_a P_f = 37.5 \times 0.35 \times 0.7 \times 0.8 = 7.35$ t ha^{-1} year^{-1}.

This example shows that on the greater part of the agricultural land of the CSSR, where moderately erodible soils, slope gradients of up to 10%, and slope lengths of up to 100 m are common, rainwash erosion of the soil may be controlled effectively by straightforward management procedures. Only on the more erodible soils and on steeper and longer slopes are other control measures required, such as shorten-

Table 72. Values of the crop factor, C, for principal agricultural crops according to Wischmeier

Crop, crop sequence	C	
	Winter	Spring
Cereals following clover	0.25	0.30
Cereals following cereals, sown in ploughed land	0.35	0.55
Cereals following cereals, sown in the stubble	0.30	0.40
Cereals following root and tuber crops	0.40	0.50
Maize following meadow	0.24	
Maize following meadow, strip ploughing (stalk yield, 3 t ha^{-1})	0.05	
Maize two years after meadow	0.46	
Maize three years after meadow	0.55	
Maize following cereals, stubble turned with the plough, stalks remaining	0.46	
Potatoes, sugar beet	0.60	
Lucerne, clover	0.02	
Clover and grass meadow	0.005	

ing the erosive parts of slopes by strip cultivation, establishing infiltration and erosion control belts, increasing soil unevenness, etc.

For the sake of comparison, let us examine the calculation of potential soil erosion by the method of Frewert–Zdražil and Wischmeier–Smith.

Example of use of Frewert–Zdražil equation, $x = DGPsL$: annual precipitation 500 mm, climatic factor $D = 0.32$; rock permeable, rock factor $G = 1.1$; soil sandy to loamy, humus content 2 to 3%, soil factor $P = 1.25$; slope gradient 10%, slope gradient factor 1.15; slope length 100 m, slope length factor 2.5. Thus potential erosion $x = 0.32 \times 1.1 \times 1.25 \times 1.15 \times 2.5 = 1{,}265$ mm $year^{-1}$ = 12.65 m^3 ha^{-1} $year^{-1}$ $\doteq$ 19 t ha^{-1} $year^{-1}$.

Example of use of Wischmeier–Smith equation, $A = RK\,SL$: annual precipitation 500 mm, $R = 34$; soil slightly to moderately erodible, $P = 0.38$; slope gradient 10% and slope length 100 m, $LS = 2.5$. Potential erosion is therefore $A = 34 \times 0.38 \times 2.5 = 32.3$ t ha^{-1} $year^{-1}$; for medium resistant soils the potential erosion would be 42.5 t ha^{-1} $year^{-1}$.

For regions in the CSSR with a mean annual precipitation of 720 mm the following values are obtained

$$x = DGPsL = 0.63 \times 1.1 \times 1.25 \times 2.5 = 2.49 \text{ mm year}^{-1} = 37.35 \text{ t ha}^{-1} \text{ year}^{-1},$$

$$A = RK\,LS = 48.96 \times 0.38 \times 2.5 = 46.51 \text{ t ha}^{-1} \text{ year}^{-1},$$

or, for medium resistant soils, 61.2 t ha^{-1} $year^{-1}$.

The examples show that for soils of low elevations in agriculturally important regions, the values of potential soil erosion obtained by the method of Wischmeier–Smith are 1.7 to 2.24 times greater than the values obtained by the Frewert–Zdražil method. For soils of moderate elevations, the former values of potential soil erosion are 1.25 to 1.64 times greater than the latter. In the first method there are greater differences in the rain factor and smaller differences in the soil factor, compared with the second method.

A relatively large difference between these two methods is to be found in the respective degrees of importance attached to the role of vegetation. According to the author's measurements, the magnitude of the erosion control effect of cereals growing over a large area relative to the erosive effect of the rainfall, is greater than that quoted in the literature. When the first downpours of the season arrive, the cereals, especially winter cereals, are already well-developed and reduce erosion to 3 to 5% of the level occurring on unprotected soil or soil only slightly protected by specialized crops. Indeed, winter cereals should be able to reduce the potential level of soil erosion to a tolerable level, even if the former were as high as 150 t ha^{-1} $year^{-1}$. This is valid in the case of land that is ploughed transversely to the line of steepest descent, and which supports a cereal crop of at least average biomass.

It follows that as well as correctly assessing potential erosion, it is also necessary to obtain the possibly most reliable assessment of the effectiveness of erosion control measures. A separate study is to be devoted to the problems of soil conservation by erosion control.

It is obvious that any considerations concerning erosion control measures will greatly affect values set for levels of admissible erosion, A_p, which the author takes as being 0.5 mm $year^{-1}$ (7.5 t ha^{-1} $year^{-1}$) for conditions in the CSSR. Extreme values for admissible erosion are 0.05 to 0.8 mm $year^{-1}$. Based on the assumption that the intensity and overall effect of weathering manifest in changes in the depth of the soil, and in the thickness of the weathered layer, respectively, the following values for tolerable erosion may be derived for soils of different depths:

Soil depth [cm]	Tolerable soil loss [mm $year^{-1}$]	[t ha^{-1} $year^{-1}$]
<30	0.05	0.75
30—60	0.2	3.0
60—120	0.5	7.5
>120	0.8	12.0

If valid relationships are established between the various indicator parameters of soil erosion, the admissible slope length can be determined regressively and compared with the general equation for the calculation of slope length (Sec. 4.2.6.3). The value for the admissible slope length may be determined using the Wischmeier–Smith equation as follows

$$L'S = \frac{A_t LS}{A_p CP},$$

where A_t is the level of admissible erosion, A_p the potential erosion, C the vegetation factor, and P the erosion control factor.

In the example given for the calculation of erosion, the fertilization factor, $P_f = 0.8$, was omitted; this results in a reduction in the estimated admissible slope length from 100 to 65 m, as can be seen from the following

$$L'S = \frac{7.5 \times 2.5}{37.5 \times 0.35 \times 0.7} = \frac{18.75}{9.19} = 2.04.$$

Figure 116 shows that at a constant gradient of 10% the $L'S$ value corresponds to the slope length (65 m).

In a similar way, it would be possible to calculate the value of the tolerable inclination S' at a constant slope length L, where terracing is contemplated. Such a calculation would be valid for steeper inclinations. For the sake of objectivity, it is possible to make a calculation using as follows

$$LS' = \frac{A_t LS}{RKCP} = \frac{18.75}{9.19} = 2.04.$$

The S' is the change of inclination – 8.65% – valid for a constant slope length. If the admissible erosion has not been exceeded, then it is necessary in this case to adjust the gradient to 8.65%, i.e. to reduce it by 1.35% (terrace height, 1.35 m per 100 m).

The admissible slope length, L', may be derived as follows

$$L = \frac{A_t}{RKSCP} = \frac{7.5}{34 \times 0.38 \times 1.17 \times 0.35 \times 0.7} = 2.03.$$

The admissible slope length for a gradient of 10% and for a L factor value of 2.03 is 65 m (Fig. 116).

When water is converging in channels on uneven terrain or for other reasons, the admissible slope length becomes shortened according to the erosion load factor of the soil – a factor which represents the increase in the size of the collection area on the upper fanned-out part of the slope relative to that of a straight slope. If, for example, the erosion load factor is 1.2, the admissible slope length is then reduced in this instance from 65 to 54.2 m (65 ÷ 1.2 = 54.2).

It follows from this that the purpose of the USLE is the calculation of the intensity of soil erosion – that is, erosion losses arising from the dissipation of the energy of precipitation, and the raindrop impact effect. Wiliams (1972) modified the USLE, replacing the R factor in the calculation of sediment weight by the empirically derived erosion activity of the surface runoff *(G)* according to the relationship

$$G = \frac{\alpha\,(Qq_p)^{\beta}}{A_r}\,KSLCP\,,$$

where G is the sediment yield [t acre^{-1}], Q the runoff volume [ft acre], q_p the peak flow rate [ft^3 s^{-1}], A_r the drainage area [acre], α, β are constants, and K, S, L, C, P are USLE factors.

The constants α and β, determined from events on 20 watersheds in Texas, Nebraska and Iowa, were 95 and 0.56, respectively. The exponents are nearly equal, indicating that a change in the product, Qq_p, has about the same effect.

Both of these equations represent extremes of approach to the erosive activity of rain-water. The first equation is based on the energy of the rain, whereas the second considers only the surface runoff of downpours. As a matter of fact, both raindrop action and runoff activity participate in the erosion and transport of eroded particles, and they complement one another.

Foster and Wischmeier (1973) proposed a modified equation which takes into account both rainfall and runoff

$$A = aR + bcQq_p^{1/3}\ KSLCP,$$

where a, b are weighting factors ($a + b = 1$), c is the equality coefficient, R the rainfall factor, Q the runoff volume [m], q_p the runoff rate [m h^{-1}], and K, S, L, C, P are USLE factors.

The weighting parameters reflect the relative amounts of erosion caused by rainfall and by runoff under unit conditions (i.e. when $S = L = C = P = 1$). There is a limited amount of experimental evidence which indicates that erosion is approximately equally divided between rainfall erosion and runoff erosion under these conditions (i.e. $a = 5b$). The equality coefficient c can be estimated from simulation plot data under unit conditions

$$c = 2A_u(K - R)\ Qq_p^{1/3},$$

where A_u is the soil loss occurring under unit conditions. The value of c for erosion in Indiana in the USA varies around 30. Substituting these values of a, b, and c into Foster's equation, the following equation is obtained

$$A = EKSLCP,$$

where $E = 0.5R + Qq_p^{1/3}$

From this equation any of its components can be derived. Foster and Wischmeier (1973) computed transport capacity from the equation

$$A = \frac{EKSCP\,l^{0.5}}{8.521}\,.$$

Transport over x metres distance (in lbs/foot) was calculated according to

$$A' = \frac{EKSCP\,x^{1.5}}{185.58},$$

and the transport capacity at a point at distance x was found from

$$A_{cx} = \frac{E_x\,KSCP}{185.58}\,1.5\,,$$

where A_{cx} is the transport capacity at x.

According to this equation the soil removal from any part of the slope, or from the entire eroded territory may be computed.

Foster also derived a relationship for calculating approximately the ratio between rill erosion and inter-rill erosion for a length of slope, in which the removal rate of loose soil is at a maximum. According to Foster

$$\beta_j = \left(\frac{K_r}{K_i}\right)_j \left(\frac{15\,Qq_p^{1/3}}{0.5R}\right)_j S_s\,,$$

where K_r/K_i is the rill/inter-rill soil erodibility ratio; when $K_r/K_i = 2$, "severe rilling" occurs, when $K_r/K_i = 1$, "moderate rilling" occurs, and when $K_r/K_i = 0.5$, there is "little evidence of rilling"; S_s is the S factor for a portion of slope of known length.

The relative proportions of *sheet erosion* and *rill erosion* change with distance down the slope, and the total erosion increases as rill erosion increases causing a simultaneous reduction in the selective effect (Frere et al. 1975).

Snow thaw erosion

The intensity of *soil erosion* caused by thawed snow water is largely determined by the rate of production and total quantity of melt water, the permeability of the soil, the disintegration of soil aggregates by frost, and the moisture content of the soil.

The rate of thawing is usually substantially lower than the rate of rainfall. The former is recorded in millimetres per 24 hours [mm day^{-1}]. However, in view of the fact that soil is frozen in winter and saturated with water in the surface layers, the rate of infiltration is minimal and varies on loam and clay soils between 0.01 and 1.0 mm day^{-1}. Consequently a considerable portion of the melt water flows away, so that the runoff coefficient for melt water on frozen soil is usually higher than that for precipitation water. Surface runoff occurs mainly during a continuous thaw when a substantial part of the lying snow melts within 10 to 20 days. This effect, as far as erosion is concerned, is greater than the melting of snow during an influx of warm air accompanied by rain.

The greatest rates of surface runoff from thawed snow water are from 0.001 to 0.08 mm min^{-1}, whereas the highest rates of rainwater runoff are 4 to 5 mm min^{-1} Normal values for snow thaw runoff lie between 1.0 and 15 mm day^{-1}.

Although erosion caused by thawed snow water does not reach the same intensity as that caused by rain, it affects large areas only slightly protected by vegetation, and soil dispersion is one of the primary consequences.

The *intensity of snow thaw erosion* may be computed from the empirical formula

$$E_s = mhk\ LS\ CPK,$$

where E_s is the intesity of soil erosion [t ha^{-1} $year^{-1}$], m the rate of thawing of snow [mm day^{-1}] in the 20-day period, in which the most intensive thawing takes place; in regions where the melting of snow is hastened by rain, the m value increases by 50 to 100%, h the amount of water derived from snow during the 20-day period [cm], k for rain-water runoff multiplied by a number between 1.5 and 3, and L, S, C, P, K are USLE factors, C and P pertaining to the period after the growth season.

Thus where $m = 2.0$ mm day^{-1}, $h = 4$ cm, $K = 0.5$, $k = 0.5 \times 1.5 = 0.75$, and $LS = 2.5$, then $E_s = 2.0 \times 4.0 \times 0.75 \times 2.5 = 15$ t ha^{-1} $year^{-1}$. It has been assumed that the soil is saturated with water in the period before the intensive thawing of the snow.

The snow cover may be distributed unevenly, and this can influence the distribution of snow erosion. Most snow is accumulated on the leeward sides of slopes, behind barriers and in depressions where the intensity of snow thaw erosion may be several times higher. On the windward sides of slopes and on sunny slopes the quantity of snow gathered is smaller, and therefore snow thaw erosion is less intense.

Snow thaw erosion is one manifestation of *snow erosion* which is generally denoted by the term *nivation* and includes also snowdrift erosion of the soil. Nivation phenomena are significant mainly on mountain massifs and they are an important factor in the formation and destruction of both the soil and the relief. The intensity of nivation is closely associated with the intensity of *pluviation,* i.e., the aggregate influence of rain-water and snow is affected mainly by anemoorographic systems.

Chemical erosion

The methods of assessing erosion mentioned so far have been concerned with the mechanical destruction of soil by virtue of the kinetic energy of raindrops, hail, rain-water flowing over the soil surface, thawing of snow, and by other mechanisms included in this section for the sake of completeness. Precipitation water also affects the soil chemically, mainly by dissolving the more readily soluble substances contained within the soil including ecologically active substances produced by the metabolism of plants for the regulation of growth, substances added to the soil as fertilizers, and also damaging organic chemicals, such as pesticides and industrial contaminants.

In dealing with *chemical soil erosion* the main problem is to limit (a) losses from the soil of easily soluble and ecologically active substance, (b) losses of fertilizers and the leaching out of pesticides or other agrochemicals, and (c) the leaching out of damaging substances and contaminants which arrive on the soil surface from industrial fumes.

An important part of this task is reduction of the precipitation water proportion going into the surface runoff, and improvement of the absorption capacity of the soil.

In this context the control of erosion losses caused not only by the mechanical force of raindrops and surface precipitation water, but also by the force of surface runoff in general is important. By diverting some of the surface runoff into underground flow, so that chemically harmful substances are filtered out of the water, not only are erosion losses reduced, but an improvement in the quality of the water is achieved; on intensively cultivated land this is an exceptionally important factor in the conservation of the living environment.

Special attention should be given to maintaining the chemical stability of the soil in regions where high salt content may lead to salination, intrasoil washing of salt, and intrasoil and tunnel erosion.

The importance of chemical soil erosion is illustrated by the fact that in the CSSR more than 250 kg ha^{-1} of chemically purified nutrients are added to the soil annually, and that precipitation water deposits on the soil surface a quantity of harmful substances which is 210 to 350 kg ha^{-1} more than the quantity deposited in the period before large-scale industrialization. The amount of harmful matter which enters watercourses from precipitation water is about 2.0 million tons out of a total quantity of material washed from the land surface of 8.5 million tons. About 4.0 million tons of this are represented by suspended matter and 0.25 million tons by bed load (i.e. particulate proportion, 4.25 million tons). According to this estimation, about 2.25 million tons of the material getting into watercourses does so on account of the chemical erosion of soil by surface precipitation water. The proportion of material which arrives in the watercourses from underground and mineral water outlets is about 1.0 million tons. These quantities of material are carried by about 28.4 million m^3 of polluted surface water per annum.

Harmful matter brought by precipitation water	2.0 million tons
Matter brought by underground and mineral water	1.0 million tons
Other material washed from the land	6.5 million tons
(including the particulate fraction	4.25 million tons)
Total pollution	9.5 million tons

Of this amount, about 1.8 million tons sediment in water reservoirs and ponds annually.

The quantities of harmful chemicals carried down by rainwater (2.0 million tons) and chemical substances washed from the fields (2.25 million tons) amount to about 50% of the total quantity of material. Of the 2.25 million tons of chemical substances leached from the fields, about 1.45 million tons account for the removal of agrochemicals (0.7 million tons), soluble substances washed out by selective erosion (0.75 million tons), and substances removed from the soil together with the particulate material of soil (0.8 million tons).

Although this is an approximate balance sheet for soil erosion based on the proportions of suspended load, bed load, soluble matter, quantities of nutrients in the soil and amounts of leached agrochemicals, it shows that chemical soil erosion is of great importance, particularly for the reason that this material plays an active role in determining soil fertility. Chemical soil erosion can be assessed by the proportions of soluble matter in surface precipitation water, but if these data are not available, the intensity of chemical erosion may be estimated from the proportion of "non-erosive precipitation" and its surface runoff, the amounts of chemicals added to the soil in the form of fertilizers and agrochemicals in general, and the amounts of pollutants deposited on the soil surface and carried away with surface water.

As an example, let us consider a region with 720 mm precipitation, a precipitation runoff coefficient of 0.25, and a runoff coefficient for "erosive precipitation" of 0.7.

In this case, "erosive precipitation", erosive snow melt water, and erosive surface runoff with a runoff coefficient of 0.7, are equivalent to 6.8% (49 mm), 5.6% (40 mm), and 62.3 mm of the total precipitation, respectively; "non-erosive precipitation" is $100.0\% - 6.8\% - 5.6\% = 87.6\% \doteq 630$ mm, and "non-erosive" runoff is equivalent to 126 mm, i.e. about twice the "erosive" runoff. Whereas in "erosive" precipitation the particulate fraction in the eroded substance ranges from 70 to 95%, in "non-erosive" precipitation the relative proportions of particulate and dissolved material are reversed.

Stehlík (1968) calculated ***chemical erosion*** according to the suspended load, using the formula

$$Qe = Qm - \frac{Q1 + Q2 + \ldots + Qn}{n} mk ,$$

where Qe is the quantity of chemical constituents displaced by water as a consequence of intensive soil erosion, Qm the quantity of chemical constituents displaced by water during large discharges of water or at high concentrations of suspended load, $Q1 \ldots Qn$ the daily discharge of suspended load during low discharge of water and at low concentrations of suspended load, n the number of days of low discharge and low concentration of suspended load, m the number of days of high discharge and high concentration of suspended load, and k a factor representing the contribution made by other chemicals (Stehlík used $k = 2.0$).

Using this method, Stehlík established that in the catchment area of the Jihlava Creek, 24% of the applied fertilizer was washed away during the period of observation. Moreover, in the investigated catchment area erosion was 18 times less intense than in the heavily eroded catchment area of the Trkmanka Creek.

When fertilizer applications amount to 10 t ha^{-1} year^{-1}, and 20% of the applied substance flows away, chemical losses represent 2 t ha^{-1} year^{-1}. In industrial regions where the fall-out of pollutants is 1,000 t km^{-2} and runoff is 20%, again 2 t ha^{-1} year^{-1} of this material is washed away. These approximate data indicate that for an annual application of about 1.75 million tons of agrochemicals, and a fall-out of about 6.3 million tons of pollutants over the land surface of the CSSR, an estimated chemical erosion of 3 to 5 t ha^{-1} year^{-1} is most probable.

Thus the quantity of chemical substances finding its way from the soil surface into watercourses is considerable, and must be taken into account in the balance of erosion losses. It must be added, unfortunately, that methods for the calculation of chemical erosion are still at an early stage of development, although a significant contribution was made in A Model Chemical–Soil–Water Interactions, prepared by the ACTMO (Frere et al. 1975).

Total losses caused by sheet precipitation erosion and ploughing

Following from the above-mentioned examples of the calculation of the various sheet, or laminar components of precipitation and aration erosion, it is possible to move on to the calculation of the total erosion (E_t) according to the formula

$$E_t = E_i + E_r + E_s + E_{ch} + E_s,$$

where E_t is the total precipitation erosion of the soil [t ha^{-1} year^{-1}] without protection from vegetation, E_i the impact erosion caused by raindrops; in the CSSR this averages about 1 t ha^{-1} year^{-1}, E_r the rainwash erosion, sheet, inter-rill and rill erosion together average about 40 t ha^{-1} year^{-1}, E_s the snow thaw erosion, this is about 15 t ha^{-1} year^{-1}, E_{ch} the chemical erosion, estimated to be 3 t ha^{-1} year^{-1}, and E_a the aration erosion, in one ploughing operation of 0.3 m depth, in which the soil is displaced 0.3 m down the slope, the aration erosion is 13.5 t ha^{-1} year^{-1}.

The *total loss of soil* from these types and forms of erosion is thus

$$E_t = 1.0 + 40.0 + 15.0 + 3.0 + 13.5 = 72.5 \text{ t ha}^{-1} \text{ year}^{-1}.$$

For an admissible erosion level, E_p, of 10 t ha^{-1} year^{-1}, $E_t - E_p$ is double the value of $E_r - E_p$ as computed by the USLE equation. The larger the intensity of snow thaw erosion and aration erosion, the greater is the difference between the calculated total losses and admissible loss.

The calculation may be verified by calculating back from data on the flow of soluble and insoluble particles in watercourses. Such an experiment was carried out by Stehlík (1976) in selected watercourses of the CSSR and involved the calculation of the transition stage of soil erosion according to the equation of Lopatin (1950, 1958)

$$k_q = \frac{E_0}{Q_p - E_l},$$

where k_q is the ratio between sheet precipitation erosion on tilled land and the quantity of erosion products entering the watercourses, E_0 the intensity of sheet precipitation erosion on tilled land [t], E_l the product of linear erosion in the catchment area [t], and Q_q the discharge of suspended load into watercourses [t].

Stehlík calculated E_0 according to the Frewert–Zdražil–Stehlík equation: $x = DP\ LS\ HOC$, where $DP\ LS$ is the "potential erosion", and HOC are factors representing the effects of fertilization, crop rotation, and soil erosion control procedures, respectively. For 18 catchment areas with a total surface area of 989,567 km^2, the average annual soil erosion, E_0, was 1,075 t km^{-2} (0.7169 mm year^{-1}), the average annual linear erosion, E_l, was 1.17 t km^{-2} (0.011 mm year^{-1}), the average discharge of suspended load into watercourses, Q_p, was 47.61 t km^{-2}, and therefore k_q was 1,075 t (47.61 − 1.17 = 46.44). The percentage of the soil product that is washed into watercourses is expressed by

$$E_k = \frac{Q_p - E_l}{\frac{E_0}{100}} = \frac{47.61 - 1.17}{\frac{1{,}075}{100}} = 4.32\ \%.$$

According to these calculations the value of E_k varies under different conditions from 0.6 to 12%, the lower values (from 0.6 to 4.2%) occurring on permeable cretaceous formations, moderate values (from 2.6 to 7.8%) in broken hill country and plateaus, and high values (from 8.0 to 12%) being found for rugged mountain regions. For the USSR, Lopatin gives an E_k range from 3 to 20%.

Given a quantity of 4.0 million tons for the suspended load in the rivers of the CSSR (Zachar 1970) and an E_k value of 10%, the intensity of soil erosion amounts to about 40 million tons, i.e. an average of 8 t ha^{-1} year^{-1} on cultivated land, the total area of which is 5.0 million ha. Of this area, about 2.9 million ha – the area of cultivated land (2,890,000 ha) which exceeds the critical slope inclination – are endangered by precipitation erosion, and if the average intensity of precipitation erosion on cultivated land were about 14 t ha^{-1} year^{-1} for an E_k value of 4.3% (as was established by Stehlík on selected catchment areas), then the average soil removal in the CSSR would amount to 93 million tons (4 million tons : 4.3 × 100%), and 18.6 t ha^{-1} $year^{-1}$ (93 million tons : 5 million ha) of arable soil, respectively, and 32 t ha^{-1} $year^{-1}$ (93 million tons : 2.9 million ha) of arable soil endangered by erosion, respectively.

This figure does not include soil displacement over short distances by raindrop action, ploughing of the land, and the washing out of soluble substances, nor does it include erosion over the remaining territory of the country. The latter erosion is far from negligible and occurs mainly on pastures, field and forest roads, and on forest land during logging (about 2% of potential erosion), as well as in the alpine and subalpine belts where an important degree of soil destruction occurs (Midriak 1982).

According to levels of potential soil erosion caused by precipitation, maximum expected levels of erosion over large land areas may be established and mapped in terms of potential erosion. The potential erosion expresses the highest possible level of precipitation erosion that can occur in the soil is unprotected by vegetation, as mentioned earlier. In any other circumstances, the actual erosion should be of a lower value, except in the case of erosion taking place under unnatural conditions where, for example, artificial channelling of the surface runoff may occur as a consequence of exceptionally steep man-made slopes, etc.

Where vegetation provides an adequate cover or control measures have been put into practice, the soil is protected against erosion to a certain degree, i.e., actual erosion is less than potential erosion. The degree of soil protection, and the effect of erosion control measures may be expressed by the formula

$$A_e = \frac{E_p - E_e}{E_p} ,$$

where A_e is the anti-erosion effect, E_p the potential erosion, and E_e the actual level of erosion achieved through soil management.

In practice A_e varies from 0.05 for the smallest anti-erosion effect of control to almost 1.0 for permanent, closed stands. The higher the potential erosion and the smaller the level of tolerable erosion on soils easily damaged by erosion, the stronger are the erosion control measures that need to be applied.

4.2.7 Gully erosion

Soil wash occurs as soon as the velocity of the surface runoff exceeds the critical limit at which external forces of the flow expressed by its energy are greater than the internal forces expressed by the coherence of the soil. The critical – tolerable, velocity of water is the highest water velocity at which the wash of soil, or rock, does not yet occur.

According to Mirtskhulava (1970), the intensity of gullying may be expressed by the formula

$$I_g = 0.2 \times 10^{-4}\, d\omega \sqrt{\frac{v_{\Delta i}}{v_{\Delta k}} - 1}\,,$$

where I_g is the intensity of gullying, d the mean diameter of the soil particles, ω the area of the transversal section of the flow, $v_{\Delta i}$ the initial water velocity, and $v_{\Delta k}$ the extreme velocity of the water flow.

Water velocity $v_{\Delta x}$ in the point x of the beginning of gullying (top of the gully) depending on the surface F of the catchment area above this point, i.e. $v_{\Delta x} = \mathrm{f}(F, x)$ and the length of gullying $l = L - x$, where L is the distance of the bottom of the gully. The shorter the distance x, the smaller the area and intensity of growth of the gully. The depth of falling water, i.e. the depth of gullying (h) in time (t) is, of course, also of decisive importance

$$h_t = h\left(\frac{t}{T}\right),$$

where h_t is the depth of gullying in the time t, h the expected depth of gully stabilization, t the time period of gullying, T the time period of gully stabilization,

$$T = 10^4 h/0.2d\,\sqrt{v_i/v_k - 1}\,.$$

Knowing the catchment area, the amount of surface runoff caused by rainwater or thawing snow in the given region, the expected water velocity and its gullying effect may be computed from the slope length, inclination and configuration which determines water concentration. The basic equation according to Chézy is

$$v_x = c\sqrt{RI}\,,$$

where v_x is the water velocity at distance x metres from the divide, c the coefficient of water determined by the surface roughness of the soil; according to Bazin $c = 87\sqrt{y}/(\gamma + \sqrt{y}) \doteq 87/\gamma \times \sqrt{y}$, R the hydraulic radius, I the inclination of the terrain, y the depth of runoff, and γ the roughness coefficient varying from 2.0 for fields ploughed up and down the slope to 15.0 for soils with a very rough surface.

The *critical water velocity* for erosion of upper soil horizons is best derived using the formula of Velikanov (1948)

$$v_k = 3.13\sqrt{14d + 0.006}\,,$$

where v_k is the critical water velocity [m s^{-1}], and d the mean diameter of the soil particles.

For clay and fine particles $v_k = 3.13\sqrt{0.006} = 0.24$ m s^{-1}. For soils in the USSR, Soviet authors have calculated values for the critical water velocity at which gullying is just about to start, these varying between 0.3 and 25 m s^{-1} (Kosov et al. 1976). Data are given in Table 73.

Table 73. Critical water velocities for various soil and rock types

Characteristics of soil and rock	Water velocity [m s^{-1}]
A. Loose quaternary sediments	0.3—2.0
1. Sand of all grain sizes	0.3—0.55
2. Loess and loess earth, peat	0.65—0.75
3. Loam of marine origin	0.55—1.0
4. Heavy clay	1.0—1.3
5. Compact earth	1.0—1.5
6. Gravel loam of different origins	1.5—2.0
B. Sediments and frozen rocks	2.1—6.0
7. Soft sediments (cretaceous, solonetzic)	2.1—3.1
8. Limestones	2.5—4.5
9. Hard sediments (sandstones)	5.0—6.0
10. Frozen quaternary rocks	5.0—6.0
C. Very hard crystalline rocks	6.0—25.0
11. Tuffs and similar rocks	6.0—16.0
12. Monolithic crystalline rocks	16.0—25.0

In fact, the critical water velocity is associated with the ***resistance of earth*** and ***rocks to gullying***. As shown earlier, doubling the water velocity increases the erosive capacity (expressed as a function of the kinetic energy) of the water flow fourfold, since the kinetic energy changes in proportion to the square of velocity. Doubling the water velocity also increases about 64-fold the individual volume (or weight) of particles that may be transported along the bottom of the flow channel, because the particle mass, $Q = Av^6$, where A is a constant of proportionality and v the water velocity (Eri's law).

Where water accumulates in rills, depressions, or artificial furrows and channels, etc., friction decreases, the velocity of the water flow increases, and consequently the erosive force and carrying capacity of the water both rapidly increase, too. Gullying activity increases still more when overflow from gullies takes place and the excess flow causes retrograde erosion which enhances the gullying process. In this way, a single downpour may give rise to large erosion rills or gullies (Figs. 45, 99), with the water carrying away whole blocks of earth as well as small stones, large stones and boulders. It has been observed that in those areas where the water flow becomes channelled on the lower parts of slopes or in depressions, the kinetic energy attained during one downpour may be up to 2,500 times greater than the kinetic energy of sheet runoff.

Therefore gully erosion occurs only under certain conditions and usually takes a characteristic course. Gully density is usually not greater than 10 km per km^2, the surface area covered by gullies does not as a rule exceed 15% of the total surface area, and the number of gullies is seldom more than 70 per km^2. When these values are exceeded, linear erosion changes into polymorphous erosion, in which the surface is substantially more divided and dissected; other destructive phenomena such as intrasoil erosion, landslides, solifluction, etc. occur at the same time, too.

In an analysis of gully erosion in the USSR, Kalinichenko and Il'inskiĭ (1976) established a classification of gullying which shows the relationships between various parameters of gully erosion (Table 74).

Table 74. Classification of gully erosion

Degree of erosion	Erosion parameter		
	Density of gullies [km km^{-2}]	Rate of gullies [%] or [ha km^{-2}]	Number of gullies per 1 km^2
Very slight	<0.15	<2	<1
Slight	0.15—0.6	0.2— 0.9	1—4
Moderate	0.6 —2.2	0.9— 3.5	4—17
Severe	2.2 —9.0	3.5—14.0	17—67
Very severe	>9.0	>14.0	>67

Comparing the data in Tables 73 and 74, it is clear that the critical water velocities for gullying activity in the sites of lowest and highest resistance show a difference of 83-fold, and parameters for gully erosion in defined regions of the USSR show differences of 67- to 80-fold, the rule being that the intensity of gullying increases with decreasing rock resistance. Obviously the intensity of gully erosion depends on other factors also.

The most important factor governing gully erosion is the *climate,* since this determines the aggressivity of the erosion process and the rate and type of plant growth – the most important element in erosion control. In general, the intensity of gully erosion increases from the tundra and forest-tundra regions, where potential gully erosion is very low, to the taiga zone with low to moderate erosion and to the forest zone with moderate potential erosion. In the steppe and forest-steppe zones the rain erosivity is approximately the same, compared with the former regions, but owing to the higher temperatures, the protective effect of vegetation is usually less, and therefore potential erosion in these zones is, as a rule, greater. In any one zone erosion increases with the increased precipitation that occurs in some years. In the semidesert and desert regions potential erosion is small, owing to the low amount of precipitation, but because of the high intensity of the rainfall potential erosion is usually greater than in the tundra and forest-tundra zones.

The larger the kinetic energy of the precipitation, the greater is the potential erosion. On the other hand, potential erosion increases as the availability of water decreases, as expressed by the relation

$$W_s = \frac{R - R'}{t},$$

where W_s is the available moisture, R the annual precipitation [mm], $R' = 30(t + 7)$, the annual precipitation required for growing crops, and t the mean annual temperature.

At low values of W_s, the available moisture for plant growth is low, and therefore the protective influence of vegetation is reduced. If large amounts of precipitation fall in mountainous areas, there is a danger of gully erosion developing into a flow of soil and earth (mudflow, etc.).

As far as the relative contributions made to the overall erosion by rain-water and thawed snow water are concerned, the kinetic energy of surface runoff is diverted more towards the enlargement of gullies, whereas the kinetic energy of precipitation goes into the loosening of earth. Consequently, erosion intensity depends mainly on the depth of surface runoff which is related to the degree of convergence of flow.

As an example taken from among much available data, the author presents some observations on the linear growth of slope and valley gullies in the Moldavian SSR (Rozhkov 1973) in Table 75.

Table 75 shows that valley gullies grow six times more rapidly than slope gullies, and during the growing season, gullies develop at half the rate observed outside the growth season. Detailed measurements revealed that thawing of snow resulted in soil removal from the catchment area of 11.3 t ha^{-1} where there were slope gullies, and 15.5 t ha^{-1} where there were valley gullies. In summer soil removal was less than in winter, being caused by a few episodic downpours. The average turbidity of the water in spring gullying was less than that in gullying caused by rainstorms.

Another parameter of climate which is of relevance to erosion is the hydrothermal coefficient (HTK) which is defined according to

$$\mathrm{HTK} = \frac{\Sigma R}{\Sigma t}\, 10,$$

Table 75. Growth of gullies in Moldavia

Type of gully	Gully growth [m]		
	Winter — Spring	Summer — Autumn	Annual
Valley	4.55	2.16	6.36
Slope	0.49	0.56	1.05

Fig. 117. Valley gully in an arid region of Tunisia. The surrounding land is used for agricultural purposes.

where ΣR is the aggregate precipitation, and Σt is the summation of air temperatures above 10°C.

In the CSSR, Σt varies from 1,200°C in the mountains to 3,600°C in the lowland, precipitation varies from 450 to 1,500 mm or more. Most gullies occur in regions with HTK values between 1.25 and 2.5. A detailed mapping of erosion processes according to these parameters has not yet been carried out.

Figure 117 shows part of a ramifying gully in an arid region of Tunisia increasing in dimensions; the gully was created by rainstorms in an area of low relief.

Figure 118 shows a site in Iraq entirely destroyed by gully erosion. This is again an arid region where the vegetation, and with it the soil also, have been destroyed by extensive over-utilization.

Another factor affecting gully erosion is the nature of the *soil* and the *underlying bedrock*, the resistance of which is characterized by the critical water velocity required to cause gullying. The largest gullies occur on loess sites. The intensity of gully erosion decreases in the direction: soil over sandy substrata $\rightarrow$ soil over loess, clayey and heavy loam substrata $\rightarrow$ skeleton soils.

Fig. 118. Slope gullies in the Kuh-e Sorkh ridge (Iran); in the foreground are recent deposits indicating the very intense levels of erosion in the catchment area. (Photo F. Papánek.)

Table 76. Data on gully erosion in Czechoslovakia

Substratum, soil	Slope inclination	Slope length [m]	Soil removal [m^3 ha^{-1}]
Sandstone loamy to sandy soils	2—8°	220—1940	16,400
Andesite, loess loam	3—13°	88	6,600—10,560
Flysch sandstones, shallow skeleton soils	6—20°	100—250	1,073
Dolomite, medium deep skeleton soils	27—30°	53—140	1,620—5,360
Dolomitic limestone, shallow skeleton soils	25—30°	120—230	1,120—2,640
Dolomitic limestone, very shallow skeleton soils, Tatra Mountains	30—38°	450	1,100

Table 77. Data on gullies occurring on a slope 100 m wide and 160 m long on loess loam

Distance from the border of the field [m]	Angle of slope inclination	Number of gullies	Soil removal [mm]	Surface area covered by gullies [%]
20	4°34′	4	11.0	8.1
40	6°53′	9	35.7	16.5
60	8°21′	9	49.4	18.6
80	10°06′	10	68.8	23.6
100	10°54′	11	121.0	24.2
120	14°52′	12	176.9	28.2
140	9°18′	12	174.3	29.0
160	8°02′	5	22.4	10.4
180	9°02′	12	91.8	21.5

The intensity of gully erosion is strongly influenced by the thickness of loose, easily erodible or moderately erodible sediments. Measurements in the CSSR (Zachar 1970) have shown that the total removal of soil and bedrock attributable to gully erosion depends on the depth of loose weathered material and the resistance of the bedrock (Table 76).

The part played by the relief in gully erosion is similar to that in sheet erosion, but the length of the slope, and the surface area of the collection area are both of greater importance in gully erosion. In the CSSR, most gullies occur on gradients of 5 to 10°, and about 90% are found on gradients within the range 2 to 15°. The slope aspect is of considerable importance, as discussed in the previous section. The most pronounced gullies occur on thick sediments and slope coverings, and in terrain depressions on which a lot of water converges.

Gully erosion may commence on a gradient of 1°, and on susceptible soil, flat gullies are created. An investigation of rills in loess strata in eastern Slovakia

Fig. 119. Gully erosion on slopes of 2 to 4° inclination developing in loess loam in eastern Slovakia (Czechoslovakia). (Photo D. Zachar.)

showed that rills are formed on fields on gradients of 2 to 3°, and owing to the presence of furrows, a relatively dense pattern of rills is created, occupying an average of 22% of the surface area. Some data on gullies developing parallel to the steepest axis of the slope during a period of 30 years, are given in Table 77. The mean annual removal of material by gully erosion was about 30 m^3 ha^{-1} $year^{-1}$, where the gradient was 9°, and the maximum removal was about 59 m^3 ha^{-1} $year^{-1}$ on gradients of 14°.

The dependence of gully erosion on the angle of inclination and length of the slope, as shown in Table 77, was observed in several locations. The sudden decrease in the erosive activity of water when the incline flattens out at its lower end is caused by the reduced kinetic energy of the water and on the fact that the water is also at its maximum suspended load capacity.

Figures 119–121 illustrate gully erosion always on the same research plot. Figure 119 shows erosion rills on a 2 to 4° gradient on a pasture lying over easily erodible

Fig. 120. Gully erosion on slopes of 9° mean inclination in loess loam in eastern Slovakia (Czechoslovakia). (Photo D. Zachar.)

material; the rills were caused by water confluence on field roads. Figure 120 shows slope rills formed in furrows ploughed parallel to the line of steepest inclination (the average slope inclination was 9°, and the greatest inclination was 14°). Finally Fig. 121 shows the details of a rill created in a moderate depression on a slope of average inclination 6°. Snow thawing and rain-water participated in the creation of rills at this site.

The influence of the relief on gully erosion combined with the influence of rock and climatic factors, is shown in the next series of figures.

Figure 122 shows the upper part of a valley gully developed in easily erodible material as part of an extending water network. The photograph was taken in the

Fig. 121. Valley gully on loess loam at 6° mean inclination; in the background there is a system of parallel slope rills. (Photo D. Zachar.)

Fig. 122. Valley gully in the central region of the state of Victoria. The gully developed on a pasture after deforestation. (By courtesy of Soil Conservation Authority of Victoria, Australia.)

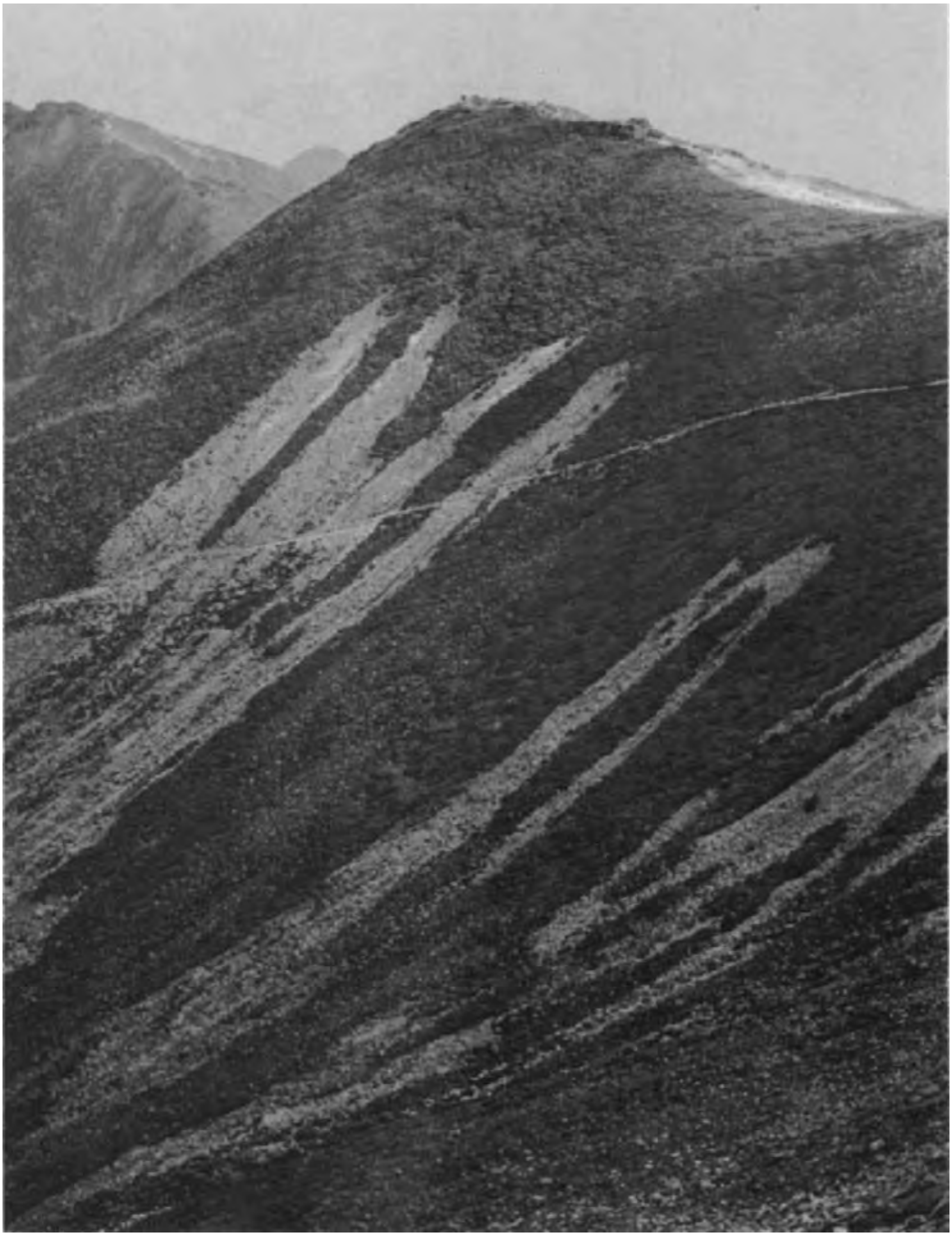


Fig. 123. Shallow ravines caused by erosion and solifluction, growing into gullies in some places (Tatra Mountains, Poland). (Photo D. Zachar.)

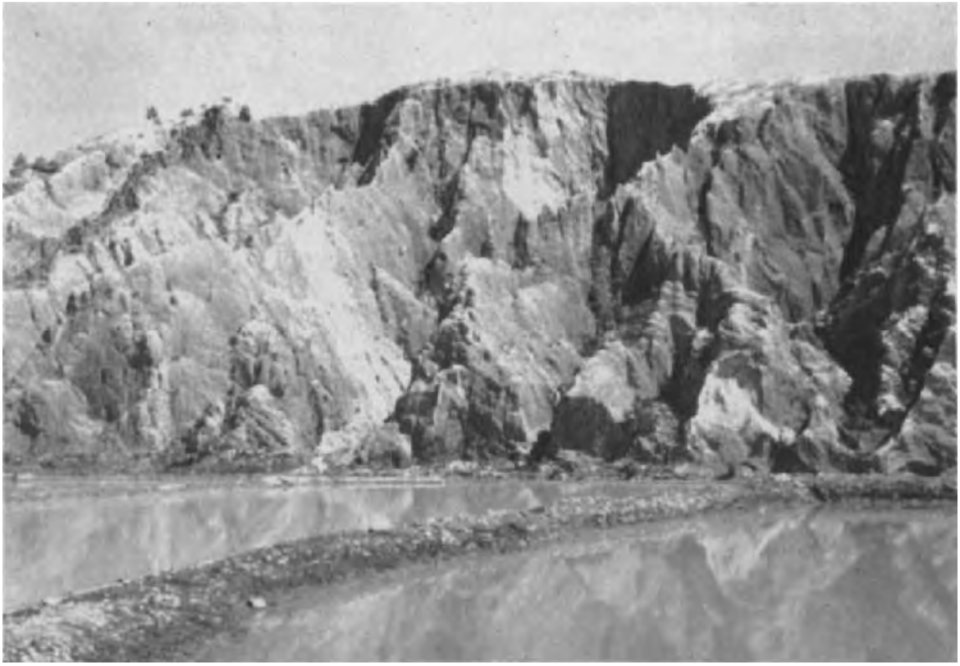


Fig. 124. Gully erosion growing into polymorphous erosion of the badland type by a combination of slipping and sliding of earth, and other destructive phenomena. Basin of the Yellow River in China. (The author's collection of photographs.)

central part of Victoria (Australia), after deforestation and conversion of the forest land into grazing. Figure 123 is a photograph taken from the alpine region of the Tatra Mountains in Poland. It shows the creation of shallow ravines in shallow gravel on a very steep slope.

In the creation of ravines surface runoff plays a certain part, but the main causative factor is the saturation of the surface layers of the soil with water, so that the internal friction of the soil is diminished and solifluction occurs with subsequent creation of gullies. A main contributory factor to this process is mechanical compaction and impairment of soil stability. Figure 124 shows the most intense form of precipitation erosion taking place in loess layers on steep slopes. Here, erosion rills develop into irregular kettle-form ravines causing the complete disintegration and destruction of the slope.

As far as *soil management* is concerned, gully erosion is most likely to occur where the permeability of the soil is low, where the confluence and channelling of surface runoff is high, and where the vegetation is impoverished. Therefore gully growth is encouraged most by ploughing of the soil, intensive grazing, incorrect planning of field and forest roads, ill conceived arrangement of fields in the

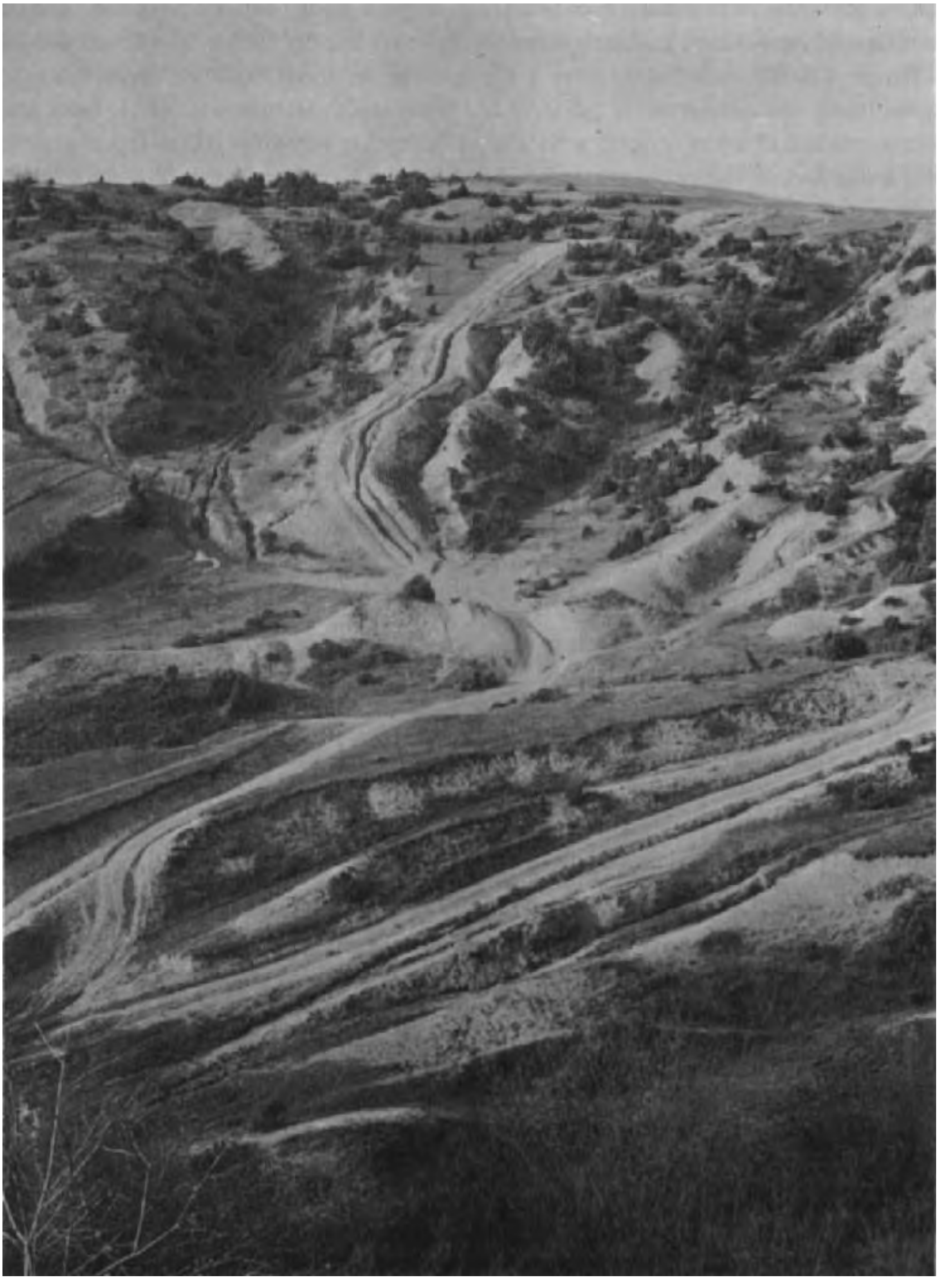


Fig. 125. Gully erosion on the loam-sand soils of central Slovakia (Czechoslovakia) caused by a badly judged lay-out of field roads and land misuse. (Photo D. Zachar.)

landscape (Fig. 125), and bad practices in ploughing, timber logging, surface drainage of rain-water, and irrigation.

Kosov (1970) established that gully growth in the USSR is attributable to agricultural mismanagement (ploughing virgin land, overgrazing, etc.), field and forest roads and areas around settlements, foresting activities (clear-felled areas), and irrigation, in the proportions 75, 15, 5, and 5%, respectively. A similar ratio between the occurrence of "agricultural" and other gullies depending, of course, on the structure of the soil mantle has been observed in other countries, too. The longest gullies occur on roads, skidding lines in forests, and in depressions on agricultural land.

An *overall assessment* of the factors and conditions involved in gully erosion has not been undertaken in as great detail as that achieved in the assessment of sheet erosion. In general, methods for predicting levels of sheet erosion have been used in the literature for the estimation of probable gully erosion also; the latter being assessed in terms of various coefficients. From among the methods that have been used for estimating gully erosion, the author has selected the method based on the flow of suspended load, as used by Kosov et al. (1976) in the mapping of erosion in the Asian part of the USSR.

The danger arising from gully erosion, and *potential gully erosion,* respectively, as a function of the depth of flow of suspended load was derived by Goncharov (1954), using the formula

$$\varrho = \frac{l_1 \varphi}{2{,}200} \; \frac{d}{n} \left(\frac{v_i}{v_k} \right)^{3.33},$$

where ϱ is the water turbidity expressed in terms of the weight of suspended load in unit volume of water [g m^{-3}], φ the turbidity parameter for the watercourse, d the average diameter of the particles of the gully bed, n the coefficient of bed roughness (approximately 0.05), v_i the water velocity, $v_i = AQ^{0.2}\, i^{0.375}/n^{0.75}$, A the morphometric coefficient of the gully bed, the depth in ratio with the width of the watercourse, Q the water discharge rate [l s^{-1}], i the average inclination of the ground in the catchment area ($i = \Delta_h/l_1 = \Delta_h \lambda$), Δ_h the vertical fragmentation, λ the density of gully network (horizontal fragmentation), $\lambda = \Sigma L/F$, L the total length of gullies, the surface area of the catchment area, l_1 the average slope length, v_k the critical water velocity,

$$v_k = (\lg 8.8H/d_5) \times \sqrt{2g/3.5\gamma_0 \times (\gamma_1 - \gamma_0)\, d},$$

H the depth of the watercourse, d_5 the average diameter of the largest grain, the content of which is ca 5%, g the gravitational acceleration, and γ_1, γ_0 are specific weights of earth particles (2,650 kg m^{-3}) and water, respectively.

Thus the calculation requires that the following information is available before the turbidity can be established. Details of relief (slope inclinations), a mechanical analysis of the soil, the water discharge rate, the surface area of the catchment area, and the density of gullies (found from maps). The depth of the watercourse (h) is obtained from the formula

$$h = \frac{Q_0^{0.375}\, n^{375}}{i^{0.185}} .$$

If hydrographic data are available, then the relative contributions of snow thaw and rain-water as factors in gully erosion may be established by this method.

The quantity of suspended load expressed in terms of its discharge rate is obtained from

$$Q_s = Q\,\varrho ,$$

and the specific flow of suspended load from

$$q_s = \frac{Q_s}{F} .$$

The depth, and intensity of gully erosion, respectively, can be derived from

$$E'_g = 315.4 q_s ,$$

where Q is the water discharge rate [l s^{-1}], Q_s the discharge rate of suspended load [l s^{-1}], ϱ the water turbidity [g m^{-3}], q_s the specific flow of suspended load [l s^{-1} km^{-2}], E'_g the gully erosion [m^3 ha^{-1} $year^{-1}$], and 315.4 the number of seconds per year x 10^{-5} (10^{-5} arises from the conversion of 1 km^{-2} into m^3 ha^{-1}).

For the territory investigated, Kosov et al. (1976) established a five-point scale of gully erosion:

Grade of erosion	1	2	3	4	5
Flow of suspended load	<0.001	0.00—0.01	0.01—0.1	0.1—1.0	1.0 +

In the author's scale of erosion intensity the intervals between the various grades of erosion are smaller (see Table 78).

It should be added that in this classification the *transit share of erosion products* in the watercourse may vary greatly. In the previous chapter it was mentioned that in sheet erosion this varies from very small values up to 12% or even 20%. In gully erosion the proportion is substantially higher, and in valley gullies adjoining the hydrographic network a figure of 100% may be reached. In slope gullies the proportion is smaller. Therefore the data given in Table 78, are valid only for catchment areas in which the soil is protected from sheet erosion and the transit share of products of gully erosion reaches 100%.

Table 78. Grading of erosion by the rate of silt discharge

Grade	Specific silt runoff [$l\ s^{-1}\ km^{-2}$]	Rate of soil removal [$m^3\ ha^{-1}\ year^{-1}$]	Mean annual depth of soil removed [$mm\ year^{-1}$]
1	<0.0015	<0.5	0.0125
2	0.0015—0.015	0.5—5	0.275
3	0.015—0.05	5—15	1.0
4	0.05—0.15	15—50	3.5
5	0.15—0.63	50—200	12.5
6	>0.63	>200	50

4.2.8 Tunnel erosion

As mentioned in the section on the classification of erosion, *tunnel erosion* occurs where there is intense penetration of the ground water. In the geographical literature, phenomena caused by tunnel erosion are also referred to as *sham karst* (pseudokarst). This form of erosion occurs both in cold *(thermo-karst)* and warm *(loamy karst)* regions. Sporadic occurrences are also known in regions of abundant precipitation *(clastic karst)*, but it occurs more frequently in regions of low precipitation, and more often on saline soils *(solonetzic karst)* than on non-saline soils. The loamy karst and solonetzic karst are of economic importance.

Tunnel erosion is now known to be much more widespread than was originally supposed, and underground forms of soil disintegration occur in practically any thick layer of finely grained sediment on each of the continents. In most cases it develops into intense gully erosion, and therefore tunnel erosion is sometimes referred to in the literature as a special form of gully erosion. It frequently occurs on forested land causing both soil losses and water loss, thus reducing the economic value of the affected forested area.

Of the various factors governing tunnel erosion, again *climate* (the amount of rainfall in relation to temperature in particular), and the available moisture for plant growth, are all of considerable importance. The type of vegetation that accompanies tunnel erosion occurs mostly in semiarid regions where there is still sufficient precipitation for plant growth, although long dry spells occur in between the rains. The annual precipitation in regions in which tunnel erosion occurs amounts 200 to 750 mm; the average monthly temperature in the dry summer months varies from 25 to 32 °C, and precipitation in the summer comes in the form of violent cloudbursts. Severe desiccation of the soil causes the opening of crevices which gradually enlarge, both in width and in depth (the latter being up to 6 m), so that surface water enters the larger fissures and causes them to grow rapidly. The formation of underground corridors partially or completely diverts the water from the surface, and surface forms of erosion therefore cease. The runoff coefficient of the surface water in catchment areas with loamy karst is usually very low (less than 0.1).

Also of importance in tunnel erosion is the nature of the *soil* and *bedrock*. Loamy karst arises mostly in loess with a distinct hard pan from 30 to 60 cm below the surface (Hosking 1967). From the morphological point of view a number of different types of loamy karst may be distinguished, four of which were described in detail by Lilienberg (1962), who also established relationships between the karst and the relief, the rock structure, and the rate of growth of the karst (Table 79).

In the case of *karst in the bedrock,* the characteristic feature is underground gullying which affects the whole area in such a way, that on the slope surface mounds and depressions of irregular forms appear. The underground erosion network ramifies and on the soil surface angular or elliptical openings appear. The surface forms depend on the rock type.

In *deluvial karst,* tunnel erosion takes a different course, in this case desiccation and salination occur, and sometimes a stratification of earth layers of different permeability becomes conspicuous. The salt content is usually higher than 1%. Tunnelling often occurs after secondary salination of the soil surface and soil substrata. Salination increases the degree of cracking and the leaching of substances from fissures. In this type of karst, erosion is much more rapid than in the previous type and may be of the third or fourth grade.

The most intense tunnel erosion is that of the *terrace type of karst,* which may be found on the borders of terraces in river valleys, and on maritime terraces, etc. A high porosity (25 to 40%) is typical of the rocks in which these forms arise, and this makes possible a rapid penetration of water into the soil with the consequent washing out of the fine particles. Owing to the high potential energy of the relief, underground forms develop into well-expressed surface forms.

The *mud volcano type of karst* is similar to the deluvial type, but is morphologically more varied, and the intensity of erosion is smaller.

In addition to these types of loamy karst, there is a number of other types, their essential common features being the development of surface forms and the high erosion intensity; in underground hollows with steep angles of inclination, the kinetic energy of water flow is higher, and the earth is not protected by vegetation. In some cases of laminar underground wash, suffosive landslides, earth subsidence, and other such phenomena occur. The most expressive form of underground and surface erosion is the *badland.*

Thus a further factor that contributes towards tunnel erosion is a large amount of potential *energy* in the relief, although vertical openings nevertheless develop on the borders of terraces, and ravines form on negligible gradients also. An essential feature of this form of erosion is the considerable drop between the mouths and the

Table 79. Morphogenetic types of loamy karst and their features

Type of karst	Nature of the parent relief	Rocks	Fissures	Angle of inclination of subterranean channels	Length of main channels [m]	Growth rate
Bedrocks	Hill slopes almost without deluvia	Sandy to loamy, cretaceous, seldom conglomerates	Tectonic weathering	10—30°	Large: tens, often hundreds of metres	Slow
Deluvial	Slopes with deep deluvia	Loessial, clayey, sometimes loamy to sandy	Weathering desiccative, dip jointing	30—40°	Medium: tens to hundreds of metres	Medium
Terrace	Terraces and terrace remnants	Loessial loams and gravels	Slope fissures, dip jointing	20—70°	Small: units to tens of metres	Rapid
Volcanic	Volcanic mud	Loamy to stony, volcanic breccia	Desiccative, vertically bulging	10—50°	Small: tens of metres	Medium

outlets of underground corridors. Lilienberg (1962) affirms that 15° is the minimum angle of inclination at which tunnel erosion occurs. According to the author's observations, the length of the main channel may be as much as 1,000 m from the terrace border of the river bank and its deeply cut river-bed. If tunnel erosion occurs on flat plains, it is of low intensity only.

Of the other factors influencing tunnel erosion, the density of *vegetation* is of considerable importance; under natural conditions in arid regions the influence of the poor plant growth is small. Vegetation protects the soil from the ravages of tunnel erosion by holding much of the surface runoff in the surface layers of the soil and preventing its penetration underground. It also protects the soil against salination, desiccation, and crevicing. No tunnel erosion has yet been observed under forest stands. If tunnel erosion is already at an advanced stage of development, control by encouraging plant growth is more difficult and further measures are needed.

Although tunnel erosion is a very undesirable phenomenon, it has not been studied in sufficient detail to allow calculation of its probable extent under different circumstances. Its intensity may be assessed without special investigation, simply according to the proportion of disrupted land surface, which may give an indication of its intensity and stage of development. In very intense tunnel erosion no surface openings can be seen in the first stage, then in the second, third and fourth stages the proportions of disrupted surface area are 25%, 50 to 60%, and 100%, respectively. Land affected by underground erosion should be protected from further expansion of the erosion during the first stage. In the less intense forms, soil disintegration is not so serious, and in weak tunnel erosion, only a negligible part of the land surface (tenths and hundredths of one per cent) is affected.

Forms of tunnel erosion are shown in Figs. 34–40. In the author's research in the CSSR, tunnel erosion was observed on three plots on loess sediments. The size of openings varied from 0.5 to 1.5 m. On one plot tunnel erosion developed into pronounced gully erosion.

4.3 Wind erosion

The main factor in *wind erosion* is the movement and circulation of air. The *wind* affects the soil by desiccating the surface layers, and drying up and removing soil particles by deflation. The stronger the wind, the greater is its influence on the soil. In some localities the soil properties are determined mostly by wind action. The influence of the wind on soil is described under the general term *aeolization*; if soil properties are predominantly affected by wind erosion, then the soil is said to be aeolized. The impoverishment of the soil by the removal of fine material and organic substance, and the burying of soil and crops under blown sand are the economically important consequences of wind erosion.

Soil offers some resistance to air currents, reducing their velocity and diminishing their kinetic energy. The effect of the wind also depends on internal soil properties, particularly the cohesion of soil aggregates as expressed by the proportion of granular material in the soil; soil cohesiveness depends on the amount of cementing substance and the soil moisture content. The overall influence of the wind is thus determined by its *erosivity (aeolisivity)* and the resistance of the soil to the wind *aeolibility*; in the absence of vegetation this overall effect represents the potential wind erosion, which may be defined as the highest possible erosion of soil left unprotected by vegetation.

Accordingly, *potential wind erosion of soil* is given by

$$E_p = VP,$$

where E_p is the potential wind erosion of the soil, V the wind erosivity, and P the soil erodibility factor.

The equation is identical to the equation for the calculation of potential precipitation erosion $E = RK$ (where R is the precipitation erosivity and K is the soil erodibility); V and R both depend on the kinetic energy of the respective erosion factor, and the resistance of the soil to erosion. The principal difference between the erosive effects of water and air arises from the difference in density between the two media

$$\frac{\text{Density of water}}{\text{Density of air}} = \frac{0.99913}{0.00122} = 819\ .$$

Air density, and consequently atmospheric pressure also, decreases with increasing elevation above sea level, and with increasing temperature. Owing to the differences in the physical properties of air and water, the velocity of air currents is tens or hundreds of times greater than the velocity of flowing water, and it may be said that air is constantly in motion, both in a horizontal direction and a vertical direction. In this way soil particles may be carried over long distances.

4.3.1 Erosion force of the wind

The *erosion force of the wind* depends mainly on the velocity and pressure of the wind. The relationship between the wind pressure (q) on a surface perpendicular to its direction and the wind velocity (v) is expressed by

$$q = \frac{\varrho}{2g}\ v^2\ ,$$

where q is the wind pressure [kg m^{-2}], v the wind velocity [m s^{-1}], ϱ the specific weight of the air, and g the acceleration due to gravidity; this depends on the temperature (t) and the barometric pressure (p)

$$\varrho = \frac{1.293}{1 + 0.00367t} \frac{p}{101.3} .$$

For a temperature $t = 15$ °C and a pressure $p = 101.3$ kPa, $q = 0.0625v^2$ kg m^{-2}. Wind velocities and pressure for various Beaufort numbers and an atmospheric pressure of 101.3 kPa, are given in Table 80.

According to some authors, soil particles are carried away even under gentle breeze conditions (BF No. 3); with a moderate breeze (BF No. 4) particles 0.05 cm in diameter (sand) are moved, in a fresh breeze (BF No. 5) sand is lifted into the air, in a strong breeze (BF No. 6) heavy erosion sets in, in a fresh gale (BF No. 8) dust storms arise, and in strong gales (BF No. 9) and whole gales the violence of the dust storms is maximal. In Czechoslovakia, the highest observed velocity of wind squalls in the High Tatras was 240 km h^{-1} at a wind pressure of 300 kg m^2.

Wind velocity increases with height above the ground surface. According to Yakubov (1946), the mean wind velocities in successive layers above the ground are as follows:

Height above the ground [m]	0.05	0.25	0.50	1	2	16	32	123	500
Wind velocity [m s^{-1}]	1.3	2.01	2.44	2.84	3.33	4.69	5.40	8.26	9.25

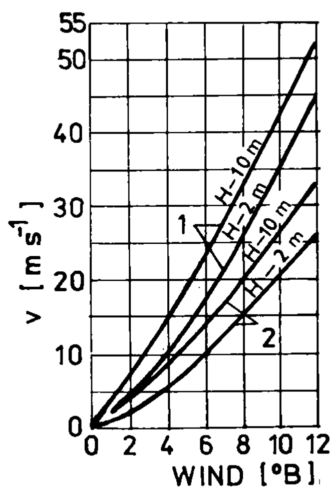


Fig. 126. Mean wind velocities (v) as a function of height above ground level (H) (B – wind intensity according to Beaufort). 1 – velocity of wind squalls, 2 – average wind velocity.

As a guide, the wind velocities at heights of 1, 6, and 16 m, are 2.18, 3.28, and 3.61 times greater, respectively, than the velocity $v_{0.05}$ at 5 cm height. The mean velocities of wind squalls recorded at heights of 2 and 10 m above the ground are shown in Fig. 126 (Zvonkov 1962).

Table 80. Beaufort scale of wind force in 1946 at 40 m height

BF No.	Verbal description	Velocity					Wind pressure [kg m^{-2}]
		Range		Mean			
		[m s^{-1}]	[km h^{-1}]	[m s^{-1}]	[km h^{-1}]	[mile h^{-1}]	
0	Calm	0—0.2	1	0	0	0	0
1	Light air	0.3—1.5	1—5	0.9	3	2	0.05
2	Light breeze	1.6—3.3	6—11	2.4	9	5	0.36
3	Gentle breeze	3.4—5.4	12—19	4.4	16	9	1.2
4	Moderate breeze	5.5—7.9	20—28	6.7	24	13	2.8
5	Fresh breeze	8.0—10.7	29—38	9.3	34	18	5.4
6	Strong breeze	10.8—13.8	39—49	12.3	44	24	9.5
7	Moderate gale	13.9—17.1	50—61	15.5	55	30	15.0
8	Fresh gale	17.2—20.7	62—74	18.9	68	37	22.3
9	Strong gale	20.8—24.4	75—88	22.6	82	44	31.9
10	Whole gale	24.5—28.4	89—102	26.4	96	52	43.6
11	Storm	28.5—32.6	103—117	30.5	110	59	58.1
12	Hurricane	32.7	118	34.8	125	67	75.7

Within the range of naturally occurring wind velocities, as in the case of water velocity, four critical velocities are distinguished: v_{k1} – earth particles begin to move on the soil surface, v_{k2} – the saltation of particles begins, v_{k3} – the particles begin to sedimentate, v_{k4} – the motion of particles stops.

Experimental measurements made by Chepil (1959), Zvonkov (1962), and others, have proved that v_{k1} lies between 3.5 and 4.0 m s^{-1} for soil particles of 0.05 to 0.1 mm diameter.

The general formula for the calculation of the critical wind velocity, according to Velikanov (1948), is

$$\frac{v^2}{gd} = \alpha + \frac{\beta}{d} ,$$

where v is the critical wind velocity [m s^{-1}], g = 9.81 m s^{-2} (the gravitational acceleration), d is the diameter of soil particles beginning to move [m], and α, β are experimentally established constants α = 14, β = 0.006 m.

The wind velocity at 8 m height, v', is 14.88 times the wind velocity at ground level. Substituting $v = v'/14.88$ in the equation, the value of the equivalent wind velocity at 8 m height, at which grains begin to be moved on the ground is obtained

$$v' = 46.5 \sqrt{14d + 0.006} \qquad [\mathrm{m\ s^{-1}}] .$$

Zvonkov (1962) derived the following formulae for the various wind velocities

$$v_{k1} = 1.414 \sqrt{\frac{\lambda_t + \lambda_c}{\lambda_\Phi} (1 \pm \sin\varphi)\ (0.66\,\gamma d + p_0)\ K_3} ,$$

where v_{k1} is the critical wind velocity [m s^{-1}], λ_t the coefficient of friction, λ_c the coefficient of cohesion, λ_Φ the aerodynamic coefficient, λ the coefficient of resistance, φ the angle of inclination of the soil surface, γ the specific weight of soil particles, d the grain diameter, p_0 the atmospheric pressure (1.03 × 10^6 g cm^{-1} s^{-2}), and K_3 the coefficient of protection of soil surface (K_3 = 1 for unprotected land)

$$v_{k2} = 1.414 \sqrt{\frac{\lambda}{\varrho} \frac{1 \pm \sin\varphi}{\mathrm{tg}\,\alpha_1} (0.66\,\gamma d + p_0)\ K_3} ,$$

where v_{k2} is the wind velocity at which soil particles begin to soar;

$$v_{k3} = 1.414 \sqrt{\frac{\lambda_m}{\lambda_\Phi^\varrho} \frac{1 \pm \sin\varphi}{\mathrm{tg}\,\alpha} (0.66\,\gamma d + p_0)} ,$$

where v_{k3} is the wind velocity at which sedimentation of particles begins;

$$v_{k4} = 1.414 \sqrt{\frac{\lambda_t}{\lambda_\Phi^\varrho} (1 \pm \sin\varphi)\ (0.66\,\gamma d + p_0)} ,$$

where v_{k4} is the wind velocity at which particles are deposited.

When the ratio $v_{k4} : v_{k1} = 0.66$ then $\lambda_t/(t \pm \lambda_c) \doteq 0.44$, the coefficient of friction $\lambda_t = 0.44\lambda_\Phi\lambda$, and aerodynamic coefficient $\lambda_\Phi = 9.35(\gamma d/\varrho v^2 c)$. The absolute values of the coefficients are shown in Fig. 127. The critical wind

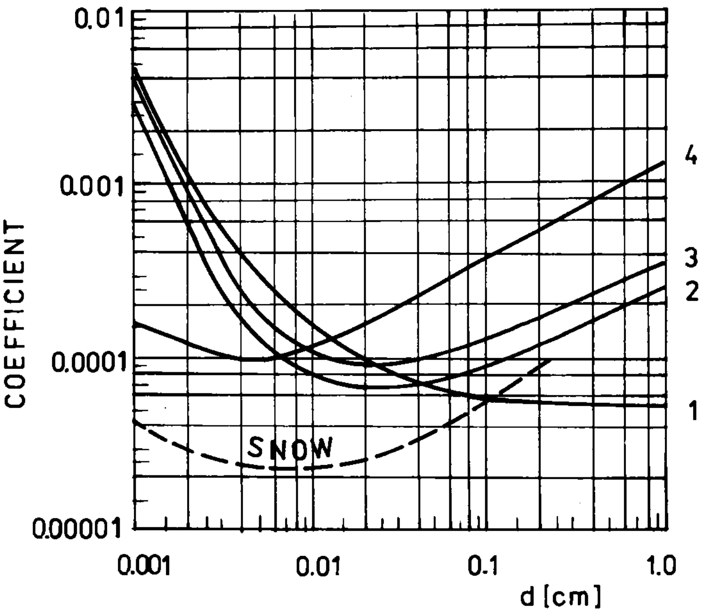


Fig. 127. Variation of coefficients with soil grain diameter. 1 – streamlining (λ_Φ), 2 – friction (λ_t), 3 – cohesion (λ_c), 4 – resistance (λ) for $H = 1$ m (d – grain diameter in cm).

Fig. 128. Critical wind velocities (v_k) in relation to the diameter of soil grains for $H = 1$ m. 1 – v_{k1}, 2 – v_{k2}, 3 – v_{k3}, 4 – v_{k4} (AB – straight line for grains of $d < 0.03$ cm).

velocities v_{k1}, v_{k2}, v_{k3}, and v_{k4} in relation to the particle diameter, d, are given in Fig. 128. It should be noted that v_{k4} for grains of diameter $d < 0.03$ cm decreases according to the straight line AB.

Some data on critical wind velocities in relation to average grain sizes, are given in Table 81.

Table 81. Wind velocity v_{k1} and v_{k2} [m s^{-1}]

Critical wind velocity	Grain diameter, d [mm]					
	0.01	0.1	0.25	1.0	1.5	2.0
v_{k1}, grains begin to move	3.65	3.83	4.57	6.62	7.65	8.57
v_{k2}, grains lifted up	3.72	5.41	6.60	10.71	13.41	16.25

The distances of transportation of soil particles in wind erosion are observed to vary; grains of diameter greater than 1.0 mm being carried only a few metres, those of 0.125 to 1.0 mm diameter being transported over distances of 1 to 1.5 km, those of 0.0625 to 0.125 mm diameter for several km, those of 0.0165 to 0.0624 mm diameter for 300 to 1,500 km, and those of less than 0.0156 mm diameter moving indefinite distances.

The *coefficient of soil resistance* to wind erosion may be expressed by the relationship

$$K_e = \frac{P_v}{R},$$

where K_e is the coefficient of soil resistance, P_v the kinetic energy of the wind, and R the resistance force of the soil.

The kinetic energy of the wind

$$P_v = 0.392\lambda_{\Phi}\varrho_t\, d^2v^2,$$

where P_v is the kinetic energy of the wind [g cm^2 s^{-2}], ϱ_t the density of the air [g m^{-3}] for t = 0°C and atmospheric pressure = 101.3 kPa, ϱ = 0.001293 [g cm^{-3}], d the diameter of soil particles [cm], v the wind velocity [cm s^{-1}].

Resistance force of the soil for v_{k1} and v_{k2}

$$R = (\lambda_t + \lambda_c)(1 \pm \sin\varphi)(P_t + P_a)K_3,$$

where R is the resistance force of the soil, P_t the gravitation force due to mass soil particles ($P_t = 0.525\gamma d^3$), P_a the pressure ($P_a = 0.785d^2\varrho_0$), ϱ_0 the atmospheric pressure (1.03×10^6 g cm^{-1} s^{-2}).

Substituting for P_t and P_a

$$R = 0.785(\lambda_t + \lambda_c)(1 \pm \sin\varphi)(0.66\gamma d + \varrho_0)d^2K_3.$$

After adjustment

$$K_e = \frac{0.5\varrho v^2}{\lambda(1 \pm \sin\varphi)(0.66\,\gamma d + p_0)K_3}.$$

If $K_e < 1.0$, earth particles are not moved by the wind. When $K_e = 1$, and $P_v = R$, earth particles vibrate on the point of lateral movement over the soil surface, and when $K'_e > 1.0$ movement occurs. Particles begin to soar when $K'_e > 1.0$, where

$$K'_e = \frac{0.5 \text{ tg } \alpha_1 \varrho v^2}{\lambda (1 \pm \sin \varphi)(0.66 \gamma d + p_0) K_3} .$$

The larger the values of K_e and K'_e, respectively, the more intense is wind erosion.

Finally, the amount of blown and deposited material in wind erosion will depend on h_m, the height up to which material is transported, and v_m, the maximum wind velocity involved in this transport; $h_m = (v_m - v_{k1})^2/2g$, and the transportation distance (the horizontal trajectory of the particles), l, for grains of different diameter is $h_m/\text{tg } \alpha$. For example, when $v = 18$ m s^{-1}, $v_{k1} = 6.9$, $v_{k2} = 8.7$, $v_{k3} = 5.65$, $v_{k4} = 4.47$ m s^{-1}, $\alpha_1 = 32.5°$, $\alpha_2 = 12.5°$, the transportation distances, l, of grains of diameter $d = 0.58$, 0.2, and 0.1 mm, are 25, 35, and 44 m, respectively.

The *amount of blown* and *deposited material*, and the *amount of material transported* in the shifting of sand (as in *barkhans* and *dunes*) is given by

$$G = gm = 0.5 h_m b v_t,$$

where G is the amount of transported material [m^3 s^{-1} m^{-1}], g the volume of sand dune [m^3]; $g = 0{,}5 h_m l b$ [m^3], m the rate of creation of barkhans or dunes over a given distance, h_m the greatest height at which particles are transported [m], b the relative width of sand drift [m], v_t the transit velocity of moving particles [m s^{-1}]; $v_t = v - v_k$, and l the horizontal projection of sand dune; $l = (1/\text{tg } \alpha_1 + 1/\text{tg } \alpha_2) h_m$.

For calculating wind erosion the following figures are assumed: $d = 0.058$ cm (d for sand grains is 0.02–0.03 cm), $\gamma = 2{,}000$ g cm^2 s^{-2}, $\lambda = 0.00029$, $\lambda_\Phi = 0.64$, $\lambda_t = 0.000078$, $\lambda_c = 0.00018$, $g = 9.81$ m s^{-2}, $\varrho = 0.00123$ g cm^{-3}, $K_3 = 1.0$, $\varphi_1 = 0°$, $p_0 = 1{,}013{,}000$ g cm^{-1} s^{-2}. Accordingly, the kinetic energy of the wind is

$$P_v = 0.392 \lambda_\Phi \varrho d^2 v^2 = 0.392 \times 0.64 \times 0.00123 \times 0.058^2 \times 500^2 = 0.26 \text{ g cm}^2 \text{ s}^{-2},$$

when $v = 10$ m s^{-1}, $P_v = 1.04$; when $v = 15$ m s^{-1}, $P_v = 2.34$; when $v = 20$ m s^{-1}, $P_v = 4.16$, etc.

The gravitation force acting on soil particles in the air is

$$P_t = 0.525 \gamma d^3 = 0.525 \times 2{,}000 \times 0.058^3 = 0.205 \text{ g cm}^2 \text{ s}^{-2}.$$

The resistance force of the soil with respect to the wind velocities v_{k1} and v_{k2}, is

$R = 0.785(\lambda_t + \lambda_c)(1 + \sin\varphi)(0.66\gamma d + p_0)\, d^2 K_3 = 0.785\,(0.000078 + 0.000108)$ $1.0\,(0.66 \times 2{,}000 \times 0.058 + 1{,}013{,}000) \times 0.058^2 \times 1.0 = 0.5$ g cm^2 s^{-2}.

The critical wind velocity, v_{k1}

$$= 1.414\sqrt{\frac{\lambda}{\varrho}\ (1 \pm \sin\varphi)\,(0.66\,\gamma d + p_0)\,K_3}$$

$$= 1.414\sqrt{\frac{0.00029}{0.00123}\ 1.0\,(0.66 \times 2{,}000 \times 0.058 + 1{,}013{,}000) \times 1.0} = 6.91 \text{ m s}^{-1}.$$

The critical wind velocity, v_{k2}

$$= 1.414\sqrt{\frac{\lambda}{\varrho}\ \frac{1 \pm \sin\varphi}{\operatorname{tg}\alpha_1}\ (0.66\,\gamma d + p_0)\,K_3}$$

$$= 1.414\sqrt{\frac{0.00029}{0.00123}\ \frac{1.0}{0.637}\ (0.66 \times 2{,}000 \times 0.058 + 1{,}013{,}000) \times 1.0} = 8.7 \text{ m s}^{-1}$$

The critical wind velocities v_{k3} and v_{k4} are, according to the above-mentioned formulae

$$v_{k3} = 5.65 \text{ m s}^{-1}, \text{ and } v_{k4} = 4.47 \text{ m s}^{-1}.$$

The velocity of the free fall of particles, v_s

$$= 1.16\sqrt{\frac{\gamma d}{\lambda_{\Phi}\varrho}} = 1.16\sqrt{\frac{2{,}000 \times 0.058}{0.64 \times 0.00123}} = 4.45 \text{ m s}^{-1}$$

(for $d = 0.01$, 0.1, and 1.0 cm, $v_s = 1.17$, 6.15, and 20.9 m s^{-1}, respectively).

The greatest altitude of particle flight, h_m

$$= \frac{(v_m - v_{k2})^2}{29} = \frac{(12 - 8.7)^2}{19.62} = 0.56 \text{ m};$$

(for $v_m = 15$ m s^{-1}, $h_m = 2.02$ m, etc.).

The coefficient of soil resistance, K_e

$$= \frac{P_v}{P} = \frac{0.5\,\varrho v^2}{\lambda\,(1 \pm \sin\varphi)\,(0.66\,\gamma d + p_0)\,K_3}.$$

$$= \frac{0.5 \times 0.00123 \times 6.9^2}{0.0002 \times 1.0\,(0.66 \times 2{,}000 \times 0.058 + 1{,}013{,}000)\,1.0} = 1.0$$

(for $v = 0.5$, 8.7, 15, 20, and 30 m s^{-1}, $K_e = 0.52$, 1.40, 4. 65, 8.30, and 18.7, respectively).

The amount of transported material G

$$E = 0.5h_m bv_t = 0.5 \times 4.4 \times 1.0 \times 12.0 = 26.4\ \mathrm{m^3\ s^{-1}}.$$

The *amount of material removed from the soil surface*

$$E = \frac{h\Delta_x v_t \,.\, 10^4}{100}\, t = 100\, h\, \Delta_x t \quad [\mathrm{m^3\ ha^{-1}}]\ ,$$

at an altitude of soil transfer of 0.0026 m, and a distance of transfer Δ_x, derived from v_t (velocity of particle transfer) = 12 m s^{-1} and t (the time of transfer) = 60 s,

$$E = 100h\Delta_x v_t t = 100 \times 0.0026 \times 12 \times 60 = 187.2\ \mathrm{m^3\ ha^{-1}}.$$

Thus the intensity of wind erosion of the soil under the given conditions is 187 m^3 ha^{-1}. The conditions represent a fresh gale (BF No. 8) lasting for one minute in an area of very easily erodible soil. Since some of the transported particles are simultaneously deposited in the same locality, the total soil loss is a consequence of the movement of particles over longer distances, and this depends mainly on the granular composition of the eroded soil.

According to the above-mentioned relationship between grain size and critical velocity, it may be supposed that particles and microaggregates with a diameter of less than 0.01 mm (these being dispersed over the soil surface), about half of the particles of diameter 0.01 to 0.1 mm, a quarter of the particles of diameter 0.1 to 0.5 mm, and a very small fraction of those of diameters greater than 1.0 mm would be moved away from the endangered field. In deserts, grains of diameter 0.2 to 0.3 mm prevail, but these are transported mostly in layers near the ground at low velocity, and they are deposited after moving short distances, when the velocity decreases. Therefore the total soil losses from a particular area will be substantially lower.

4.3.2 Soil resistance to deflation

As mentioned in the previous chapter, *soil resistance* depends mainly on the *granular composition (texture), moisture content,* and *surface roughness of the soil.* All these properties are mutually interrelated and interact with one another. Among them, the granular composition is of prime importance, as has been clearly demonstrated in the context of critical wind velocities.

Extensive research has shown that *sand* of uniform grain size representing the result of the selective influence of the wind on the soil, is the least resistant to wind erosion. On account of the continuous blowing away of the finer particles, the particulate fraction of the surface layer diminishes, and consequently soil erosion decreases also. Thus wind-eroded soil loses its fine particle content, and sand has an exclusively coarse-grained structure. Finally, soils which have evolved on finely

Table 82. Soil resistance to wind erosion and v_{kl} critical wind velocity

Resistance category	Predominant grain diameter [mm]	Wind velocity, v_{kl} [m s^{-1}]	Erodibility
1	0.1—0.15	3—4	Very high
2	0.05—0.1 and 0.15—0.5	4—5.5	High
3	0.01—0.05 and 0.5—1.0	5.5—7	Moderate
4	0.005—0.01 and 1.0—2.0	7—10	Low
5	Under 0.005 and above 2.0	Above 10	Very low

textured deposits dropped a long distance from the site of origin (loess) have a high proportion of loam and clay material, and are fairly resistant to further wind erosion; these soils, however, are more susceptible to water erosion, whereas sand is less easily eroded by precipitation water.

According to Chepil (1945) and other authors, soil containing particles of diameter 0.1 to 0.15 mm is the most easily eroded. The second category of less erodible grains is formed by the fractions of diameter 0.05 to 0.1 mm, and 0.15 to 0.5 mm, respectively, and the third category is represented by grains of diameter 0.5 to 1.0 mm. Particles with diameters less than 0.01 mm and above 1.0 mm are moved by the wind with difficulty, and only in small quantities. Critical wind velocities for the different soil categories vary, as shown in Table 82.

Chepil (1958), suggested that as a standard for comparison, soil with a 60% content of particles exceeding 0.84 mm in diameter, should be chosen. Such air-dry particles are hardly moved by wind action. However soils with a 60% content of particles larger than 0.84 mm in diameter are by no means common, and it is therefore difficult consistently to use them as a standard. It would be more convenient to use as an etalon sandy soils of aeolian origin which have similar properties under different geographical conditions.

Research was undertaken on this basis by Stredňanský (1977) with the purpose to investigate the frequency of wind occurrence between 1971 and 1975 on soils of different grain composition and moisture content at three sites in southern Slovakia. Rates of soil removal were determined also, in an aerodynamic tunnel. Thus soil erodibility relative to that of sandy soil was established, and its dependence on soil moisture content and granulation was quantified (Table 83).

Table 83. Erodibility of soils of different structures and moisture contents

Type of soil	1	2	1	2	1	2
Sandy	5—8	89.4	10—13	10.3	16—19	2.5
Loamy to sandy	7—9	85.9	9—13	11.6	13—15	2.5
Loamy	9—12	63.6	19—21	27.8	28—31	8.6

1 — per cent of soil moisture content, 2 — erodibility.

Similar results pertaining to the relationships between granulation, moisture content and soil erodibility were obtained by Pasák (1967), who observed that the greatest influence of soil moisture on erodibility occurred in loamy-sand soils. According to Pasák, soil erodibility depends mainly on the content of "non-erodible" particles and on the relative moisture level, as expressed by the relation

$$E' = 22.02 - 0.72P - 1.69V + 2.64R,$$

where E' is the soil erodibility [g m^{-2}], P the content of "non-erodible" particles in the soil, V the relative moisture content of the soil (see below), and R the wind velocity at the soil surface [m s^{-1}].

"Non-erodible" particles refer to fractions with a diameter exceeding 0.8 mm. The relative soil moisture content is given by $V = V_m - V_n$, where V_m is the instantaneous soil moisture content, and V_n the matrical or bound moisture content of the soil, expressed by Solnár in terms of the size of the first soil fraction

$$V_n = \frac{\text{\% of the first fraction}}{2.4}.$$

Soil moisture levels are expressed as percentages of the dry earth weight. As can be seen, soil erodibility is determined in this equation basically by the content of barely erodible particles – those of diameter greater than 0.8 mm and less than 0.01 mm. As shown in Table 82, soils with a predominance of such material are of medium to very low erodibility.

Chepil (1958), in making calculations of the threshold wind velocity, takes the soil moisture content into account as expressed by the relation $c = W/v$, where W is the instantaneous moisture content, and v the number of hygroscopicity.

Erodibility is a function of the cohesive forces between soil particles surrounded by a film of adsorbed water. The soil moisture content changes in direct proportion with the amount of precipitation and inversely with the square of the temperature, since the temperature influences the rate of evaporation from the soil.

Erodibility is influenced not only by soil moisture content, but also by another variable property of the soil, namely its structure, particularly the proportion of water-soluble aggregates. Soil structure is expressed mainly in terms of the proportions of particles of diameter less than 0.02 mm and the humus content. The mechanical soil stability of four types of soil [according to Chepil (1958)] is given in Table 84.

When soil is wetted, some of the cementing substance which gives the soil its cohesiveness and thus its resistance to erosion obviously becomes dispersed throughout the aggregates. Aggregates of low resistance are disintegrated by the mechanical action of wind-blown particles, so that gradually, with progressive wind erosion, the intensity of the erosion increases as more and more aggregates disintegrate.

Table 84. Relationship between mechanical stability of soil surface and proportion of particles of d <0.02 mm dislodged by water

Soil. 4 types	Soil material	Particles above 0.02 diameter [%]	Mechanical soil stability [%]
Sandy to clayey	Deposited	7.3	50.8
	Original	13.3	65.3

Chepil expresses this *acceleration of erosion* in terms of the abrasion coefficient, which is associated with the amount of soil material derived from the erosion of the soil aggregates

$$K_a = a\left(\frac{25}{v}\right)^2,$$

where K_a is the abrasion coefficient of the soil aggregates, a the weight of abraded soil, and v the wind velocity [mile h^{-1}].

Interesting data pertaining to the relationship between the size of the aggregates and soil erodibility (expressed in terms of the threshold velocity) were obtained by Gossen (in Zaïtseva 1970) in an investigation of carbonate chernozems. He established that for soil aggregate of diameter less than 1 mm, v_{k1} ranges from 3.8 to 6.6 m s^{-1}, and if the diameter increases to 2 mm, v_{k1} increases to 11.2 m s^{-1} (Table 85). Gossen concludes from this observation that soil resistance to wind erosion rapidly increases when aggregates of more than 1 mm diameter predominate, although erosion may continue in the presence of aggregates exceeding 6 mm in diameter, owing to the lower specific weight of these larger aggregates.

Corresponding to these data, a direct relationship between the sizes of aggregates in the surface layers of the soil and soil erodibility was discovered by Shiyatyĭ

Table 85. Values of wind velocity v_{k1} at 15 cm height above the ground for different sizes of soil aggregates and moisture contents

Diameter aggregates [mm]	Wind velocity [m s^{-1}]	Moisture content of aggregates [%]
0.25	3.8	6.1
0.25—0.5	5.3	7.4
0.5—1.0	6.6	7.6
1—2	11.2	6.5
2—3	13.1	7.0
3—5	17.6	6.8

(1965a, b). With a 60% content of particles of diameter more than 1 mm, the soil is almost totally resistant to wind erosion, even at a wind velocity of 12.5 m s^{-1} (Fig. 129). A diameter of 1 mm is close to the value obtained by Chepil ($d = 0.84$ mm).

Almost all authors who have investigated soil erodibility have come to the conclusion that the structure of the surface layers up to 5 cm depth is a reliable pointer to soil erodibility. If the content of particles of diameter more than 1 mm (Pasák gives 0.8 mm and Chepil gives 0.84 mm) exceeds 60%, the soil has an adequate resistance to wind erosion, and does not require protection; if the proportion of this material varies between 50 and 60% the soil is less resistant, and if the quantity is less than 50%, the soil needs to be protected against wind erosion.

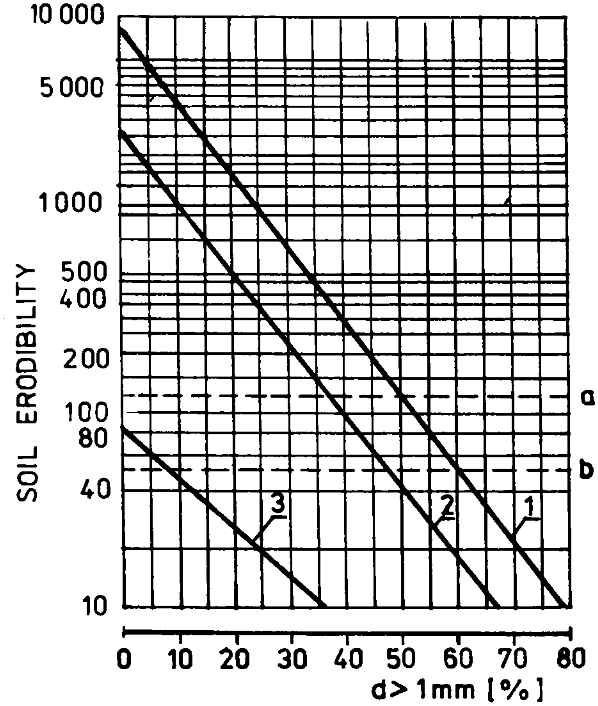


Fig. 129. Variation of soil erodibility (e, g m^{-2}) in relation to soil structure (content of particles with $d < 1$ mm, %) and wind velocity (v) at height of 50 cm, 1 – $v = 12.48$ m s^{-1}, 2 – $v = 10.4$ m s^{-1}, 3 – $v = 8.05$ m s^{-1} (a – highest permissible erodibility, b – permissible erodibility).

In the course of making a complex assessment of soil properties in wind-endangered regions of the USSR, Dolgilevich et al. (1973) came to the conclusion that the intensity of wind erosion depends on the soil resistance, which can be estimated by means of a number of diagnostic characteristics.

Among the most important of the latter is the content of erodible aggregates (A,%), which are carried away during 30 minutes of wind blowing with a velocity of 20 to 24 m s^{-1}. Other diagnostic characters are the content of loamy-type particles (Ca), the content of clay (Si), the content of microaggregates of diameter less than

0.01 m *(a)*, and the total quantity of Ca^{2+} + Mg^{2+} per 100 g of soil. Relations according to these signs are as follows: $A = 87.8 - 0.915$ Ca; $A = 72.7 - 1.043$ Si; $\lg A = 1.9 - 0.015a$; $A = 1.97 - 0.008$ $Ca^{\cdot\cdot} + Mg^{\cdot\cdot}$. The mechanical resistance of aggregates may also serve as a diagnostic character (Table 86).

The validity of these characters as indicators of soil resistance was tested by measuring soil erodibility in an aerodynamic tunnel (Table 87). About 40% of the soils in the USSR have other, more important characteristic properties, and cannot be assessed by the above diagnostic characters alone; these soils are, for example, several carbonate and saline soil types, humus-carbonate soils, soils of high humus content, etc. Some of these other properties are manifested by the amount of

Table 86. Soil erodibility in the steppe zone of West Siberia and North Kazakhstan

Erodibility	Diagnostic parameters					
	A [%]	Ca [%]	Si [%]	a [%]	$Ca^{2+} + Mg^{2+}$ mekv. 100 g^{-1}	Mechanical stability of aggregates [kg $aggreg.^{-1}$]
Very low	<20	>65	>28	>28	>35	>0.80
Low	21—33	64—53	>28	>28	>35	0.79—0.50
Moderate	34—49	52—40	27—20	27—14	35	0.49—0.30
High	50—73	39—2.1	19—3	13—3	34—14	0.29—0.15
Very high	<74	<20	<2	<2	<13	<0.14

Table 87. Classification of West Siberian and North Kazakhstan soils by degree of wind erodibility

Soil group	Erodibility* [t ha^{-1}]	Erodibility
Solonetz and southern chernozems	0.2	Very low
Dark chestnut and solonetz types and southern sandy to loamy chernozems	0.3—0.7	Low
Dark chestnut and solonetz type southern chernozems	0.8—2.0	Moderate
Southern loamy to sandy and leached chernozems	2.1—5.0	High
Chestnut solonetz types, loamy to sandy chernozems, etc.	5.1	Very high

*Soil removal in 30 min at wind velocity from v_{kl} to 20—24 m s^{-1}.

eroded aggregates present. However the content of carbonate, humus, etc. is affected by frost and alternate freezing and thawing of the soil, which is of great importance in alpine sites and cold regions. Regelation and freezing of the soil with the formation of ice crystals results in serious soil blow off in winter and spring.

In order to standardize the classification of soil erodibility given in Table 86, may be noted that on very easily erodible soils (dark chestnut loams to sandy soils), soil losses occurring during dust storms amount to 480 t ha^{-1} and wind velocities attain values of 11 to 27 m s^{-1}. The time during which the velocity, v_{k1}, is reached varies from 0.4 to 69.0 hours in this region. Occurring once every four to five years, these dust storms cause an average annual loss of 96 to 120 t ha^{-1}. If the losses caused by wind erosion in the periods intervening between dust storms are added to this value, it turns out that in terms of the author's classification, the intensity of wind erosion in this region is very considerable indeed, and is even catastrophic.

In the course of research on soil erodibility in the Ukraine, Dolgilevich (1973) found that the wind erodibility of loess soil increased with the $CaCO_3$, Ca, and Mg contents; the content of erodible aggregates *A* was smaller than that of other soils. Differences in soil erodibility for various particle sizes of the fine earth were found to be as much as 42-fold.

Of the many methods of assessing soil resistance, that involving a comprehensive soil factor which expresses soil resistance to wind erosion, as recommended by Shiyatyĭ (1972), is worthy of mention. According to Shiyatyĭ, resistance to erosion may be obtained from the mechanical structure of the soil, using the formula

$$S = a + bx_1 - cx_2 - dx_3,$$

where *a, b, c, d* are regression coefficients, x_1 is the clay content (particle size less than 0.001 mm), x_2 the sand content particles of diameter 0.05 to 0.25 mm, and x_3 the content of particles greater than 0.25 mm diameter.

For the investigated soils, the following coefficients were obtained

$$S_1 = 34.7 + 0.9x_1 - 0.3x_2 - 0.4x_3,$$

and according to this equation, Shiyatyĭ classified soils into six categories as follows:

Scale of soil susceptibility	I	II	III	IV	V	VI
Index of soil resistance, *S*	Above 65	55—65	45—55	30—45	15—30	Below 15

The contribution of the carbonate content to soil resistance to wind erosion was expressed by

$$S_2 = 20 + 5K$$

where *K* is the carbonate content of the soil.

The total soil resistance to wind erosion was taken as

$$S = S_1 + S_2,$$

where S_1 represents mechanical resistance factors, and S_2 represents chemical resistance of the soil.

Most methods of assessing soil resistance to potential wind erosion rely on a knowledge of the ratio of erodible to non-erodible particles, erodible particles being taken as those with diameters from 0.01 to 1.0 mm.

4.3.3 Common assessment of climate and soil

The previous chapters have shown that with respect to wind erosion, there is a close relationship between climate and soil, which may be expressed in general form by

$$E_v = KP,$$

where E_v is the wind erosion, K a factor of the climate, and P a factor of soil resistance.

The climatic factor is expressed in most formulae by the wind factor, V, which is basically a function of the kinetic energy and the lifting force of the wind. An additional factor of the climate is its effect on soil dryness, or degree of wetting, which influences the properties of the surface layers of the soil.

Chepil et al. (1962) established that the removal of soil particles in wind erosion depends mainly on wind velocity and the soil moisture content according to the formula

$$q = \mathrm{f}\,\frac{v^3}{w^2}\,,$$

where q is the soil removal by wind erosion, v the wind velocity at 10-m altitude, and w the effective soil moisture content, computed as follows

$$w = \mathrm{f}\;\frac{P - E}{T^2}\;,$$

where P is the amount of precipitation [mm], E the amount of evaporation [mm], and T the average annual temperature [°C]. The climatic factor of wind erosion was expressed by Chepil and his collaborators as

$$C = 100\;\frac{v^3}{(P - E)^2}\;.$$

Pasák (1978) used the following formula for the calculation of the climatic factor, C, for conditions in the CSSR

$$C = 100(6 + 0.52n)^3\,(I_z + 60)^{-2},$$

where C is the climatic factor of wind erosion, n the frequency of wind $\geq$ BF No. 5 (expressed as a percentage of all winds in a year), and I_z the index of climatic humidity, according to Konček

$$I_z = \frac{R}{z} + \Delta r - 10t - (30 + v^2),$$

where R is the total precipitation in the growth season (April to September), r the positive deviation of the total winter precipitation from the value 105 mm, and t the mean wind velocity at 14 hours throughout the growth period [m s^{-1}].

This method gives results identical with Thornthwait's moisture index which was used by Chepil. In the USSR, the hydrothermal coefficient HTK = $\Sigma R/\Sigma t \times 10$ is used in the calculation of the coefficient of climatic humidity, where ΣR is the total precipitation for the period [mm], and Σt is the total temperature for the period in which the average temperature exceeds 10°C.

According to those parameters, the climate can be divided in a number of ways into various types. By the formula $w = (P - E)/T^2$, the climate is more rigorously assessed than with the formula $w_s = (R - R'/t$, and Konček's formula $I_z = R/2 + \Delta r - 10t - (30 + v^2)$ is still more precise. Thus for a region, in which the annual precipitation, R, is 600 mm, the annual average temperature, t, is 10°C, and the annual evaporation, E, is 500 mm, the following result is obtained: $w = (600 - 500)/10^2 = 1$; $w_s = (600 - 510/10 = 9$. According to Konček's criteria, the region is moderately dry, with a value of I_z around 20. The hydrothermal coefficient according to the formula HTK $\Sigma R/\Sigma t \times 10$ varies between 1.0 and 1.1.

For the identification of macroregions of the USSR endangered by different degree of wind erosion, Kosov et al. (1976) used a relatively simple method which can be applied in other instances also. In this method, soils were divided into two regional types: I. soils resistant to wind erosion occurring where winds of velocity greater than 10 m s^{-1} are common (loam and clay soils), II. soils of low resistance to erosion occurring where there are winds of velocity greater than 6 m s^{-1}. The territory was then divided into types of region, according to the wind force: for the first group of soils, the three categories of wind velocity were 6 to 9, 10 to 15, and more than 15 m s^{-1}, respectively, and for the second group of soils they were 10 to 15 and more than 15 m s^{-1}, respectively. By multiplying wind velocity by the percentage probability of its occurrence, values of up to 510 for the first soil category and up to 209 for the second soil category were obtained. On the basis of these parameters of wind force, v_s, the territory was divided into four categories with respect to the degree of danger from wind erosion:

Intensity of erosion	Wind force level
1 Slight erosion	<50
2 Moderate erosion	50—100
3 Severe erosion	100—200
4 Very severe erosion	>200

The authors divided the territory into five categories, according to the moisture parameter HTK: I. > 1.33, II. 1.33−1.0, III. 1.0−0.77, IV. 0.77−0.33, V. < 0.33. The greater the value of the moisture parameter, the more intense is the level of

Table 88. Intensity of wind erosion in West Siberia, Kazakhstan and the central Asia regions of the USSR

Geographical zone	Wind erosion parameter		
	v_s	HTK	Grade of erosion
Tundra and forest tundra	428—510	>1.33	4
Taiga	100—350	1.0—1.33	1—3
Forest steppe	<50—209	0.77—1.0	1—4
Steppe	110—377	0.33—1.0	2—4
Semidesert and desert	204—419	<0.33	3—4

wind erosion. Wind erosion data for the various zones of the USSR are given in Table 88.

It therefore appears that potential wind erosion in the tundra regions is very high in spite of the high moisture levels, and with the lower wind levels in the forest tundra, taiga, and steppe zones, the potential erosion may be no higher than the first grade; in the forest steppe, however, it quickly rises with increasing dryness. Differences in the levels of actual erosion are much greater because in humid regions the soil is well protected by vegetation, whereas in deserts actual erosion is close to potential erosion.

4.3.4 Complex methods of assessment of wind erosion

Besides the methods for determining erosion intensity on the basis of climatic and soil factors, methods which take a whole complex of factors into account have also been developed.

One of the first experimental equations for the calculation of the *wind erosion intensity* (Chepil and Woodruff 1954) took into account soil "cloddiness", surface roughness, and the *amount of plant residue* in the soil. The equation was intended for the calculation of erosion on the Great Plains (North America)

$$W_e = 491.3 \frac{I}{(RK)^{0.835}} ,$$

where W_e is the soil removal by wind erosion [t acre^{-1}], I the soils "cloddiness" factor, R the crop residue factor, and K the ridge roughness equivalent factor.

The factor K, for unploughed fields is 1.5, for wheat stubble 3.2, for tilled land with deep furrows 10.0 (without crop residues), for pastures 1.0, and for forested land 0.1 to 0.5. The factor R varies from 100 to 300 for soils without crop residues, from 300 to 600 for soils with average amount of crop residues (wheat stubble, maize roots, etc.), and from 600 to 1,000 for soils with abundant crop residues. The

R value is expressed as the mass of crop residue in pounds per acre, and for perennial stands may attain higher values than those quoted above. Finally, the factor *I*, is expressed in terms of the content of particles of diameter greater than 0.84 mm.

The authors constructed a nomogram according to this equation, and classified areas endangered by wind erosion into three categories according to the intensity of the erosion: I. erosion of up to 0.25 t $acre^{-1}$ (0.6 t ha^{-1}), II. erosion of 0.25 to 5.0 t $acre^{-1}$ (0.6–12.4 t ha^{-1}), III. erosion of over 5.0 t $acre^{-1}$ (12.4 t ha^{-1}). According to the author's scale of wind erosion, this covers a range from the first to the third grade of erosion intensity.

This equation was later adapted by Chepil (1959) for calculating wind erosion on cultivated land

$$E = IRKFBWD,$$

where *I* is the soil "cloddiness" factor, *R* the crop residue factor, *K* the ridge roughness equivalent factor, *F* the soil "arability" factor, *B* the windbreak factor, *W* the field width factor, and *D* the wind direction factor.

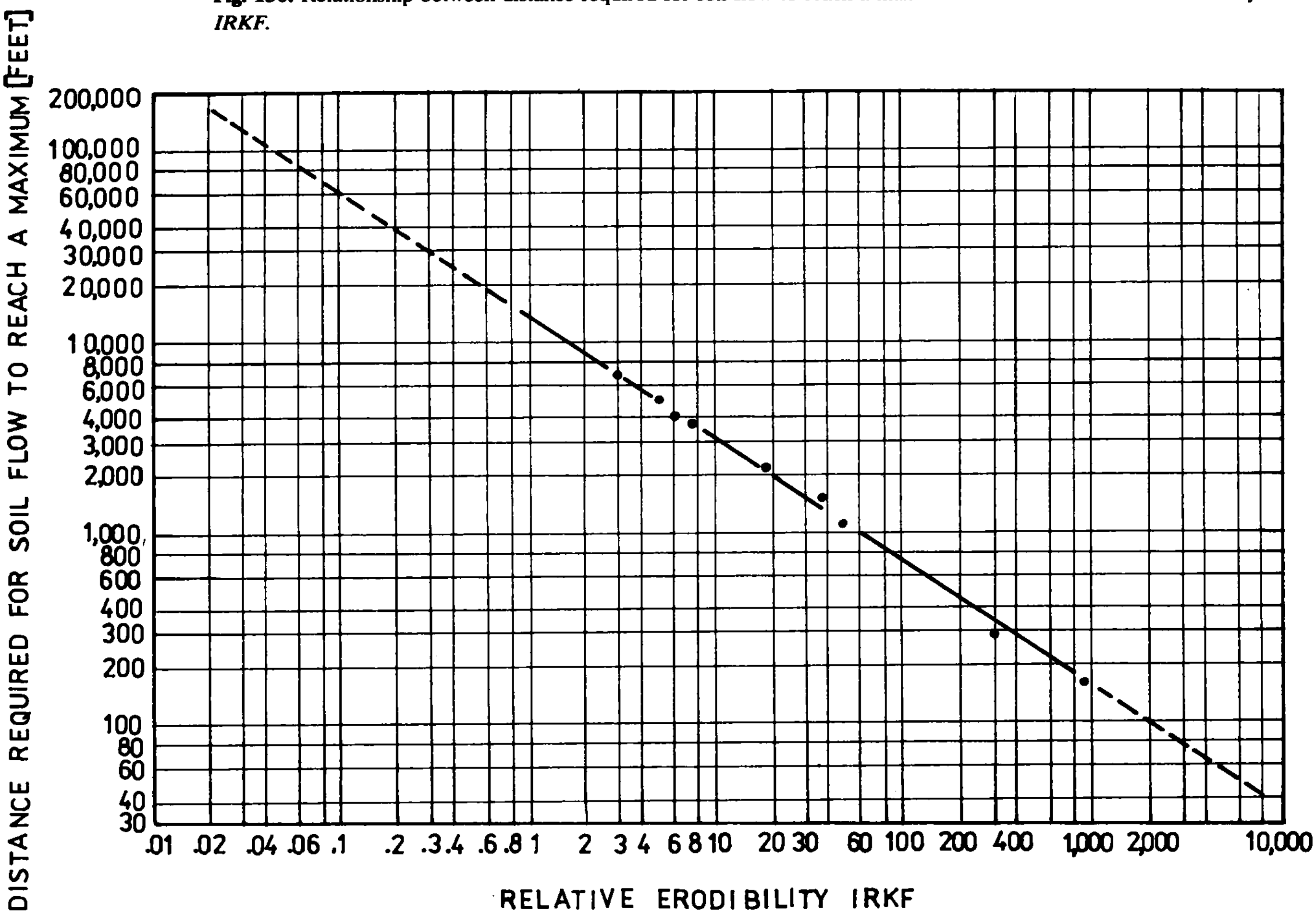
In this equation Chepil places much importance on the fifth factor, *B*, which expresses the effect of the width of the eroded field on wind erosion. He assumes that owing to bombardment from wind carried particles, the abrasion of aggregates and the shifting of loosened particles increase with increasing width of the field. The increase in the rate of soil movement with distance downwind across an unsheltered wind erosion area is called *soil avalanching*. Its recognition has contributed to the adoption of various forms of strip cropping as a means of wind erosion control.

The wind velocity and altitude on which this analysis is based are 40 m.p.h. at 50 feet, respectively, above a smooth, level, and unsheltered terrain. Such a wind occurs in the region approximately once every two years in April, and lasts about six hours.

The factors *I* and *F* are expressions of soil erodibility, and the *R* and *K* factors also have a bearing on soil erodibility; on the basis of already known relationships, these factors together represent field erodibility. An alignment chart and table have been drawn up from which the independent influences of factors *I, R, K,* and *F* may be determined (Chepil 1958); it is also explained by Chepil how procedures may be reversed in order to determine what values of *I, R, K,* and *F* are necessary to reduce the level of erosion to any specified degree. The relationship between the wind erodibility of the soil and the *distance* required for soil movement to reach a maximum rate is shown in Fig. 130.

Figure 130 shows that the field erodibility values, $F_e(IRKF)$, are in this case up to 1,500 times greater in range than values D_m, the maximum width of the field, these

Fig. 130. Relationship between distance required for soil flow to reach a maximum and the relative wind erodibility of the soil, *IRKF*.



two ranges of values being in inverse relationship; for about 100 m D_m one part of $F_e = 1$ m (1 foot = 0.3048 m).

From the maximum width of the field, D_m, Chepil deducts d – that width of the field, which is fully sheltered from wind erosion by some barrier such as *stubble, growing crop, hedge,* or *tree windbreak.* This is based on Woodruff's and Zingg's (1952) formula, in which

$$d = \frac{h}{17} \quad \frac{v_0}{v} \quad ,$$

where d is the width of area fully protected from wind erosion, h the height of barrier, v_0 the minimum wind velocity at a height of 50 feet required to move the most easily erodible soil fraction, and v the actual velocity at a height of 50 feet.

The minimum velocity required to initiate soil movement on a smooth, bare surface after erosion has already started and before wetting by rain and subsequent surface crusting take place, is about 21.5 m.p.h. at 50 feet above this surface. Under these conditions, wind erosion occurs (according to Chepil) in the distance

$$d = \frac{365h}{v} \,,$$

where d is the distance of full protection from erosion caused by a wind velocity of 40 m.p.h. at a heigth of 50 feet, h the height of barrier, and v the actual velocity at a height of 50 feet.

The *wind barrier factor, B,* determined by this method may be used in the equation only if the barrier can be considered as being permanent, or if the width of the field, D_m, always changes in relation to the wind break barrier. The width unprotected from the wind, d_t, may be determined by the formula

$$d_t = D_m - d.$$

Example: if $h = 2$ feet, $RKF = 400$, $D_m = 300$ feet, $v = 40$ m.p.h. at 50 feet, $d = 365 \times 2 + 40 = 18.25$ feet, and $d_t = 300 - 18.25 = 281.75$ feet.

However, if D_m is given as 300 feet and the intention is to protect the entire width of the field from wind erosion, i.e. $D_m = d$, then the barrier height, h, is computed from

$$h = \frac{D_m v}{365} \,.$$

In this case the height of the barrier is $h = 300 \times 40 + 365 = 32.88$ feet $\doteq$ 10 m. This means that a 10 m high barrier is sufficient to protect the field from wind erosion at a width of 92 m.

An evaluation of factors W and D may be made from the chart shown in Fig. 131. According to the nomogram, wind erosion as expressed by the formula *IRKFBWD* is determined from the *IRKF* value, which is shown on the left side of the diagram in Fig. 131. From the *IRKF* value one projects along the thick lines

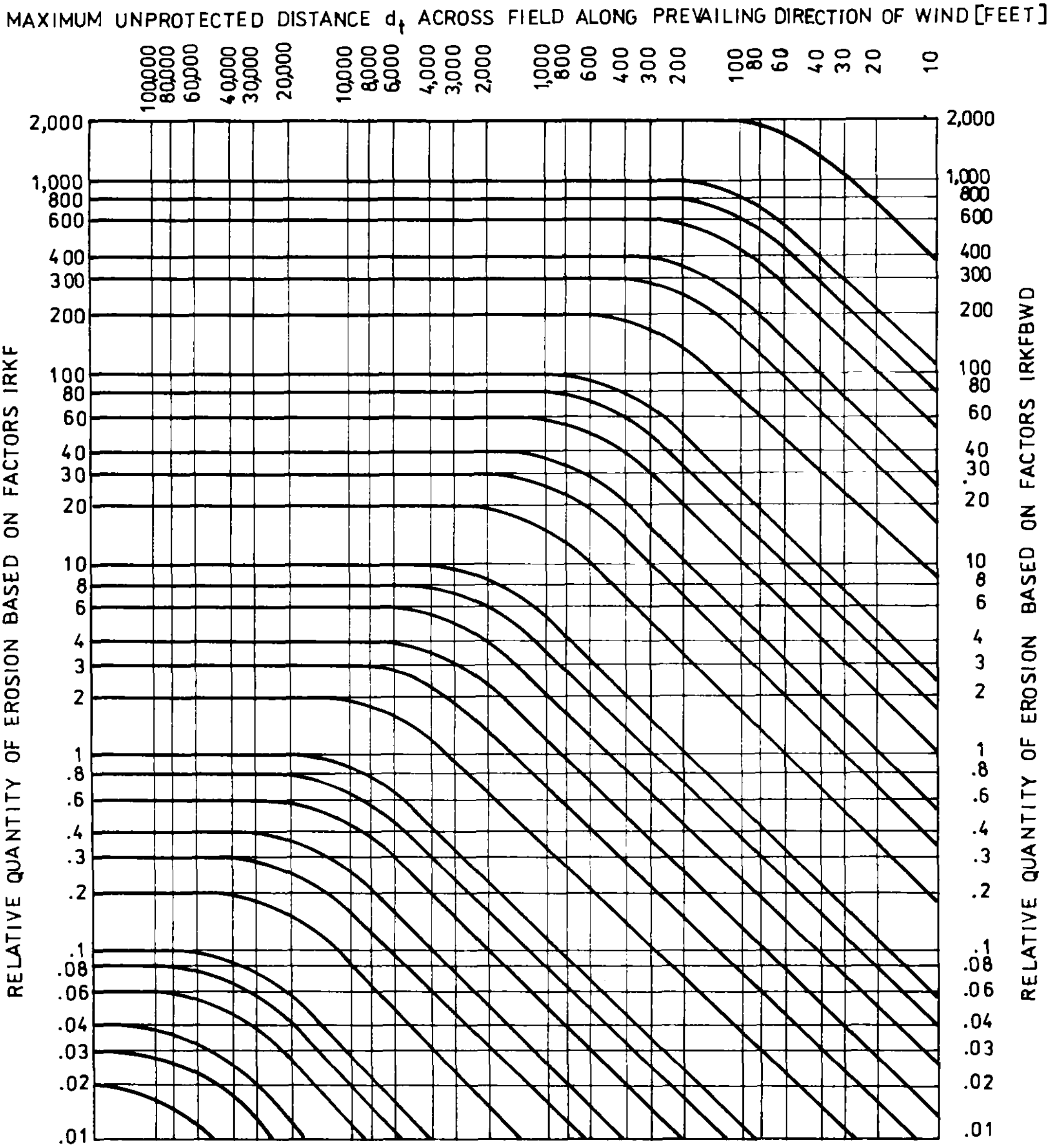


Fig. 131. Diagram showing the relationship between the relative degree of wind erosion, and maximum unprotected distance (d_t) across field in prevailing wind direction.

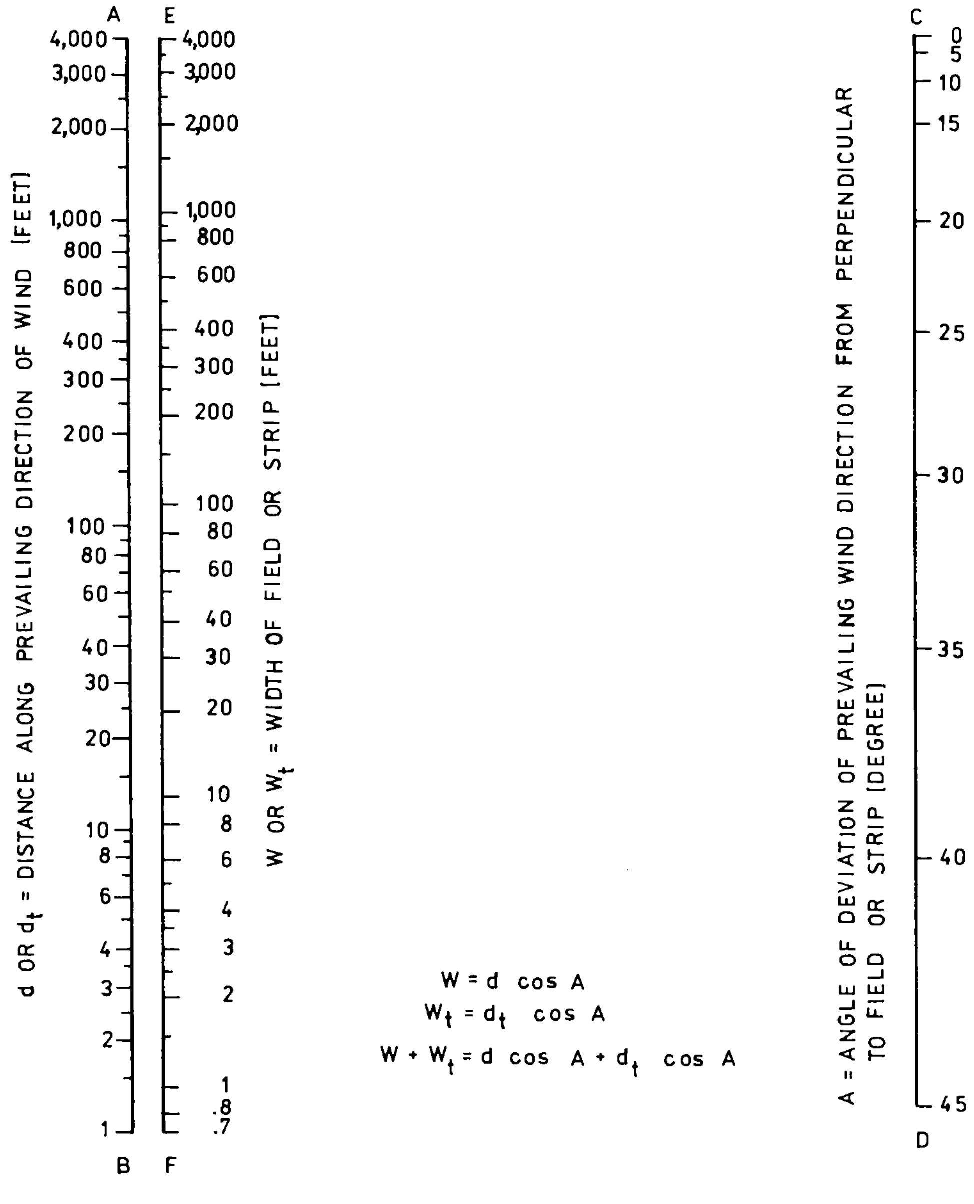


Fig. 132. Alignment chart for determining protection distance across field parallel to prevailing wind direction, from the width of the field and the direction of wind.

until the required value of d_t, shown on the upper side of the diagram is reached. The procedure may be reversed, and the amount of erosion on the unprotected part of the field or within a certain distance of the barrier from the place where erosion reaches its maximum may be computed.

Example: if width of field $W = 1{,}000$ feet, stubble height $h = 1$ foot, $IRKF = 2$, then $d_t = 1{,}100 - 8 = 1{,}092$ feet. From Fig. 131 it is found that the relative erodibility for $d_t = 1{,}092$ feet is 0.42. In order to diminish the erodibility from $IRKF = 2$ to $IRFKBWD = 0.25$, the width d_t must be reduced to 750 feet. If, for the other reason, a reduction in the width of the field is not feasible, it is then necessary to decrease erosion by increasing soil "cloddiness" *(I)*, the content of crop residue in the soil *(R)*, the surface roughness *(K)*, or the barrier height *(B)*.

Finally, the wind direction factor, D, can be determined from the nomogram shown in Fig. 132. If the width of the field, d and d_t (AB connecting line), respectively, and the angle of the direction of the prevailing wind (CD connecting line) are known, the width of field W and W_t, respectively, may easily be found on the EF connecting line. E.g., for a wind direction angle of 22.5 degrees and a field width $W = 1{,}000$ feet, the distance $d = 1{,}100$ feet.

Chepil's equation examines first and foremost the increase in wind erosion with increasing width of the field up to a certain maximum at which erosion is constant, and the effect of a *windbreak*, or *shelterbelt*, on the level of wind erosion. The examination of these relationships makes it possible to determine the widths necessary to reduce wind erosion to an admissible level in areas supporting crops offering different resistances to wind erosion, and in acutely threatened areas; the effect of the distribution of permanent shelterbelts can also be predetermined. Since these are basic issues in the control of wind erosion, the method used in the USSR for the calculation of the necessary distances between shelterbelts, is also discussed here.

According to research carried out by Dolgilevich and his colleagues (Dolgilevich et al. 1973) shelterbelts with an air current permeability of 30 to 40% provide the best *protection against wind erosion*. As a result of the positive effects observed on the yields of agricultural crops, a distance between shelterbelts $L = 30$ is being used, where H is the height of the shelterbelt in m. After an evaluation of the effects of 225 shelterbelts it was concluded that the distance between shelterbelts bears a close relationship to the ratio between the maximum wind velocity (frequency of occurrence, $n = 20\%$, velocity at 10 m height, $v = 11$ to $27\ \mathrm{m\,s^{-1}}$) and the wind velocity at which tolerable erosion occurs. This ratio can be expressed

$$L = \mathrm{f}\,\frac{v_t}{v_m}\,,$$

where L is the distance between shelterbelts expressed by H, i.e. shelterbelt height [m], v_t the tolerable wind velocity [$\mathrm{m\ s^{-1}}$], and v_m the maximum wind velocity [$\mathrm{m\ s^{-1}}$].

For a soil surface roughness parameter of 0.7 cm

$$\lg L = 1.2\,\frac{v_t}{v_m} + 0.58\,.$$

Research has shown that for a wind velocity $v_m = 21$ m s^{-1} and wind duration $t = 20$ h year^{-1}, the amount of erosion per hour is $E_h = 9.4$ t ha^{-1} h^{-1}, and the overall intensity of erosion in this case is $E = E_h t = 9.4 \times 20 = 188$ t ha^{-1} year^{-1}. The level of permissible erosion is $E_t = 2.0$ t ha^{-1} year^{-1}, and the corresponding permissible wind velocity is $v_t = 11.5$ m s^{-1} at 10 m height. The ratio v_t/v_m is 0.55 and the distance L (from lg $L = 1.2(11.5/21.0) + 0.58$) is equal to $17H$. The distance between shelterbelts where the intensity of wind erosion is between 2.0 and 188 t ha^{-1} year^{-1} varies from $17H$ to $47H$, and for a shelterbelt height of 20 m, the distance L is within the range 340 to 940 m, although in practice the range 500 to 600 m can be taken. For erosion of intensity 8 to 28 t ha^{-1} year^{-1}, $L \doteq 28H = 540$ m. At these lower levels of erosion, reduction to a permissible level may be achieved by other measures also.

As an exemple of the positive effect of shelterbelts in erosion control in Kazakhstan, some data may be cited on the reduction of wind erosion for distances between shelterbelts of 260 to 1,000 m, with shelterbelt heights of 50 to 70 m over land tilled into coarse furrows. The rate of soil removal [according to Astakhov (1973)] was below the limit of permissible erosion (Table 89) even with a distance between shelterbelts of 400 to 500 m.

Table 89. Soil removal by wind in the Stavropol region in 1970

Distance between shelterbelts [m]	1,000	750	400—500	260—290
Fine earth removal [m^3 ha^{-1}]	65	49	1.5	0

Of other methods, that have been used, calculations of the intensity of wind erosion on the basis of ten-year measurements made in the neighbourhood of Belgrade may be mentioned (Gavrilović 1972). His experimental equation for the calculation of wind erosion was

$$W_e = T I_v D_e Y X_a F,$$

where W_e is the annual wind erosion [m^3 year^{-1}], T the temperature coefficient ($T = (t^0/10) + 0.1$) [t is the annual mean temperature (°C)], I_v the mean annual wind velocity [m s^{-1}], D_e the mean annual number of windy days in the period without snow cover, Y the coefficient of soil resistance, X_a the coefficient of the structure of the catchment area, and F the area of the catchment area [km^2], up to 300 km^2.

Fot the Y coefficient, Gavrilović gives a table in which values vary from 2.0 for sand, to 0.25 for the most resistant soil. The coefficient X_a varies from 1.0 for barren land, to 0.05 for forested land. For tilled and barren land X_a varied between 0.9 and 1.0.

According to this equation, if $t = 10°C$, $I_v = 2\ m\ s^{-1}$, $D_e = 100$ days, $Y = 2.0$, $X_a = 1.0$, and $F = 0.01\ km^2$ (1 ha); then $W_e = 1.1 \times 2.0 \times 100 \times 2.0 \times 1.0 \times 0.01 = 2.24\ m^3\ ha^{-1}\ year^{-1}$. The number of windy days has the greatest weighting in the equation, since a lower limit for the wind velocity of a "windy day" is not specified. It may be noted that wind erosion also occurs during the winter months.

The most widely used equation is, of course, that which was devised by the Agricultural Research Service of the US Department of Agriculture and the Kansas Agricultural Experiment Station with the cooperation of W. S. Chepil, P. N. Woodruff, F. H. Sidoway, and others. This Wind Erosion Equation has the following form

$$E = IKCLV,$$

where E is the annual loss of soil [t acre^{-1}], I the soil resistance, K the ridge roughness equivalent factor, C the climatic factor, L the field width, and V the vegetation factor.

The equation is described by Hayes (1965), Woodruff and Sidoway (1965), Skidmore et al. (1970), and others. According to Hayes, soils are classified with respect to wind erodibility into eight groups, the content of non-erodible fractions (A fractions) being of the greatest importance for the assignment of a soil to the Wind Erodibility Group (WEG). The variation of soil erodibility in relation to the latter factor is shown in Table 90.

Table 90. Erodibility of standard soil

	Percentage content of 0.84 mm fraction							
A factor	0	10	20	30	40	50	60	>60
t acre^{-1} year^{-1}	(313)	134	98	76	56	38	21	<21
t ha^{-1} year^{-1}	(773)	331	242	188	138	94	57	<57

The erodibility for 0% content of fraction A is the value derived by extrapolation of the curve. In the same way it is possible to extrapolate the curve in the opposite direction to obtain the erodibility for $A = 70\%$, which is thus found to be 5 t acre^{-1} year^{-1} (12 t ha^{-1} year^{-1}). Soils with no non-erodible fraction are rarely found in nature, the lowest values being around 3%, with an erodibility of 220 t acre^{-1} year^{-1} (544 t ha^{-1} year^{-1}).

The factor K takes account of the resistance to wind erosion caused by ridges of given heights and spacings compared with a standard ridge spacing ratio of 1 : 4. (If ridges running at right angles to the prevailing wind direction are 6 inches high and spaced 30 inches apart, their spacing ratio is 1 : 5.) In this way, the value K_r is defined (in inches) by

$$K_{\mathrm{I}} = \frac{\text{standard spacing ratio (1 : 4)}}{\text{field measured ratio (1 : x)}} \text{ height of ridge} .$$

The climatic factor, C, depends on wind velocity and the difference between precipitation, P, and evaporation, E, C being expressed as a percentage. For standard conditions in Kansas, $C = 5\%$.

The factor L is determined according to the nomogram shown in Fig. 131.

Thus wind erosion on a tilled, 100 m wide field of high erodibility, unprotected by vegetation and without organic matter left in the soil is

$$E = IKCL = 0.2 \times 5 \times 0.52 = 172 \text{ t acre}^{-1} \text{ year}^{-1}.$$

This level of erosion is that, which occurs at 100 m distance from the border of the field, whereas at the border, the intensity of erosion amounts to about 50 t acre^{-1} year^{-1}, the average intensity over the field (according to this method of calculation) is about 110 t acre^{-1} year^{-1}.

A detailed study of the effect of wind direction on wind erosion was made by Chepil et al. (1964) for conditions in the Great Plains (North America). The relative level of erosion behind the barrier is calculated from the formula

$$E_{\mathrm{p}} = \frac{10\, l_{\mathrm{p}}}{X} ,$$

where E_{p} is the relative level of wind erosion of the soil, l_{p} the longest line of the erosion compass card when dividing the compass card into eight segments, and X products and each of the resulting quotients.

Thus total length of lines of the wind erosion compass card

$$l = X(FV^3)\,(\Sigma FV^3)^{-1}$$

where V is the wind velocity, and F the first multiplying per cent duration.

The width of the fully protected part of the field, D_{b}, is derived from the total width of the field, D_{f}, where $D_{\mathrm{f}} = W_{\mathrm{f}}/\cos\alpha$

$$D_{\mathrm{b}} = \frac{W_{\mathrm{b}}}{\cos\alpha} .$$

The width of the unprotected part of the field, L, is obtained from $D_{\mathrm{f}} - D_{\mathrm{b}}$, or $L = W_{\mathrm{f}}/\sec\alpha - W_{\mathrm{b}}\cos\alpha$, respectively.

Example: if $D_{\mathrm{f}} = 525$ m, $W_{\mathrm{f}} = 500$ m, $W_{\mathrm{b}} = 190$ m, and $D_{\mathrm{b}} = 200$ m, $D_{\mathrm{f}} - D_{\mathrm{b}} = 525 - 200 = 325$ m.

In addition to the field width the angle of *inclination of the ground* (slope) is also taken into account in the calculation of the intensity of wind erosion. Some authors believe that wind erosion is most acute on a plain, where the angle of inclination is 0°. In reality, as the steepness of the slope increases, the intensity of wind erosion increases on the windward side and on the top of the ridge; this increase is observed to occur up to a certain limit, which is close to 45°. At this point wind erosion by rubbing decreases, and the turbulent erosion of material which is so typical on rock walls begins.

The increased intensity of wind erosion on steeper slopes may be explained by the fact that above the ridge the air velocity and air pressure both increase. In this context one may speak of the drag velocity, V_d [cm s^{-1}], and the surface drag, τ [dyne cm^{-2}], which is equal to τV_d^2, where ϱ is the density of the air [g cm^{-3}].

Chepil et al. (1964) derived relationships between V_d, τ and the relative wind erosion for different slope inclinations (Table 91).

Table 91. Wind erosion in relation to slope inclination

Wind erosion parameter	Slope inclination [%]			
	0	3	6	10
Drag velocity [m.p.h.]	1.06	1.15	1.34	1.55
Surface drag force [dyne cm^{-2}]	2.7	3.2	4.3	5.7
Relative soil loss at the tops of knolls [%]	100	150	320	660
Relative soil loss at the windward slope [%]	100	130	230	370
Angle tan φ	24	25.6	27.5	29.75

Relative soil losses on the tops of knolls, $I = \tau^{2.5}$, losses at the windward slope of knolls, $I_s = (\tau/\tan\varphi)^{2.5}$, and general losses with respect to slope inclination as a factor on which potential wind erosion depends, can be obtained from the formula

$$I = as^b + (cd)^{-1},$$

where I is the relative wind erosion as a function of slope inclination and the equivalent degree of erosion on a plain [%], s the knoll slope [%], and a, b, c, d are constants depending on the state of the soil and other conditions.

In the erosion of the windward of knolls, the removal of soil is concentrated in the upper third of the slope. Relative wind erosion is shown in Fig. 133.

Stredňanský (1977) investigated the relationship between the angle of inclination of soil samples in a wind tunnel and the relative erodibility of the soil, and found smaller values for wind erosion than those given in Fig. 133.

Inclination of soil sample	0°	2°	4°	6°	8°	10°	15°
Relative rate of soil removal [%]	100	147	185	226	296	372	705

Increased wind erosion on the protruding parts of the relief is observed not only on a meso-scale, but also on a micro- and nano-scale on the one hand, and on a macro-scale on the other. Windiness increases with elevation above the sea level and this enhances the erosion force of the wind. Katabatic winds also exert a considerable erosive influence, and bring about a higher intensity of wind erosion on the affected territory. As an example, the occurrence of force 6 and force 8 winds in various parts of the Carpathian Mountains in Czechoslovakia may be quoted:

Station, relief	Number of days in the year with wind force	
	6	8
Hurbanovo, plain	80	10
Sliač, basin-shaped valley	28	5
Bratislava, site with katabatic wind	101	20
Skalnaté Pleso, massif	186	100

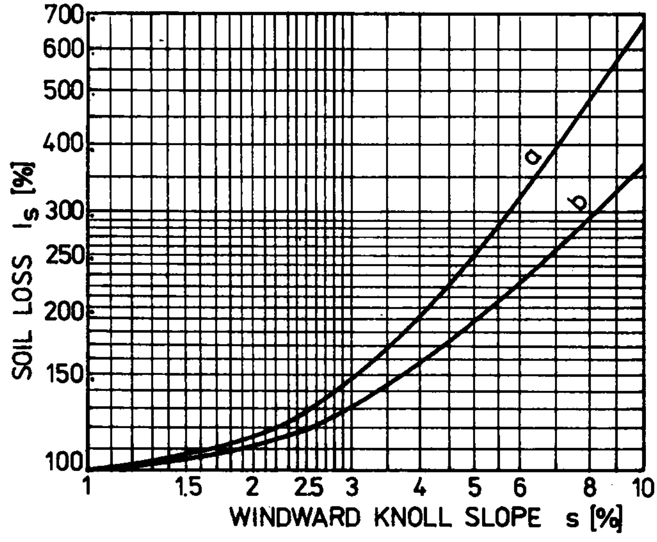


Fig. 133. Soil loss (l_s,%) on undulating terrain relative to that on level terrain, as a function of windward slope inclination of knoll (s, %). a – top of knoll, b – windward slope of knoll.

High winds prevail in the mountain regions and at the foot of the slopes of the Little Carpathians where foehn (warm and dry) winds occur, the observed frequencies being 27, 18, 13, and 4 days per year for winds of force 8, force 9, force 10, and force 11, respectively.

It can be seen from the above example that the relief has a considerable effect on the wind, and therefore on wind erosion also; the relief refers not only to dissection of the topography and the steepness of slopes, but also to the lengths and aspects of slopes. Thus it is possible to speak of ventoerosive orographic systems which are specific for each aeolian relief.

4.3.5 The assessment of wind erosion and evaluation of quantitative data

In the previous chapters procedures for the assessment of the various factors in wind erosion as they have been used in different countries were given. A comprehensive procedure for general application has not yet been developed, since not enough data are available on the intensity of erosion under known natural conditions, and because various authors have different sources of basic data at their disposal. Therefore, no proposal for a generally acceptable comprehensive procedure can be made here, but perhaps a few remarks will suffice.

The first requirement in assessing the intensity of wind erosion is a precise knowledge of the erosion force of the wind, and this depends mainly on the wind velocity or the energy of the wind. From wind energy, *wind erosivity* is established which expresses the ability of the wind to erode the soil, and is otherwise known as the aeolian erosion factor, or ventoerosion factor. The kinetic energy of the wind is established from meteorological measurements, from which the average velocity of winds of a particular frequency of occurrence (according to the purpose of the calculation) is obtained. The kinetic energy is computed from these values, according to the relation

$$W_k = 2.25 v_{0.2}^2,$$

where W_k is the kinetic energy [MJ ha^{-1} h^{-1}] of winds with a 0.2 frequency of occurrence, derived from the formula $W_k = 0.0625 v^2 \times 3{,}600 \times 10^4$, $v_{0.2}$ the velocity of winds with a 0.2 frequency of occurrence [m s^{-1}].

Wind erosion for a given velocity $v_{0.2}$, is determined in aerodynamic tunnels, by the use of instruments as described in the section of methods, or by direct measurement. As an illustration, values given by Dolgilevich et al. (1973) for soils of different resistance are listed in Table 92.

If direct quantitative data are not available, soil erodibility may be assessed by the use of the diagnostic parameters given in Table 86 and the accompanying text.

Table 92. Soil erodibility [t ha^{-1} h^{-1}] in relation to wind velocity

Soil	Wind velocity [m s^{-1}]			
	11	15	19	23
Heavy clay solonetz	0.05	0.05	0.06	0.06
Loamy, salty, dark chestnut	0.0	0.13	0.20	0.34
Loamy chernozem	0.44	1.00	1.60	2.14
Carbonate-loam chernozem	0.10	0.50	2.00	3.90
Sandy chernozem	1.50	3.40	5.80	8.00
Sandy dark chestnut	2.40	7.00	14.00	21.20

Table 93. Soil erodibility [t ha^{-1} h^{-1}] for wind velocity 20 to 25 m s^{-1} and *A* fraction of erodible particles

Soil erodibility category	1	2	3	4	5	6
Fraction A content [%]	<10	10—20	20—33	33—50	50—80	>80
Erodibility [t ha^{-1} h^{-1}]	<0.1	0.1—0.5	0.5—1.5	1.5—5	5—15	>15

These parameters, such as the proportion of erodible soil particles or the content of fraction *A* aggregates, provide a measure of soil erodibility. Table 93 gives values of soil erodibility, for different fraction *A* contents at $v = 20-25$ m s^{-1}.

The value of *A* may be established directly, or may be derived from another diagnostic parameter, such as the content of clay, loam, micro-aggregates, carbonate, salt, humus, etc. Interconversions between some of these factors are given in Table 86 and the corresponding text.

In assessing the effect of erosion and the wind erosivity, it is necessary to find out how much soil is eroded by wind with a kinetic energy of 100 MJ ha^{-1} year^{-1}; in assessing soil erodibility, it is necessary to know how much kinetic wind energy is needed to remove 1 ton of soil from an area of 1 ha in 1 hour. The wind energy is computed from velocity readings at 10 m height above the ground.

In establishing soil erodibility in tunnels, the tunnel wind velocity is multiplied by approximately two (according to tunnel construction); the erosive effect of the wind is expressed in terms of the weight of soil removed [g] per 10 J m^{-2} h^{-1}, and soil erodibility is expressed in J per 1 g eroded soil per hour.

Potential erosion may be calculated from

$$E_p = nte_p,$$

where E_p is the potential wind erosion [t ha^{-1} year^{-1}], n the wind frequency (occurrences per year), t the duration of wind [h], and e_p the soil erodibility [t ha^{-1} h^{-1}].

Example: if $v = 15$ m s^{-1}, wind frequency $n = 0.2$, duration of wind $t = 10$ h, and soil erodibility $e_p = 1.0$ t ha^{-1} h^{-1}, then $E_p = 0.2 \times 10 \times 1 = 5$ t ha^{-1} year^{-1}.

If the potential soil erosion, E_p, and the level of tolerable erosion, E_t, are known, the amount by which erosion control measures should reduce potential erosion is obtained. Since wind erosion has a stronger selective effect than water erosion, and since this effect differs according to the granular composition of the soil, it is recommended that the values for potential erosion, given in Table 94 be taken for soils of different grain structures.

A *reduction in the level of wind erosion* may be obtained in the following ways:

– by reducing wind velocity with a wind control barrier erected on the adjacent land,

Table 94. Values for potential wind erosion of the soil

Soil	Soil depth	Tolerable loss	
	[cm]	[mm year^{-1}]	[ha^{-1} year^{-1}]
Sandy	120	0.33	5.0
Loamy sand	60—120	0.2	3.0
Sandy loam	30—60	0.13	2.0
Loam	30	0.05	0.75

– by reducing wind velocity on the eroded field by means of increased surface roughness,
– by increasing soil resistance to erosion,
– by bringing about simultaneously a reduction in wind velocity and an increase in soil resistance and soil binding.

The *reduction of wind velocity* was described in the previous chapter. The size of the reduction needed, is established from a chart of soil erodibility for the given soil; the wind velocity which corresponds to tolerable erosion is found, and, according to the effect of the barrier chosen, its height and best position are determined.

As an example of the erosion control effect of low shelterbelts on heavily eroded sandy soils (the Chir sandy massif in the USSR), the data of Khimina (1973) may be cited. On unprotected territory in the investigated region a rate of soil removal of 366 t ha^{-1} was established (Table 95).

Table 95. Sand transport of fields protected by shelterbelts [t ha^{-1} h^{-1}]

Height of shelterbelt [m]	Width of field [m]	Distances from shelterbelts, H				
		2	5	10	20	30
5.5	180	0.4	2.6	19.2	131.7	347.1
6.0	240	1.1	0.3	2.9	172.8	358.1

In this case shelterbelts were effective when they were spaced at a distance of about ten times the height of the shelterbelt. Overall, the shelterbelts reduced wind erosion about 3.5-fold.

A similar effect is produced by *coulisses* formed by the stalks of high plants which may be distributed according to cropping pattern and crop rotation. Coulisses have proved to be effective especially in dry regions where crop yields depend on the amount of winter and spring moisture. Coulisses prevent the blow off of both soil and snow, and thus improve the soil moisture contents in adjacent fields. Baraev (1971) reported that mustard coulisses 70 to 95 cm high increased the height of the snow cover from 8 to 10 cm, to 40 to 60 cm, and improved the yields on experimental stations in Kazakhstan by 52%, from 16.4 cwt ha^{-1} to 25.0 cwt ha^{-1}. The considerable beneficial effect of coulisses has been observed in other steppe regions also.

On undulating terrain erosion is more severe on the windward than on the leeward sides of slopes. The required spacing between *shelterbelts* on the windward side is computed from the formula

$$L = \frac{\alpha H}{1 + \alpha I},$$

where L is the spacing between shelterbells [m], α the coefficient of shelterbelt effect on level terrain, H the shelterbelt height [m]; distance between shelterbelts on level terrain $L' = \alpha H$, and I the angle of ground inclination.

Example: if the shelterbelt height *(H)* is 20 m, the coefficient of shelterbelt effect (α) is 25, and the gradient *(I)* is 0.05, then $L = (25 \times 20)/(1 + 25 \times 0.05) = 222$ m.

A reduction in the wind velocity on eroded fields may be achieved, in the first place, by adjustment of the soil surface to obtain a ***greater degree of roughness;*** fences, grass strips, and other forms of soil conservation may also be employed. The relationship between grass density and wind erosion is shown in Figs. 134–136.

A very considerable degree of soil conservation results from the residues of various crops which also protect the soil while they are growing. Shiyatyĭ (1965b) established the following relationship between soil erodibility and both soil structure and the number of stalks in cereal stubbles

$$Q = 10^{4.03691 - 0.03845 - 0.00406N},$$

where Q is the soil erodibility [g m^{-2}], S the soil structure (content of particles), and N the number of stubble stalks over 20 cm long per m^2.

Example: if $S = 45\%$ (contents $d > 1$ mm), $N = 0$, then lg $Q = 2.26441$, and $Q = 183.9$ g m^{-2}, as may be seen in Fig. 128 without calculation. An erodibility of 183.9 g m^{-2} with a wind velocity of 12.48 m s^{-1} in the tunnel equivalent to 22m s^{-1} at 10 m height, means that erosion exceeds the tolerable limit several times. From Shiyatyĭ's formula and Fig. 137, it may be seen that a stubble field with as few as 50 stalks per m^2 reduces erosion to a tolerable level, and if the number of stalks is 150, the soil is very adequately protected against wind erosion, since erodibility is reduced to 50 g m^{-2} – a reduction of 3.68 times. When the number of stalks increases to 300, erodibility decreases to approximately 20 g m^{-2}.

From a knowledge of stand height and stand density, the protective effect of *vegetation* may be determined with relative accuracy. A dense grass or forest stand fully protects the soil from erosion. For the calculation of actual or expected wind

Fig. 134. Soil protection against wind erosion provided by tussocks of *Ammophila arenaria* planted with a wide spacing on a littoral dune (Beachport, South Australia). (By courtesy of Department of Agriculture, Adelaide, A.S.)

Fig. 135. Soil protection against wind erosion on a littoral dune densely planted with grass tussocks. (Photo D. Zachar.)

Fig. 136. Soil protection against wind erosion on littoral dunes achieved by means of rows of densely sown rye (*Secale cereale*). The tall stubble protects the soil adequately even after the harvest.

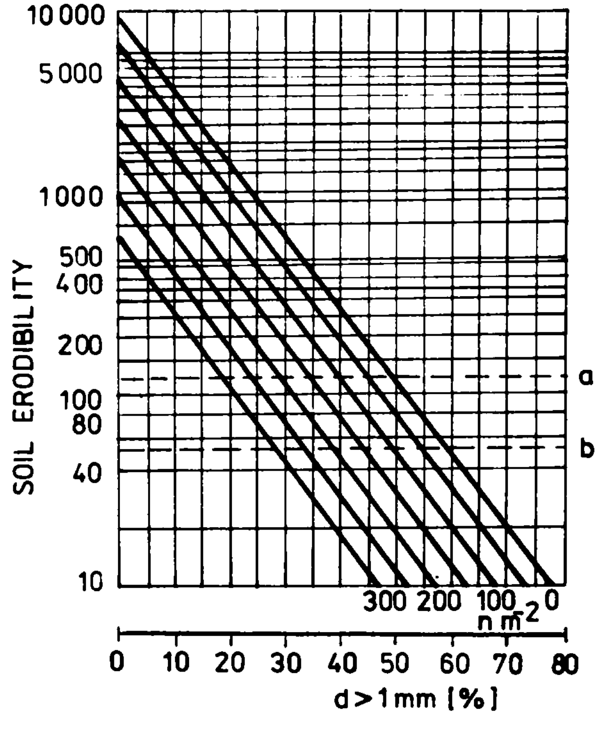


Fig. 137. Variation of soil erodibility in relation to soil structure and density of stubble stalks at a wind velocity of 12.48 m s^{-1} (soil erodibility in g m^{-2}, d – diameter of soil grains, n – number of stalks of stubble per square metre).

erosion, use may be made of the coefficients derived for the corresponding calculation with respect to water sheet erosion; these coefficients vary around 1, decreasing on unprotected land to values close to zero. Whereas in water erosion the effectiveness of the stand is determined by the stand density, as well as the penetration of roots into the soil and by the length of the period in which soil is protected by vegetation, in wind erosion the stand height is of importance in determining the effect of the stand on wind erosion on the adjacent land.

From among the many data confirming these relationships, one example will be given which fully demonstrates the relationship between stand height and the size of the protected area. In 1957, after dust storms had occurred over the "Kubanskiĭ" kolkhoz (USSR), it was established that winter cereals situated on the "lee side" of permanent grass stands were damaged to an extent in proportion to the distance from the border with the grass, as shown in Table 96 (Zaĭtseva 1970).

Table 96. Number of plants of spring wheat surviving after dust storms in 1957

Number of stalks	Distance from edge of grass stand [m]						
	25	50	100	150	200	300	500
Stalks m^{-2}	299	261	238	151	67	48	Sparse
% of number at 25-m distance	100	87	80	51	22	16	0

Inhibition of erosion increases not only with the number of stalks or plants, but with the quantities of plant parts both below and above the ground also. Zharkova et al. (1973) established the following relation for easily erodible soils

$$V_k = a \frac{FR}{F + R} ,$$

where V_k is the erosion control coefficient of vegetation, a the vegetation cover [%], F the weight of plant matter above the ground [cwt ha^{-1}], and R the weight of underground plant matter [cwt ha^{-1}].

The formula was originally intended for the assessment of the inhibiting effect of vegetation on water erosion, but it is valid for wind erosion also, provided that the height of the stand is taken into account. The following values were established for V_k on a soil threatened by very severe erosion:

V_k	Above 48	28 to 48	14 to 28	6 to 28	Below 6
Erosion	Negligible	Slight	Moderate	Severe	Very severe

As far as *crop rotation* is concerned, it is recommended that alternating strips be used as in the case of water erosion control, various combination of the protected and unprotected part of the field being possible. Shiyatyĭ et al. (1973) recommended a crop rotation in strips, as shown in adapted form in Table 97 for various grades of soil susceptibility to wind erosion.

In practice, from the third grade of erosion danger upward, shelterbelts are required for adequate protection against wind erosion. Shelterbelts such as these are also known as ***deflation control barriers*** or ***ventoadverse shelterbelts.***

For the sake of clarity, a scale for the deflation control effect of various crops is given, in which a grade is assigned to each case (Table 98).

As in the case of plant roots, organic matter ploughed into the soil also tends to reduce wind erosion. Data collected by Yakubov (1946) on soil blow off on soils of high susceptibility to wind erosion are given in Table 99.

In addition to the amount of ploughed-in straw, the length of the straw also has an effect on the erosion resistance of the soil; the longer the straw length, the greater is the inhibitory effect on erosion.

Soil management and the nature of the soil surface resulting from cultivation are of basic importance to soil erodibility. During ploughing, less resistant parts of the

Table 97. Crop rotation on soils under different degrees of danger from wind erosion

Grade of potential wind erosion	Crop rotation
1	Crops rotated as desired; agricultural measures adequate; sowing carried out perpendicularly to the wind
2	Strips of annual crops varying in protective effect
3	Strips of annual crops in rotation with perennial grass
4	Strip culture, predominantly with perennial grass
5	Permanent crops raised in some areas
6	Extensive permanent crops and forest plantations

Table 98. The protective effect of crops against wind erosion

Grade of wind erosion	Crop	Coefficient V_k
1	Complete protective effect from dense forest	<0.01
2	Complete protective effect from perennial grass	0.1—0.1
3	Diminished protective effect from annual grass	0.1—0.3
4	Low protective effect from winter cereals	0.4—0.6
5	Slight protective effect from spring cereals	0.4—0.6
6	Very slight protective effect from root and tuber crops	0.7—0.9

Table 99. Effect of the amount of ploughed-in straw on soil removal by wind

Susceptibility of soil to erosion	Wind velocity at 30 cm height [m s^{-1}]	Amount of soil blow-off [t ha^{-1}]					
		Control	0.6	1.3	2.5	5.0	10.0
High	7.6	6.4	3.0	1.7	1.3	0.1	0
	9.8	13.4	8.5	7.7	3.2	1.4	0
Very high	7.6	60.9	26.8	13.6	6.5	1.0	0.3
	9.8	123.6	62.5	37.6	23.6	7.5	0.6

Table 100. Soil erodibility arising from various types of fallow processing

Soil processing	Soil erodibility [g m^{-2} 5 min^{-1}]	
	In autumn, after ploughing	In spring
Ploughing with furrow turning to 25—27 cm depth	59.2	149.3
Soil loosening without furrow turning or straw incorporation	22.2	48.4
Ditto, with incorporation of the straw	6.4	10.9

soil not penetrated by plant roots come to the surface, thus increasing its erodibility. Ploughing also causes changes in the moisture content and other soil properties; in particular, the soil structure may be altered. Data on the combined effects of ploughing and crop residue on soil erodibility (Zaïtseva 1970), are given in Table 100.

It should be mentioned finally that any measures taken to increase soil fertility also provide an effective means of deflation control. These measures include *fertilizing,* making *improvements to the water regime of the soil,* and *raising the humus content* and levels of *organic matter in the soil,* etc. The relationship between wind erosion and soil fertility is similar to that between water erosion and soil fertility.

Various *chemical substances* produced in large quantities in various countries have also been used as a means of increasing soil resistance to wind erosion. Among the most successful of these preparations is "Neerozin", which is produced in the USSR and which protects the soil to wind velocities of 28 to 30 m s^{-1}, and even to winds of 40 to 42 m s^{-1}. Neerozin is a modified kastrobiolit; it is permeable to water, allows seed germination, but is slightly toxic. It is recommended for the stabilization of shifting sand. In the CSSR, "Antieroza", which has a similar composition to that of "Neerozin" but includes accelerating and nutrient components which improve the growth and protection given by vegetation, is used for the stabilization of eroded land. In West European countries "Krilium", and in the USA, "Turbifer" are used.

The problems of soil conservation will be discussed in a separate work.

In concluding this section on wind erosion, it should be emphasized that the issue of wind erosion is complex. Until now it has been studied mostly with respect to cultivated land where wind erosion causes significant economic damage. A serious problem is posed by the expansion of sand deserts and the extension of wasteland as the result of a gradually increasing intensity of wind erosion and a weakening of the protective effect of vegetation. Therefore the control of wind erosion is important both in terms of world food production for the growing populations of mankind, and as a factor in the protection of the living environment. An evaluation of present information on wind erosion and the possibilities for its control shows that little more than a start has been made, but even our present knowledge, if correctly applied in practice, entitles us to look to the future with some optimism.

Among the main causes of the acceleration of wind erosion are:

- deforestation, reductions in the Earth's surface roughness, and increasing wind velocities,
- the removal of permanent vegetation and its replacement by crops with a smaller protective effect,
- soil disintegration, the breakdown of soil structure, destruction of humus, and desiccation and salination of the soil,
- damage to vegetation and soil, caused by the grazing of cattle and wild animals (rabbits, kangaroos, etc.), and also by burning,
- the acceleration of water erosion and the formation of light sediments which are highly susceptible to wind erosion,
- damage to vegetation caused by industrial fumes, and other injurious agents which are ecologically undesirable.

On the other hand, positive results are obtained by

- improvement and fertilization of the soil together with optimalization of the water regime of the soil,
- raising varieties and species of crops with a higher biomass productivity and thus with higher conservation effect also,
- applying more intensive management systems, increasing yields of grass stands, suppressing forest grazing, and organizing efficient fire control, etc.

The most important protective measure on intensively cultivated land are:

- the correct distribution of tall vegetation – optimizing the structure of forest belts to form a permanent skeleton of an ecologically balanced landscape,
- the additional distribution of coulisses formed by the taller agricultural crops,
- the raising of crops of lower protective effect in rotation together with plants that give good protection against wind erosion,

- the alternation of crops in strips running at right angles to the direction of the prevailing harmful wind, and contour management on terrain with a 3 to 6° angle of inclination, where wind erosion occurs together with water erosion (strips following the contour lines),
- mulching of the soil surface with organic matter and ploughing in, or otherwise incorporating the organic matter into the soil,
- increasing the content of organic matter in the soil so as to improve the water and nutrient status of the soil together with its general fertility,
- increasing the roughness of the soil surface by judicious selection of cultivation methods and utilization schemes,
- finding comprehensive solutions to the problems of soil conservation in the context of the overall ecological optimalization of the landscape.

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Chapter 5

DISTRIBUTION OF EROSION

A number of related matters can be included in the general subject of soil erosion, such as the influence of erosion on the soil, the ecological consequences of erosion, its economic assessment, the control of erosion, etc., but as far as the scope of this work is concerned, it will suffice to include a chapter on the *distribution of erosion phenomena* in various natural and economic environments, together with some mention of *erosion control.* In this chapter only a brief survey can be presented, which is unfortunately incomplete, and therefore unable to give a fully balanced account. One of the reasons for this is the scarcity of data relating to some parts of the world.

5.1 Europe

Europe is a continent with intensive agricultural production and relatively favourable conditions. Nevertheless a large area of agricultural land and also some forest land in Europe is impaired by soil erosion, and there is no prospect of erosion being reduced to a harmless level in the near future. The FAO estimated the proportion of unused land in Europe to be 22% in 1960, and 42% in the world as a whole. Of this proportion a large part is accounted for by erosion damage.

In studying the distribution of soil erosion it is necessary to delve into the past, because the acceleration of erosion is connected with the removal of vegetation, the conversion of forest land into agricultural land, the burning of vegetation, the extensive rearing of cattle, and many other negative consequences of human activity. The greater the effects of man's intervention and the more extreme the conditions of the *climate,* the more serious the consequences of erosion have been. Thus the countries around the Mediterranean are the most affected by erosion, with the *scrubby underbrush "maquis"*, the *rocky desert "garringues"*, the *water-eroded soil of the "calancos"*, etc. The most dangerous forms of erosion are *torrent erosion,* the formation of *mud flows, snow avalanches* and *snow erosion,* the formation of *sand dunes* especially in maritime regions, and the acceleration of *suffosis phenomena* and *underground erosion.*

A short survey of the distribution of erosion in various countries is given in the following.

The European part of the USSR

In the European part of the USSR a great variety of conditions and predominant forms of vegetation can be found, resulting in a corresponding variety of forms of erosion. In general, erosion of the soil increases from North to South and from West to East. The most intense forms of soil erosions occur in the plains and mountains of the southeastern region of this part of the world.

In the cold northern regions, forms of linear, aeolian and thermokarst erosion occur, and in the forest belt less intense forms of gully erosion occur together with various forms of sheet erosion associated with the forest steppe belt. With increasing aridity of the climate the intensity of precipitation increases, and in the dust storm belt, which includes a large proportion of the cultivated land, the intensity of wind erosion increases towards the South. Typical erosion forms develop in regions of extensive loess deposits where a network of *gullies "ovragi"* and *ravines "balki"* accelerates the erosion process. In semideserts and deserts *wind erosion* is the predominant form and in mountain massifs mud flows (Russian *sel, selevyĭ potok*) are frequent. Forms of underground erosion are also widely distributed. Erosion phenomena are most common on agricultural land, in desert and semidesert regions, and on mountains with sparse forest and meadow vegetation.

According to Sobolev (1961) about 50 million ha are damaged by precipitation erosion in the European part of the USSR, including 10 to 11 million ha of moderately and severely eroded land (harvests reduced by 70 to 80%) and 2 million ha of soil completely destroyed by erosion. On slightly rain-washed soils, the yields of agricultural crops are diminished by 30 to 40%. According to older data (Sobolev 1948), the annual increase in the area of soil impoverished by sheet erosion is estimated at 150,000 ha, and the annual increase in the area of soil damaged by gully erosion is thought to be 45,000 ha. Drifting sands are estimated to increase at the rate of 140,000 ha year^{-1}. Underground erosion of the soil occurs in regions consisting of carbonate rocks, rocks of high salt content or loess, and also in some mountain massifs mainly in the Caucasus (Maksimovich 1955).

About 6.5 million ha of sandy soil under cultivation is damaged by wind erosion annually (Zaĭtseva 1970), although the total area covered by wind-eroded soil is much larger. The annual increase in the area of soil damaged by wind erosion is about 140,000 ha (Sobolev 1960). Conservation measures against wind erosion in the steppe and forest steppe are urgently needed on 10 million ha of sandy and sand-loam soils.

Approximate calculations made by Zvonkov (1962), indicate that in the entire USSR (land area: 22,402,200 km^2) about 200 million ha are damaged by water erosion, including 7 million ha of gullies and completely ruined land, 40 million ha

Fig. 138. General distribution map of soil erosion in the USSR. 1 – sandy soils severely damaged by wind erosion, 2 – northern boundary of region of dust storms. Percentage of cultivated land affected by erosion: 3 – 0 to 5%, 4 – from 5 to 20%, 5 – from 20 to 40%, 6 – over 40%, 7 – surface protected by permanent vegetation, 8 – swamps.

of severely eroded land, 53 million ha of moderately eroded land, and about 100 million ha of slightly eroded soils. There are at least 200 million ha of sandy and sand-loam soils damaged by wind erosion, including 65 million ha of drifting sand.

Figure 138 shows the distribution of soil erosion in the USSR; the map was prepared from the work of Sobolev (1960), and is based on a simplified scale of erosion distribution. For the various republics and regions, detailed maps showing soil erosion are available, and in addition to actual erosion, maps of potential soil erosion have also been prepared.

Rivers crossing the plains of the European part of the USSR carry away about 47 million tons of silt annually, and from the whole of the USSR, about 535 million tons of silt containing about 12 million tons of K_2O, 593,000 tons of P_2O_5 and 1.2 million tons of N are removed (Sobolev 1960). According to Lopatin (1952), the rivers of the USSR carry away a total of 472.3 million tons of suspended matter and 846.3 million tons of dissolved substance, the larger part of which consists of the products of water erosion of the soil. The total runoff of suspended and dissolved material in the USSR, according to Lopatin, is 1,318,600 tons, representing a mean specific runoff of silt amounting to 40.3 t km^{-2} which, as his calculations show, is 3.3 times smaller than the corresponding world average.

The intensity of erosion varies a great deal in different parts of the European part of the USSR. Sobolev (1960) found that on tilled land in the plains, erosion varied according to the inclination of the ground surface from negligible values up to catastrophic levels of erosion (Table 101).

Table 101. Approximate area affected, and intensity of sheet erosion in the plains of the European part of the USSR

Ground inclination	Area affected [million ha]	Rate of soil removal from arable land [m^3 ha^{-1} $year^{-1}$]
1—2°	50	2—31
2—4°	16	7—42
4—6°	6	8—96
6—8°	2	10—114
8—16°	?	14—154

The intensity of wind erosion varies as a function of wind velocity, wind duration and soil erodibility, from negligible values up to 200 m^3 ha^{-1} $year^{-1}$ or more. Considerable losses are caused by dust storms which are separated by relatively long-term intervals. After the great dust storm which occurred in the spring of 1882, heavier storms recurred in 1928 with the result that in certain regions of the Ukraine, 12 cm of soil (1,200 m^3 ha^{-1}) were borne away. The total soil removal in the Ukraine was estimated at 15 million tons, 5.4 million tons being completely removed from the affected regions (Zaïtseva 1970).

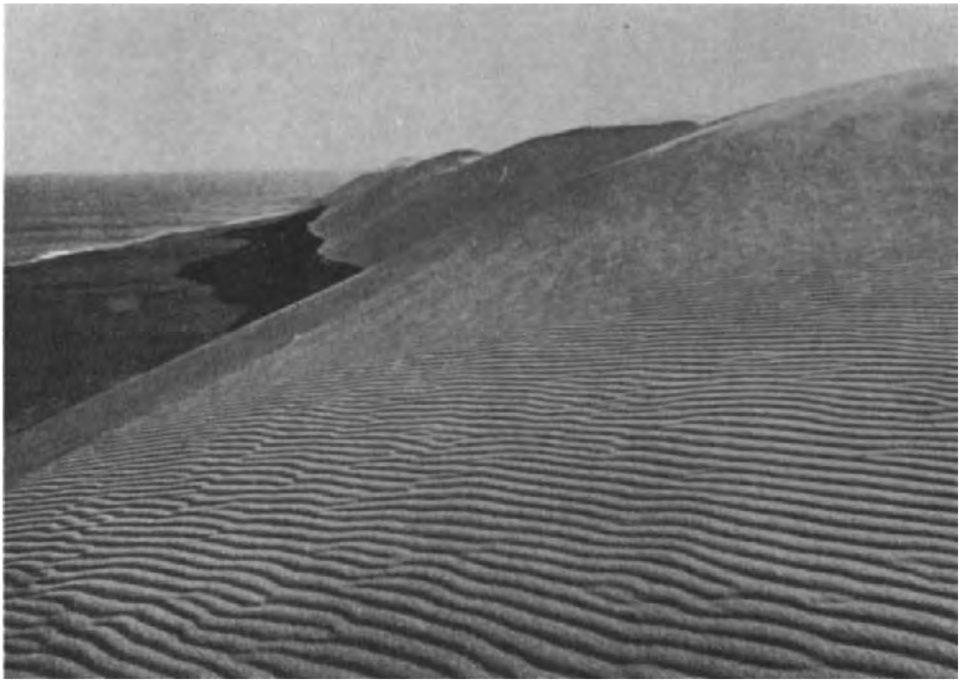
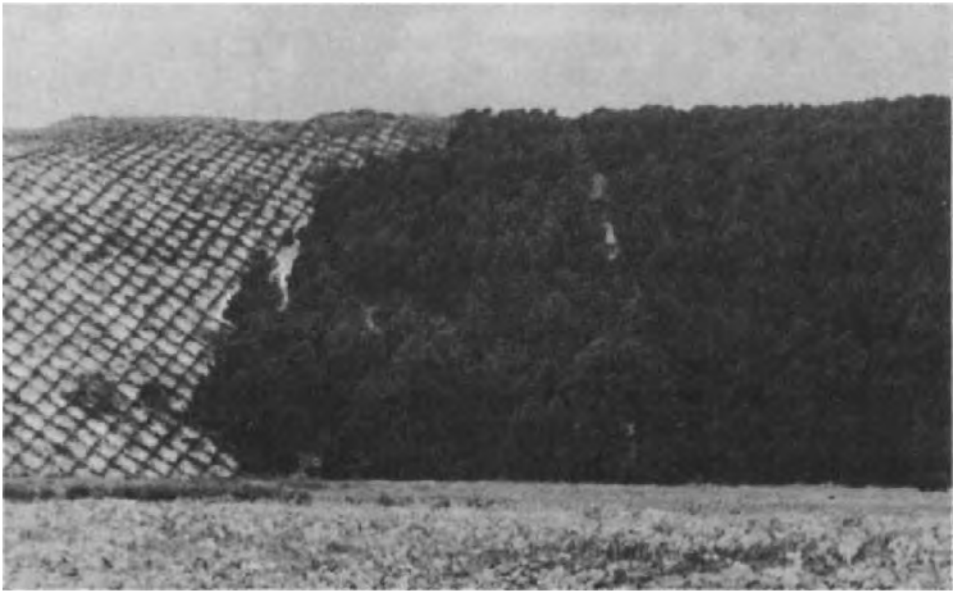


Fig. 139. a – Littoral dune covered by ripples (Kurskaya kosa, USSR). (The author's collection of photographs.) b – Binding of littoral dunes on the Kurskaya kosa by means of fences and forest vegetation. (Photo J. Novotný.)



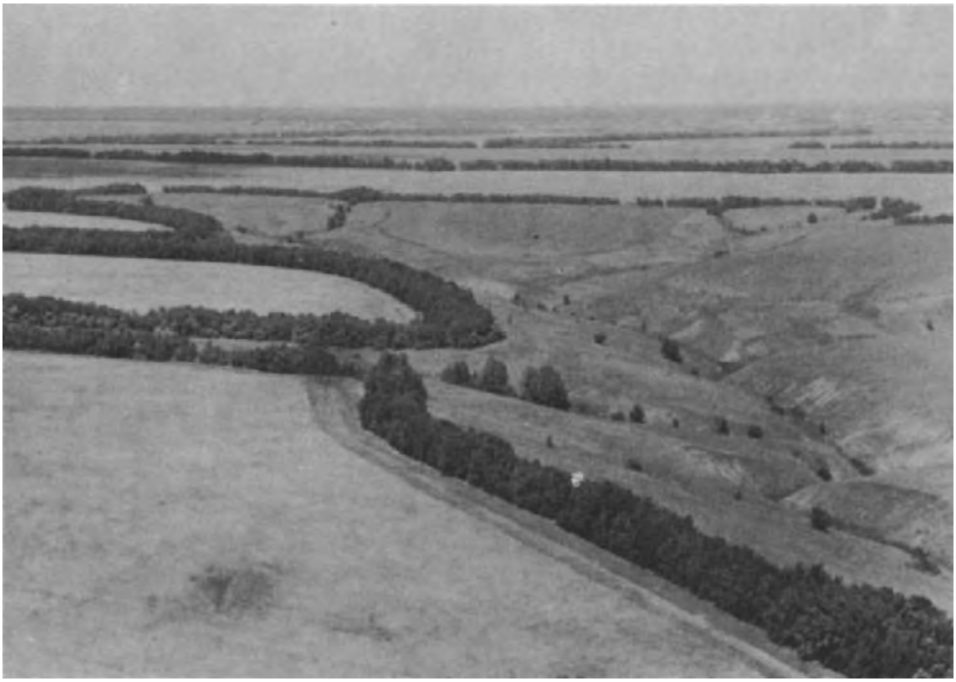


Fig. 140. System of shelterbelts on the Kelt Research Station (Don river basin, USSR). (Photo G. Lvovskiĭ.)

Among the most damaging dust storms were those which occurred in March and April 1960 affecting fields over an area of 4 million ha, from which 980 to 1,280 million tons of fine earth were removed. It is calculated that in these storms, 300 to 400 m^3 ha^{-1} of soil was blown off, and deposits in some places were up to 4.5 m thick (Doskach and Trushkovskiĭ 1963).

Of the Soviet republics, the *Ukraine* has the largest area of soil damaged by erosion (601,000 km^2); out of 42 million ha of land investigated, 12 million ha (28%) were found to be affected by erosion, including 3.7 million ha suffering from moderate and severe levels of erosion, and about 300,000 ha completely destroyed by erosion. The total area of eroded soil is steadily increasing. According to Sokolovskiĭ, water erosion affects 13 million ha, causing moderate to severe damage on 5 million ha. About the same area of land is attacked by wind erosion.

In the *Russian S.F.S.R.* (total area: 17,075,000 km^2) there are data for 1961 giving 6.1 million ha as the area affected by erosion, including 2.6 million ha of tilled land. Outside the chernozem region, about 1.3 million ha were damaged by moderate to severe levels of erosion. The total area of eroded soil increased as the amount of land under cultivation increased.

Fig. 141. Erosion control dams built along the banks of ravines (Desna river basin, Ukraine, USSR). (Photo D. Zachar.)

In *Belorussia* 59.2% of the cultivated land surface is affected by erosion (12.8% by severe erosion, 22.6% by moderate erosion, and 23.8% by slight erosion) (Medvedev 1968). A total of 587,000 ha of tilled land is damaged of which 180,000 ha are damaged by wind erosion. In the neighbouring republics erosion affects 10% of the land in *Lithuania,* 9% of the land in *Estonia,* and 2.5% of tilled land in *Latvia.*

In *Moldavia* (total area 33,700 km^2), more than 50% of the land surface is endangered by water erosion (Zaslavskiĭ 1966). In addition to sheet erosion, gully erosion is also widespread. Rozhkov (1973) established that the annual overall growth of gullies in Moldavia was 60 to 70 km, the increment in the area covered by gullies was about 200 ha, the increase in the surrounding eroded area was 800 to 1,000 ha, the annual deposition was 10 to 15 million tons of material, and about 1,000 ha of land were rendered useless in flood zones.

The erosion control measures used most commonly in the USSR are: afforestation, the establishment of infiltration, erosion and deflation control belts, stabilization and afforestation of ravines, stabilization of sands (Fig. 139), torrent and avalanche control, and other measures concerning the hydrographic network. On agricultural land, systems of erosion control measures (Figs. 140, 141) based on intensive research have been introduced, and these are adjusted to suit different regions according to the degree of soil erodedness, economic utilization, and other factors. However, serious problems are encountered when erosion control is required in mountain areas. Soil protection is decreed by law in the USSR and has been put into practice since the state came into existence. The work of Dokuchaev is of some historical importance since it is centred on a comprehensive theory of the soil and the improvement of its fertility by a system of measures including erosion control.

Romania

Of the 237,502 km^2 total land surface of this state, about 7.3 million ha (30.8%) are endangered by erosion, about 3 million ha being affected by moderate to severe erosion, and 840,000 ha being threatened by very severe erosion. Of the 6.3 million ha of forest land, about 200,000 ha are affected by erosion (Pimpirev 1957, Moţoc 1956, 1963). In addition, a considerable area (about 600,000 ha) of agricultural land is affected by wind erosion, river sands covering an area of 98,000 ha suffering the most. A high degree of protection is required by about 4 million ha of soil.

According to a more recent survey (Ionescu 1972), 2.555 million ha of agricultural land in Romania is damaged by water erosion, including 2.144 million ha damaged by sheet erosion and 0.084 million ha carved into gullies. Underground erosion phenomena are also common.

Afforestation is much used as a means of soil protection (since 1950 more than 100,000 ha of eroded land have been planted for the purpose of torrent control and the binding of silt in the basins of water reservoirs). Further action against erosion on fertile fields takes the form of crop rotation improved soil management, and terracing; terracing is carried out over a total area of about 120,000 ha, comprising 40,000 ha of vineyards and 80,000 ha of orchards.

Poland

From the total land surface of 331,730 km^2 about 4 million ha of soil in Poland (12%) is susceptible to water erosion, including about 1 million ha (3%) suffering from severe erosion (Żiemnicki and Józefaciuk 1965). About 8% of the soil in the country is threatened by wind erosion. Whereas water erosion occurs in more severe forms in southern Poland reaching its most severe form in the Carpathians, wind erosion is more intense in the northern, maritime regions. Yields of agricultural crops on moderately and severely eroded land are reduced by 40 to 70%. The rivers carry away about 5 million tons of silt annually representing a specific silt runoff of about 15 ton ha^{-1}. Annual losses of water in Poland owing to erosion are estimated to be 300 million m^3 (Żiemnicki 1968).

Erosion control schemes focus mainly on soil protection in mountain regions and the binding of sand in maritime regions. Special emphasis is laid on protective afforestation and on torrent and gully control.

Czechoslovakia

Of the total land surface of the country (127,877 km^2), approximately 2.93 million ha (23%) are damaged by precitipation erosion and 1.63 million ha (12.6%) by wind erosion. The greater part of the eroded soil occurs on agricultural land. The mean runoff of silt is estimated to be 4 million tons per annum, and the greatest intensity of precipitation erosion measured during an exceptionally heavy downpour of 100 years frequency of recurrence was 2,000 m^3 ha^{-1}; the most intense wind erosion measured was 75 m^3 ha^{-1} year^{-1}.

The distribution of soil erosion in Czechoslovakia is shown in Figs. 142 and 143.

The control of soil erosion has been practised in Czechoslovakia since 1852. Examples of the first efforts in this direction are shown in Figs. 144–147; this work included the control of torrents, gullies and ravines, whereas more comprehensive erosion control measures have been introduced only in the last 30 years, during

Fig. 142. General distribution map of rainfall erosion of the soil in Czechoslovakia. Proportion of territory affected by sheet erosion: 1 – less than 25% affected, gully erosion less than 0.1 km km^{-2}, 2 – from 25 to 50% affected, gully erosion from 0.1 to 0.5 km km^{-2}, 3 – from 50 to 75% affected, gully erosion from 0.5 to 1.0 km km^{-2}, 4 – over 75% affected, gully erosion over 1.0 km km^{-2} (from data collected by Š. Bučko, O. Stehlík and M. Holý; compiled by D. Zachar).

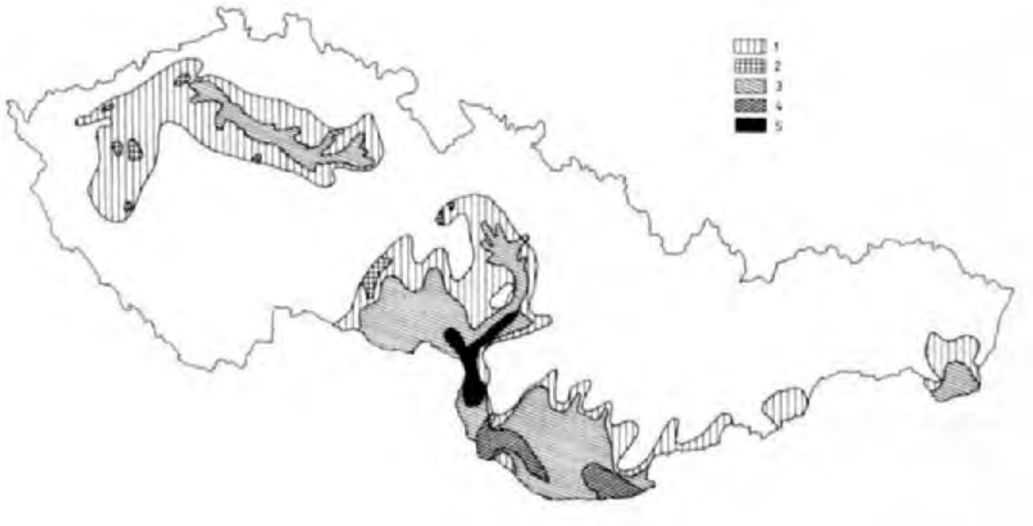


Fig. 143. General map of potential wind erosion in Czechoslovakia according to Pasák (1978). 1 – slight, 2 – moderate, 3 – moderate to severe, 4 – severe, 5 – very severe.

Fig. 144. Soil stabilization by means of a system of technical measures (structures), afforestation, and grass sowing. Neighbourhood of Rakovník (Czechoslovakia). (The author's collection of photographs.)

Fig. 145. Soil stabilization on a shallow ravine using palisade fences. Water is diverted by a stone pavement. After stabilization, the area was planted with trees (Jeseník Mts, Czechoslovakia). (The author's collection of photographs.)

which time about 150,000 ha of eroded land have been planted with forests and several types of erosion control have been put into practice on agricultural land. Terracing has been carried out for the establishment of vineyards and orchards in the most fertile regions (Fig. 148).

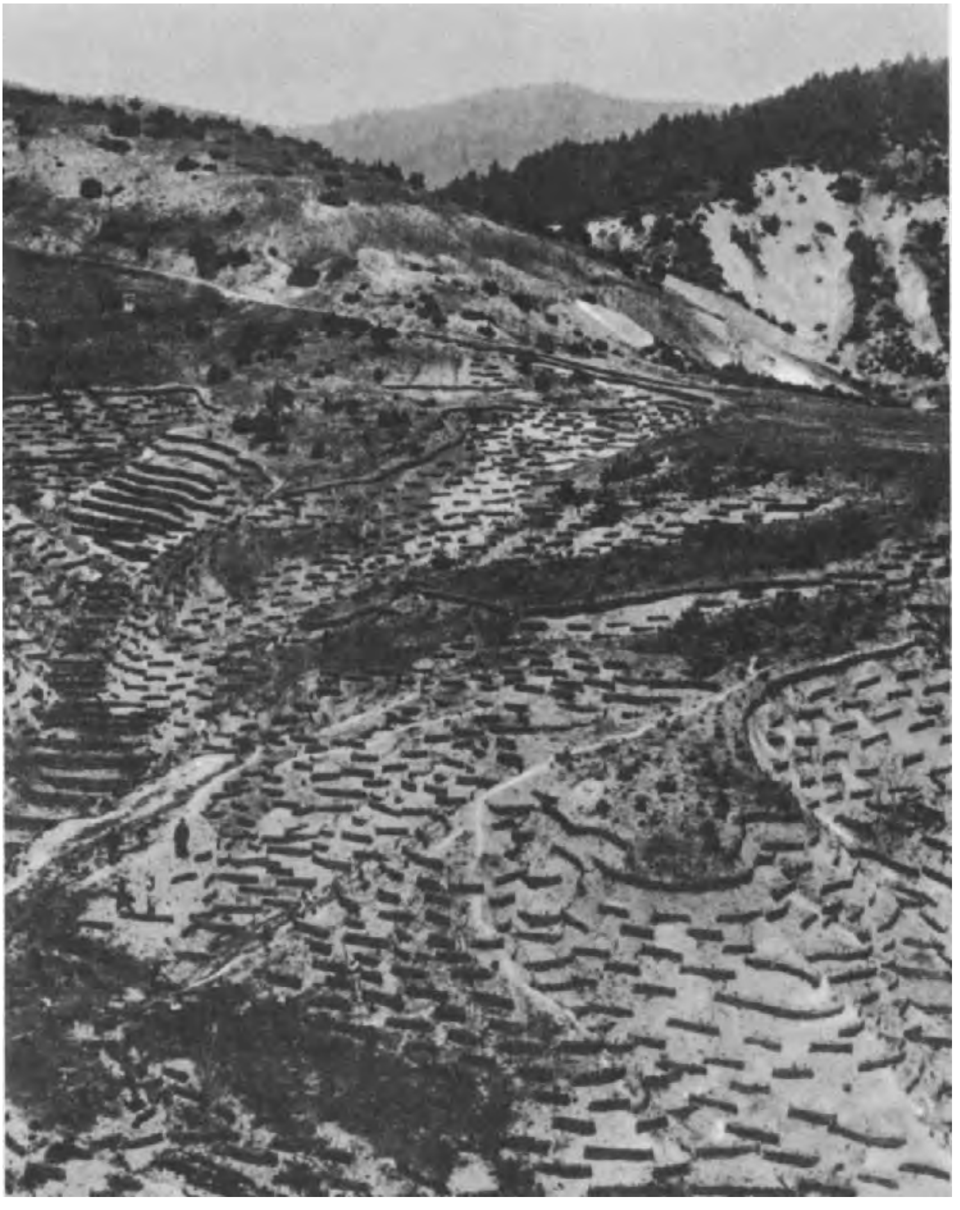


Fig. 146. Stabilization of wasteland by means of fences erected towards the end of the last century. Neighbourhood of Banská Bystrica (Czechoslovakia). (The author's collection of photographs.)

Fig. 147. View of the same area 65 years later. (Photo D. Zachar.)

Fig. 148. Soil terracing for the establishment of orchards in southern Moravia (Czechoslovakia). (Photo D. Zachar.)

Hungary

Hungary has a less broken relief, less resistant rocks, and a more extreme climate compared with Czechoslovakia. Of the 93,030 km^2 total land surface of the country, 2.297 million ha (24.6%) are affected by water erosion and 1.449 million ha (15.5%) by wind erosion. About 858,000 ha (9.2%) suffer from severe precipitation erosion 885,000 ha (9.5%) from moderate precipitation erosion, and 554,000 ha (5.9%) from slight precipitation erosion (Stefanovitz in Kovacs 1977). In addition to erosion, Hungary is affected by frequent flooding and the obstruction of fields by deposits. Yields of agricultural crops are reduced by 20 to 30% on slightly eroded soil, by 30 to 40% on moderately eroded soil, and by 35 to 60% on severely eroded soil. In addition to surface erosion, underground erosion also occurs in Hungary.

Systematic afforestation schemes are carried out for the purpose of erosion control; in the last 30 years over 300,000 ha of land have been planted, a large part of this being primarily for the inhibition of erosion. Other control measures in force in Hungary involve the establishment of shelterbelts, and the control of erosion on tilled land by means of crop rotation, improved management techniques and terracing on smaller areas.

Bulgaria

Bulgaria, which experiences rather more extreme conditions in general, is relatively more affected by erosion than other countries, and in the past a large part of its territory has been damaged by erosion. Of the 110,549 km^2 of land surface, up to 1,748,830 ha (15.9%) were lying fallow in 1951, and severe erosion occurred on forest land which at that time occupied an area of 3,670,650 ha (35%). The severe and moderate categories of erosion together damaged 73% of tilled land (Biolchev 1955) and 22% of forest land (7% of the total land area) (Aleksandrov 1966). The specific runoff of silt and bedload into the Dimitrov and Stamboliĭski reservoirs was 1,090 and 21,600 $m^3\ km^{-2}$, respectively, so that in the Stamboliĭski reservoir 6.63 million m^3 of deposits sedimented out in four years. Wind erosion is also very severe, but less well documented than water erosion (Georgiev 1977). In March and April 1957 a layer of soil 5 to 6 cm thick (500 to 600 $m^3\ ha^{-1}$) was removed from extensive areas of unprotected land, crops were completely destroyed locally, and approximately 20 million tons of humus containing fine earth were carried away. The dust storms recurred in 1976.

Fig. 149. Soil stabilization achieved with a flood retention barrier and forestation of the catchment area. (Photo D. Zachar.)

For this reason, a great deal of attention is given to erosion control in Bulgaria, and erosion has been almost entirely stopped on forested land. Most of the wasteland has been planted with trees, measures against flooding and gully erosion on a large scale have been taken (Fig. 149), and riparian stands have been established. Terracing is used to a large extent on agricultural land.

Yugoslavia

Conditions in Yugoslavia are very conducive to erosion and out of the total land surface of the country (225,804 km^2), there are 5.661 million ha of almost completely denuded karst land. Of the total area of karst, 1.075 million ha (19%) are used for the cultivation of agricultural crops, whereas another 2.105 million ha (38%) are totally barren. The total area of soil damaged by karst erosion is 3,339,480 ha, i.e. 59% (Bura 1955). Land outside the limestone region is also affected by erosion. According to Gavrilović (1972) and Lazarević (1973), 193,675 km^2, i.e. 75.71% of the territory of the country, are affected by erosion. Erosion in diverse forms may reach harmful levels over 91.1% of the land surface, and sedimentation takes place on 8.9% of the land, producing 89.76 million m^3 of deposits annually, 37.25 million tons of which are transported off the land. In terms of cost, 167,546,900 dinars were spent during the period from 1954 to 1968 on the control of water erosion (the annual economic loss due to water erosion is estimated to be 362,960,400 dinars), 74% being allocated to work on river-beds, 14% to afforestation, 2% to turfing, 4% to terracing, and 6% to other work in the catchment area (Lazarević 1973). Wind erosion occurs over a smaller land area, but causes a lot of damage (Jevtić 1973).

Albania

The territory of Albania, like that of Yugoslavia, is affected by very severe erosion. Of a total land surface of 28,748 km^2, over 80% are endangered by potential erosion. According to Nako a considerable part of the forest land which covers 41.2% of the total land area is eroded. Unfavourable natural conditions and irresponsible land management have brought about the highest specific runoff of silt in southern Europe. Fournier (1972) gives figures of 4,150 and 3,590 t km^{-2} year^{-1} for silt runoff into the Semani river near Urage Kucit and the Shkumbini river near Papere, respectively.

Greece

Greece, too, is one of the countries in Europe most afflicted by erosion. Throughout the distant and recent past the original vegetation has been destroyed over almost the whole of the 131,944 km^2 land area of the country, and land devastated by erosion occupies 2,312,800 ha (17,4%). Severely eroded grazing land supporting shrub growth of little value occurs over an area of 5,209,400 ha (29.4%), and also on the remainder of the land there is severe erosion (Moulopoulos 1960). Soil protection measures are required on 24.5% of the land, but only about 100 ha are afforested annually.

Turkey

Only a small part of Turkey is situated in Europe, this being much less damaged by erosion than the Asian part of the country. Nevertheless, very severe erosion occurs over a considerable area of the territory.

Italy

Italy is one of the most interesting countries to any study of erosion, because of the occurrence of a diverse range of natural conditions, as well as for the reason that in the past man has left behind traces of different types of soil utilization. The high relief of the Alps and Apennines, the low resistance to erosion of the sediments and some of the rocks give rise to high, even catastrophic levels of erosion, as may be seen from the values for the specific runoff of silt from the catchment areas of small rivers. Thus Lamone has a silt runoff of 2,420 t km^{-2} $year^{-1}$, and Savio has a silt runoff of 2,980 t km^{-2} $year^{-1}$ (Fournier 1972). Erosion of unusual forms and of high intensity occurs in regions with intense volcanic and tectonic activity. Special problems in Italy are caused by mass movements of soil ("frane") and badland formations ("calancos") which occur mainly in Lucania. Wind erosion occurs mostly in the maritime regions and affects about 900,000 ha (Gangemi 1963). Of the total of 301,200 km^2 land area of the country about three quarters are endangered by erosion.

Soil conservation has a long tradition in Italy. The main methods employed in the protection of agricultural land are described in the three editions of Oliva's work of 1952, and soil conservation and torrent control in the Alps are discussed by Wang (1904), and other authors. A great deal of attention has been given in the last 120 years to protective forestation; according to Mac Gregor about 2.2 million ha of eroded land in Italy were afforested up to 1957. Subsequent to this, another 600,000 ha of eroded land had to be afforested (Camaiti, 1962). Soil conservation presents difficulties in the Alpine regions of northern, as well as in southern Italy, as shown by the work of Puglisi (1963). The construction of a barrier in calanco terrain by means of a cableway is shown in Fig. 32. Figure 150 shows a system of earth dams for the stabilization of erosion gullies, and pavements along the ridges of erosion remnants for the drainage of surface water (Fig. 151). For the sake of completeness, Fig. 152 shows the "gradona" classical Italian terracing in Sicily, and Fig. 153 shows a difficult case of soil stabilization undertaken recently.

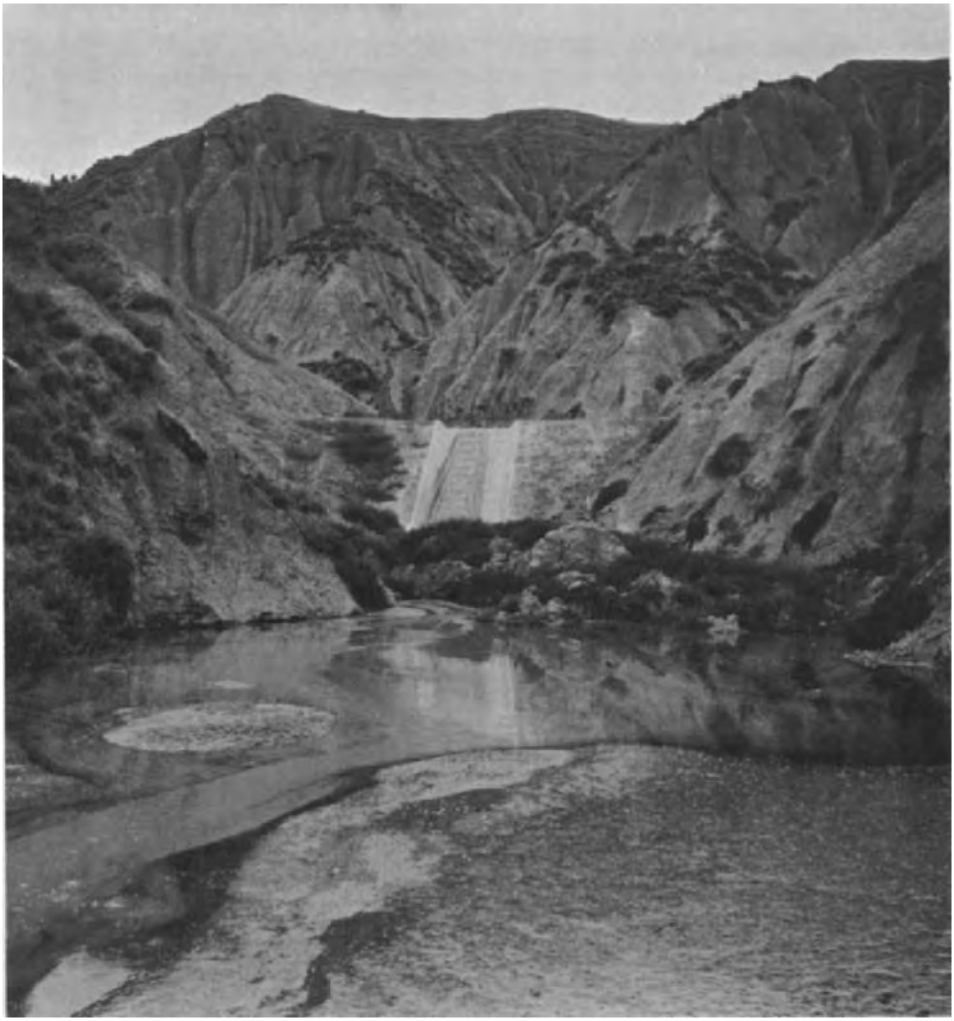


Fig. 150. Flood control achieved by a system of earth dams in the neighbourhood of Pisticci (Italy). (By courtesy of Ente Riforma, Bari.)

Fig. 151. Drainage of surface water by ridge channels, and gully stabilization. (By courtesy of Ente ▶ Riforma, Bari.)

Fig. 152. Classical stabilization of shallow, stony soil in Sicily. (The author's collection of photographs.)

Fig. 153. Difficult and costly soil protection measures in the Zonaro flood basin in Calabria (Italy). (Photo F. Rainer.)

Austria

Intense erosion processes occur in Austria which with a total area of 84,000 km^2, is host to almost all forms of erosion. Erosion control work is mainly directed at the control of torrents, ravines and avalanches, and at the stabilization of steep slopes and debris. After a successful work carried out in France, a detailed study was made of stabilization methods which were later applied in other European countries also. The credit for this work is due to Seckendorf (1884), Wang (1901), Strele (1950), and Schiechtl (1973). The reason for the mountig interest in the protection of soil and watercourses was the increasingly intense erosion approaching catastrophic proportions in easily erodible material. Gall (1953) established that the mean annual removal in the Tirol was as high as 760 m^3 ha^{-1} year^{-1}. In mud flows (Murgang) the rate of removal is even greater.

Figure 91 shows a typical ravine in fluvioglacial material, and Fig. 154 shows the stabilization of the same ravine by means of a system of concrete retention barriers and classical cordon planting, according to Couturier. At present, the slopes are fully stabilized.

Fig. 154. Stabilization of a ravine bottom together with slope adjustment before surface stabilization (a), soil stabilization by means of willow cordons and grass sowing (b). (The author's collection of photographs.)

Switzerland

This Alpine country (41,300 ha total area) is faced with problems of torrent erosion, debris flow (Murgang), avalanches, and an excessive sedimentation of deposits in valleys and water reservoirs. The relatively large flow of silt in the rivers is an evidence of the high intensity of erosion processes. According to Fournier (1972), the river Rhine at Lustenau has a specific flow rate of silt amounting to 843 t km^{-2} $year^{-1}$. Wind erosion also reaches considerable proportions in the Alps; Gall (1953) quotes an annual soil removal of 14.2 m^3 ha^{-1}. In spite of the extreme conditions, Switzerland has a relatively effective system of erosion control.

France

Over the total land area of 551,000 km^2, there can be found a great variety of erosion processes. Both Atlantic and Mediterranean influences operate here, and the Alps and Pyrenees also have a pronounced effect on the erosion pattern. According to Fournier, the Isère river in Grenoble carries a silt flow of 615 t km^{-2} $year^{-1}$, the Drac in Grenoble carries 780 t km^{-2} $year^{-1}$, the Garonne in Toulouse carries only 250 t km^{-2} $year^{-1}$, the Adige in Trident 160 t km^{-2} $year^{-1}$, and the Seine as it flows through Paris transports only 17 t km^{-2} $year^{-1}$. Hénin et al. (1954) established that the annual removal of soil unprotected by vegetation in the Durance, Drac, and Severaisse river basins amounts to 450 t ha^{-1} whereas only 15 t ha^{-1} are eroded if the soil is partially protected by vegetation. The Rhône river with the largest flow of water removes about 22 million tons of silt annually.

In addition to water erosion, intense wind erosion also occurs in France, mainly in the maritime areas of the Gascony.

Of particular note among works of soil conservation is that carried out under a law of 1860 concerning mountain reforestation (reboisement des montagnes). The greatest credit for erosion control work is due to Demontzey (1882) who is considered as the founder of the principle of protective afforestation, turfing, and the control of torrents and ravines. In France, sand dunes were stabilized as early as the 16th century. From historical photographs, two pictures are presented showing the stabilization of soil severely damaged by water erosion (Fig. 155).

Fig. 155. One of the first operations carried out in the Savoy Alps following the 1860 law on the turfing ▶ and afforestation of land. a – construction of dry barriers, b – the same territory two years later, before completion of the stabilization work (France). (The author's collection of photographs.)

Fig. 156. Protection against rain erosion, snow erosion and avalanches in the Rio Oragon basin, Pyrenees. (Photo F. Rainer.)

Spain

Spain, like France, is dominated by Atlantic and Mediterranean climatic influences which are associated with intense levels of erosion. According to Giordano (1956), 18.9 million ha of soil (33.7% of the country's total land area of 505,000 km²) are severely affected by erosion or completely destroyed. 24.4 million ha of agricultural land also suffer from very severe erosion, as do 6.5 million ha of forests which include only 2.4 million ha of productive forests (as recorded in 1953). Exceptionally intense erosion occurs in the southeastern part of the country.

In Spain, as in other European countries where there is severe erosion and destruction of the permanent vegetation, especially the forests, much emphasis is placed on stabilization. According to Austin (1958), it is planned to replant 5.7 million ha of forests in Spain over the next hundred years. Figure 156 shows a system of erosion control measures in Pyrenees avalanche area.

Portugal

Portugal also experiences relatively extreme conditions, but the unfavourable effects of the Mediterranean climate are limited to a small area. Nevertheless, Portugal has 1,575 million ha of wasteland representing 17.7% of the total of 8,777 million ha of wasteland on the European continent. Agricultural and forest land is also affected by severe and very severe levels of erosion (Haden-Guest et al. 1956). The adverse effects of Portugal's colonization of Saint Helena are well-known, the problems arising with the introduction of goat breeding in 1502; after 200 years, the lush green landscape was converted into barren, heavily eroded slopes.

In *other European countries* erosion is less widely distributed, principally because climatic conditions are more favourable. In *Great Britain* (244,100 km²), the area affected by erosion is estimated to be 5.6 million ha (22.9%), and includes mostly moors and uplands. About 40,000 farmers cultivate this land contributing only 4% to the agricultural income of the country in 1956, although the amount of land in question represents 25% of all cultivated land.

In *Scotland,* the origin of the well-known stone fields called screes is most probably connected with precipitation erosion. An unusual phenomenon found here is the wind erosion of moorland in the Highlands as a result of sheep grazing; paths are trodden across the ground by the animals and this facilitates wind erosion. Soil erosion in *Ireland* is also associated with large-scale soil management methods. Unfortunately, data on soil erosion are not available.

In other countries which are under the influence of the Atlantic climate, erosion does not cause such serious problems, although its harmful effects cannot be overlooked. In *Belgium* (30,513 km²), slight or moderate precipitation erosion occurs on small areas of agricultural land, and in the Netherlands (36,900 km^2) wind erosion occurs. Wind erosion is of greater importance in *Denmark* on the Jutland peninsula, the total area of land damaged by this form of erosion being about 60,000 ha (Fig. 157). *Luxembourg* (2,586 km^2) has no serious problems with erosion.

Norway (324,000 km^2) experiences some degree of erosion; two thirds of its land area is situated above the upper timberline where the soil receives little protection in the form of vegetation. High rates of removal of eroded material occur in those rivers fed by glaciers in the catchment area, the specific removal of debris varying from 1,000 to 2,200 t km^{-2} $year^{-1}$. Steep slopes unsuitable for agricultural land suffer from sheet erosion and also from gully erosion and landslides in some places. In the periglacial region, the soil is heavily damaged by solifluction and avalanches.

In *Sweden* also (449,750 km^2), solifluction, landslides, and on tilled land precipitation erosion – all cause some damage. Wind erosion occurs on Gotland and in some areas of southern Sweden; Hornsmann (1958) reported that 35,000 ha of soil were damaged by wind erosion.

Finland (337,000 km^2) does not escape erosion either. Although the country has a high percentage of forested land and a large area of lakes and wetlands, the northern regions contain mountain ranges where glacial rivers carry a high proportion of debris – up to 2,500 t km^{-2} $year^{-1}$ (Kukal 1964).

In the polar zones of Scandinavia and on Spitzbergen with its prevalent tundra, forests cannot exist and the chief aid to soil protection in these parts of the world is the conservation of the tundra vegetation. Without it, solifluction and wind erosion occur leading to a *periglacial desert*. These are the only desert formations in Europe.

Erosion in *Iceland* (103,000 km^2) is likewise characterized by the removal of material by glacial rivers (up to 3,200 t km^{-2} $year^{-1}$). In addition, there are specific forms of erosion caused by the activity of the 40 to 50 active volcanos. According to Fournier (1972), about half of the subarctic meadows was destroyed by excessive grazing during the 1,100 years of settlement, thus creating conditions for intense erosion. As well as this, volcanic output of ash and lava destroys the vegetation especially in the surrounding areas of active volcanos. In some eruptions a considerable amount of ice is melted, thus releasing a large quantity of water which destroys the vegetation and intensifies erosion. The thaw is also associated with mud flows, and crevices develop. The volcanic ash, glaciovolcanic material and alluvia are rapidly eroded by wind action.

This description of erosion phenomena in Europe concludes with the German Federal Republic and the German Democratic Republic.

In the *German Federal Republic* (248,000 km^2) erosion mostly affects the mountain regions where there are fluvioglacial sediments and areas of loess and sand deposits. According to Richter (1965), the area of territory damaged by

Fig. 157. Soil protection against wind erosion on the Jutland peninsula, Denmark. (Photo M. Holubčík.)

erosion varies from 7.5% (Schleswig-Holstein) to 75.2% (Marsberg) of the cultivated land, and represents 45.6% of the total land surface. Erosion is most severe in the springtime, and besides precipitation erosion, intense wind erosion also occurs in the GFR. Furthermore, local underground erosion has been observed in the loess regions. Interesting results have been obtained relating to soil transport on account of ploughing; Wandel (1950) and Richter (1965) described a step which developed at the lower border of a forest. The step, which had been formed as the result of 55 to 130 years of ploughing had an inclination of 2 to 7°, and was 50 to 160 cm high, thus representing a soil removal of 46 to 145 m^3 ha^{-1} $year^{-1}$. Wandel (1950) concludes that after the disappearance of terraces and shrub growth on the loess slopes of the Rhine valley, soil erosion set in and the bedrock was exposed within 110 years. Kuron (1956) found that on runoff plots with gradients of 9 to 11%, the mean annual soil loss on unprotected loess soil amounted to about 40 t ha^{-1}.

Table 102. Area of soil affected by erosion in the GDR [ha]

Severity	Erosion			
	By water	By wind	By water and wind	Total
High	374,060	100,630	28,220	502,910
Moderate	1,256,055	105,700	215,590	2,377,345
Total	1,630,115	1,006,330	243,810	2,880,255

Fig. 158. Stabilization of a littoral dune with palisade fences in the GDR. (Photo D. Zachar.)

In the *German Democratic Republic* (108,178 km^2) cultivated land over a considerable area is in need of protection from erosion (Flegel 1959) (Table 102). This means that soil susceptible to erosion and requiring erosion control measures extends over 28,800 km^2 (26.6% of the country), including 5,029 km^2 (4.7%) requiring protection from severe erosion. The total area of tilled land is 53,000 km^2. According to Schultze (1952), 14% of the area of Thüringen is damaged by erosion. Hempel (1951, 1954) reported that the average intensity of soil erosion on loess with a gradient of 2 to 4° during 1,000 years of cultivation has been about 10 m^3 ha^{-1} year^{-1}. During this period a layer of 1,000 to 1,200 mm of soil has been removed. The intensity of wind erosion in littoral areas has been discussed by Hornsmañn (1958); he mentions, inter alia, that after deforestation of the Prussian Haff, the communities of Karwaiten, Lattenwalde and Pillkopen were buried under sand. Shifting dunes forced the inhabitants to move out of the neighbourhood. An example of the stabilization of littoral sands is shown in Fig. 158.

From this survey of soil erosion in Europe, it may be seen that although this continent is the least threatened by erosion, erosion can nevertheless attain very dangerous and harmful proportions. In general, erosion increases with the extent to which the climate is continental in type, and with the aggressivity of the climate; it also increases with the dissection of the relief and soil erodibility. By the application of appropriate measures erosion can be reduced to harmless levels.

5.2 Asia

Asia has much higher levels of erosion and a much larger area of eroded land than Europe, and in absolute terms, it has the highest rates of erosive earth transport of all the continents. Only some brief comments on the distribution of erosion and use of erosion control measures can be made here, according to the availability of data.

Asian part of the USSR

Almost all forms of soil erosion occur in the Asian part of the USSR. They have already been briefly described in the account of erosion in the European part of the USSR. Generally the intensity of erosion and the area occupied by eroded land increase from North to South and from East to West.

High levels of erosion occur in the republics of *Georgia, Armenia* and *Azerbaijan* (Fig. 159). Alekperov (1957) established that the rivers of Azerbaijan carry away about 48 million tons of silt annualy from an area of 85,700 km^2, compared with only 47 million tons from the plains of the European part of the USSR where only about 40% of the land is eroded. This means that the specific runoff of silt in

Fig. 159. Retention barrier (height 9 m, length 100 m) as a protection against mud flows in Georgian SSR. (Photo S. G. Totashvili.)

Table 103. Amount of eroded land in the mountain regions of Central Asia

Republic	Mountain areas [ha]	Amount and per cent of eroded land		Per cent of eroded land in each erosion category			
		[ha]	%	I	II	III	IV
Kirghiz SSR*	18,800	5,996	35.7	41	27	27	6
Uzbekistan	6,300	5,806	87.9	35	25	40	0
Tadzhikistan	12,800	9,109	71.2	5	10	71	14
Turkmenistan	5,070	4,921	97.2	27	41	28	5
Total	43,000	25,838	60.0	24	23	45	7

*Survey of eroded land not yet completed.

Azerbaijan is about 1,400 t km^{-2} and locally the average rate of soil removal is as much as 500 t ha^{-1} $year^{-1}$. Large soil losses and severe erosion are caused by mud flows which occur in all the mountain ranges bordering on the southern frontiers of the USSR.

The highest levels of erosion occur in the republics of Central Asia, namely in *Kirghiz SSR, Uzbekistan, Tadzhikistan* and *Turkmenistan,* where 36 to 97% of the total land surface of the mountain regions is damaged by erosion. According to Kocherga (1965), about 25,833 million ha (60% of the 43 million ha of mountain country) are affected by erosion. More detailed data are given in Table 103.

Detailed analyses of soil erosion in the various republics show that almost the entire territory is affected by erosion or the deposition of the products of erosion. Thus in Tadzhikistan, for example, slightly eroded or uneroded land occupies only 2.3% of the territory. Another 26.3% of the territory is threatened by precipitation erosion of different degrees of seriousness and by debris flows, 4.2% by irrigation erosion, 14.9% by slight erosion, 2.6% by moderate and severe erosion, and 49% of the territory is rocky, with either accumulations of rock fragments and the remains of washed soil (33.5%), or glaciers and snow containing fractions of rock and debris (15.4%). The remaining 0.7% is accounted for by lakes and rivers (Yakutilov et al. 1963).

In like manner, the mountain regions of Turkmenistan are composed of some land that is eroded by precipitation erosion (8.3%), some by irrigation erosion (10.3%), some by precipitation and wind erosion (19.2%), and some by wind erosion (59.3%) and gully erosion (1.6%) (Stepanov 1966).

Intense erosion was found to occur in the basin of the Amu Darya which delivers about 96.7 million tons of silt per year to the Aral Sea and has an average turbidity of 3,590 g m^{-3}. However in some of the rivers in the upper part of the catchment area turbidities of up to 11,700 g m^{-3} have been recorded, the specific flow rates of

Table 104. Erosion removal and soil deposition on irrigated land, according to Presnyakova (in Mikhaĭlov 1949)

Soil	Crop	Slope inclination	Soil movement [m^3 ha^{-1}]		
			Removal	Deposition	Loss
Alluvium carbonate soil,	Beet	5°	30.8	8.8	21.0
first grade of wash	Beet	11°	73.0	18.2	54.8
	Beet	17°	231.3	35.0	195.3
Brown soil, second grade	Maize	20°	388.8	66.8	322.0
of wash	Cabbage	24°	845.0	157.5	687.5

transported matter reaching exceptionally high values of up to 50,000 t km^{-2}. The total discharge of deposits in the *Amu Darya* in the city of Kerki is 228.67 million tons, corresponding to a specific runoff of these deposits of 1,008.2 t km^{-2} $year^{-1}$. A similar situation is observed in the basin of the *Syr Darya.* The percentage of forested land in both river basins is 2.4.

High levels of erosion also occur on agricultural land; in fact in this part of the world the highest known rates of soil removal are recorded. According to Mikhaĭlov (1959), the rate of removal was as high as 3,200 m^3 ha^{-1} in some regions. High values were also registered on irrigated soils (Table 104).

Exceptionally high losses and the transport of very large volumes of material occur in *mud flows* (selevye potoki) in which rates of transport of up to 90,000 t km^{-2} (of the basin's active area) have been established. The latter represents an average reduction in the thickness of the topsoil of 90 mm per year. The total volume of eroded, displaced earth produced in one downpour can be as much as 10 million m^3 (Gagoshidze 1949, Kocherga 1965). A total of 2,245 debris flows (mud torrents) has been recorded in Central Asia.

Besides surface erosion, underground erosion also does a lot damage mainly in semiarid regions.

Whereas in the mountain regions of these republics severe general erosion occurs, wind is the main erosion factor in the plains. This is particularly valid for Turkmenistan and Uzbekistan with their extensive desert and semidesert regions (the *Kara Kum* and *Kizil Kum deserts*).

Of the total land surface of 1,237,000 km^2, mountain regions account for 430,000 km^2 (according to other sources 444,000 km^2), the remainder of the territory suffering mainly from wind erosion (Fig. 160). The percentage of forested land in the different republics varies from 1.0 to 3.7.

Intense wind erosion also seriously damages the territory of the *Kazakhstan Republic,* although conditions there are relatively more favourable. According to

Fig. 160. Territory affected by wind erosion in the Kara Kum desert (USSR). Saxaul shrubs are the predominant plant. (Photo D. Zachar.)

Dzhanpeĭsov (1977), 76.0 million ha (28% of the total land surface of 272.5 million ha in the republic) are afflicted by wind erosion. Sand (mostly desert) occupies 24.1 million ha. The total area of soil threatened by precipitation erosion is 16.8 million ha (6.2% of the total land surface of the republic), which includes 7.0 million ha affected by moderate and severe erosion. Erosion is intensified by ploughing of virgin land and by grazing of domestic animals. In the republic of Kazakhstan are situated the deserts of *Muyunkum, Sari Ishikotrau* and *Taukum.*

In the Asian part of the *Russian Soviet Federated Socialist Republic,* which takes in the largest part of the land surface of the USSR, natural and economic conditions are not conducive to erosion. A cold climate, a high percentage of forests and a sparse population are the reasons for the negligible to very slight levels of erosion. However a survey of land utilization has shown that in these regions there is a high risk of potential erosion (Mizerov 1966); average values for sheet erosion in the Far East and Sakhalin were found to vary from 1.0 to 8.4 mm, the thickness of deposits varied from 0.4 to 2.1 mm, and average soil losses were about 40 t ha^{-1}. For gully erosion the average increment in the depth to which the soil was eroded was found to be 40 cm $year^{-1}$, and the average annual soil loss varied from 45 to 337 t ha^{-1}. On easily erodible loam the average soil loss was

4.5 mm year^{-1} on gradients of 1 to 3°. Considerable damage is also caused by underground, river, and wind erosion.

China

Erosion processes on other parts of the Asian continent are much more intense. Whereas in Europe and the USSR the proportion of land put to economic use was 22% in 1960 (FAO), in Asia the corresponding proportion was estimated at 54%. Probably the most intense erosion of the soil occurs in loess regions characterized by a very broken relief. In these areas erosion reaches catastrophic proportions, although the total area of land affected by erosion in China, according to Messines (1958), is estimated to be 160 million ha – only 16.67% of the total 9,597 million km^2 land area of China. The cultivated part of the eroded land, 58 million ha, is situated in the basin of the *Yellow River* (Huang Ho). Loess deposits extend over an area of about 60 million ha.

Both the extent and intensity of soil erosion in the loess regions are indicated by data on the flow of silt in the Yellow River (which was named after the yellow colour of the water resulting from the very large content of transported erosion products). The turbidity of the Yellow River is estimated by Muranov (1957) to be 34 kg m^{-3}, by Parde (1954) and Messines (1958) to be 44 kg m^{-3}. During floods, however, turbidities of 430 to 460 kg m^3 have been measured. Messines reported even higher turbidities in various tributaries of the Yellow River, and when the turbidity reached about 560 kg m^{-3} the flow became plastic and its surface became wrinkled. It is interesting to note that at turbidities of 510 to 560 kg m^{-3} the discharge velocity of the watercourse was observed to increase. Figure 124 shows a detail from a slope which has been completely destroyed by erosion and which forms an important source of silt in the Shansi Province. An illustration of erosion in one of the most intensively eroded regions is given in Fig. 31. For the sake of comparison, the turbidity of the Nile is 1 kg m^{-3}, that of the Amu Darya is 4 kg m^{-3}, that of the Colorado River is 10 kg m^{-3} and the turbidity of the Yellow River is around 40 kg m^{-3}.

The total flow of alluviates in the basin of the Yellow River is variously estimated to be 920 million m^3, 1,380 million tons (Messines 1958), or even 2,000 million tons (Parde 1954). In some years even higher values have been recorded; Messines (1958) gives 2,643 million tons as the highest figure. Taking the area of the basin as 74.5 million ha, and the average flow of solid matter as 1,380 million tons, the mean specific runoff of silt into the Yellow River turns out to be 1,850 t km^{-2}. For some of the tributaries this value increases to 3,360 t km^{-2}. In the central part of the basin of the Yellow River the values are still higher, which in view of the size of the watercourse represents a world maximum. Thus according to Chinese investigators, the Wei Ho river carries an average of 5,800 t km^{-2} and the Lo Ho river carries 7,190 ton km^{-2}. If one disregards the area of the old delta of the Yellow

Fig. 161. The first step in the prevention of erosion is the stabilization of river courses, the control of floods and gullies. The system of earth dams shown in the picture prevents vertical erosion, and therefore also inhibits the erosion of adjacent slopes. (The author's collection of photographs.)

River (250,000 km^2) in which there is no erosion, a specific runoff of 4,000 t km^{-2} is obtained, and in the most eroded parts of the basin this may be as much as 10,000 t km^{-2}.

In order to make a comparison between conditions in China and those of central Europe, the author has selected the Bey Ho river which has a basin somewhat smaller than the area of Czechoslovakia (110,000 km^2), but has a silt flow of 386 million tons per annum and a specific silt runoff of 3,515 ton km^{-2} (half the figure for the Lo Ho river). An approximate calculation of the situation in Czechoslovakia, taking the mean runoff of silt as 5.0 million tons, suggests that the specific runoff of silt is about 38.5 t km^{-2}, which is 18.6 times smaller than the value for the Lo Ho river.

The high erodibility of Chinese loess and the advanced state of recent erosion make land use and soil management extremely difficult, even if comprehensive erosion control measures are applied where necessary in all parts of the river basins (Figs. 161–164).

The reason for the high erodibility of loess is the large proportion of fine earth fractions together with the relatively low content of binding colloidal materials. A higher cohesiveness of loess is possible only if there is a sufficiently high content

Fig. 162. Erosion control bars for inhibition of erosion on slopes on which agricultural crops are cultivated. (The author's collection of photographs.)

of calcium which, of course, is easily leached out. The most intense erosion occurs in those regions in which loess alternates with sand or gravel deposits. Erosion starts with the formation of gullies which, in already loosened and weathered loess, grow rapidly into sizable rills with perpendicular walls. Further weathering, the

Fig. 163. System of erosion control measures in the mountain regions of Shensi Province (China). (The author's collection of photographs.) ▶

Fig. 164. Systems of erosion control measures on slopes disturbed by intermittent vertical erosion (Kansu Province, China). (The author's collection of photographs.)

washing of banks, gullies and ravines, and the detachment of entire blocks of loess continuously furnish material for removal by precipitation water and river water, so that locally, mud torrents may develop. The deeper the loess, the greater is the possibility for the formation of ever larger gullies which can grow into ravines or canyons 200 to 300 m deep. The gully heads dig upward and backward into the basin and damage an increasing amount of land. Underground erosion is common, and water erosion often acts in conjunction with wind erosion, especially on ridges and projections of the relief.

Some indirect consequences of soil erosion occur in the lower lying plains of China, which are essentially the huge deltas of streams. (The delta of the Yellow River has an area of about 25 million ha.) The large amounts of transported material give rise to a continuous deposition and raising of the river-beds, thereby decreasing the dimension of the discharge profile and increasing the risk of the water *overflowing* the banks. This danger increases during *floods* when the turbidity of the water also increases and the greatest changes occur in the river-bed. It has therefore been necessary to build or raise the level of earth dams along the banks of the lower reaches of the river in order to prevent *flooding* at high water levels, which occur in spring when snow melts in the mountains, and in summer from July to October when heavy rains arrive. The bed of the Yellow River and that of the Yangtze Kiang river are continuously being raised, so that the total thickness of deposits in some sections is now some tens of metres. In such a situation, the action of rodents burrowing into dikes or the bursting of dikes at high water levels imposes a terrible threat on the surrounding territory; in the past, flooding has caused the deaths of millions of people in the basin of the Yellow River alone. For example, in the floods of 1887 and 1889 about a million people died on each occasion and about 7 million people of this fertile plain were made homeless. Some villages were buried by layers of deposits up to 3 m deep. In 1938, when the Yellow River changed its course for the seventh time, about 890,000 people died and 12.5 million people were made homeless (Muranov 1957). Altogether more than 200 million people live in the basin of the Yellow River.

It is clear that in this situation, when the serviceable life of reservoirs which could be used to regulate the flow of water from the catchment area is very short; agricultural methods of erosion control and the establishment of forests in the river basins are of the greatest importance. In the People's Republic of China, protective measures have been put into practice, over 14 million ha since 1953, and 10.3 million ha of new forests have been established.

In the northern parts of China there are extensive areas of desert and semidesert where wind erosion is the main destructive force. Chief among these areas are the Taklamakan Desert, the Alasan Desert, and the *Gobi Desert* which spread into the Mongolian People's Republic. Intense wind erosion also occurs in the Junggar basin, on several plateaus, and in almost the whole of the loess region. In the mountains of Tibet there are well expressed forms of pluviofluvial, nival and cryogenic soil destruction, all of which are aggravated by animal grazing.

Much soil erosion also occurs outside the loess regions and is caused mainly by the deep dissection of the relief and the aggressiveness of the climate. Thus, for example, the second largest river, the Yangtze Kiang with its catchment area of 1,960,000 km^2 and average water flow per annum of 720 km^3, carries a mean quantity of 1 milliard tons of silt representing a silt flow of 514 t km^{-2} (Ma Szi 1955). About 3 milliard tons of silt are carried away annually from the catchment areas (total area 2,750,000 km^2) of the two greatest rivers Huang Ho and Yangtze Kiang alone. The latter areas represent 28.2% of the land area of China.

Japan

Conditions here are less conductive to erosion, mainly owing to careful maintenance of the forests. Of the total surface area of 372,000 km^2, forests cover 249,520 km^2, equivalent to 68% of the country's territory (China's forests cover only 10% of the land). Despite this, about 315,000 ha of forest soil are damaged by

Fig. 165. System of erosion control measures on land destroyed by water erosion, including an earth dam, fences, grass sowing and tree planting (from Conservation in Japan, 1968).

Fig. 166. Stabilization of river sands in the basin of the Ganges river in India. (Photo V. Čermák.)

erosion. Heavy damage from erosion occurs in the more arid areas where the vegetation is sparse and the poorly protected land called *geya* occupies an estimated area of 1.829 million ha (5% of the country's territory) (Cummings 1956, Ogihara 1952). Severe erosion takes place among the mountain massifs where 40% of the rock is of volcanic origin; the erosion is caused by heavy precipitation exceeding 2,000 mm per annum. Thus the total area of eroded land is 280,000 km^2, and the flow of erosion products in the rivers is 70 million tons representing a mean specific runoff of 188 t km^{-2} $year^{-1}$. In the southern regions the specific runoff of

Fig. 167. Severely eroded territory in eastern Azerbaijan, Iran. (Photo F. Papánek.) ▶

Fig. 168. Promontory of the Zagros Mountains; the rocks here have a high content of salt which is washed out into the salt desert. (Photo F. Papánek.)

silt exceeds 1,000 t km^{-2} (Katayama 1968). Wind erosion occurs chiefly in littoral areas. Measures to conserve the soil are widely used (Fig. 165).

Very intense, and even catastrophic erosion occurs in, or threatens almost all the countries of southern Asia, except in the *Korean republics* where the specific runoff of silt is similar to that in Japan. Higher rates of silt runoff occur in regions with a high annual rainfall, such as the catchment areas of the Mekong, Salwin, and Iravadi rivers, western and southern *India, Burma, Thailand, Cambodia, Bangladesh, Vietnam,* and other countries. However, most of the territory which has a high rainfall also has a high percentage of forests (e.g. the *Philippines* have 72.5% forest cover), and therefore erosion is not of great importance in these regions, although the cultivation of crops (mainly rice) in the heavily broken terrain meets with considerable difficulties.

With decreasing rainfall and increasing temperature, rates of erosion increase sharply. The dependence of erosion on the climate and on the nature of the relief is clearly visible in India which has very severe erosion of all forms, including wind erosion (Fig. 166). The weakening of the protective cover of vegetation owing to deteriorating climatic conditions is apparent in *Mongolia,* where desert forms of erosion (the *Gobi Desert*) are on the increase. The main reasons for the acceleration of erosion in these countries are, of course, grazing, burning of the vegetation and ploughing up virgin land, etc. Examples of these processes are also found in *Sri Lanka* where eroded land is called *patronas.*

Erosion has very damaging effects in *Pakistan, Afghanistan, Iraq, Iran, Turkey,* and all the countries of *Asia Minor.* The total land area of these countries amounts to 7.84 million km^2, which is approximately the same as the area of the Sahara. Although the proportion of desert landscape in Asia Minor is less than that of the Sahara, it is unsurpassed with respect to the very extreme conditions prevailing in the *Indo-African desert region.* The remainder of the territory belongs to the Pontic-Central Asian and Mediterranean regions which are also characterized by extreme conditions. Only a very small part of the southwestern corner of the *Arabian Peninsula,* some mountain regions of Turkey, and the plains around the *Indus* and *Firat* are not affected, or only very little affected by erosion. In the mountain stretches of these rivers there is much water erosion, and the Firat (Euphrates), for example, as it passes through Turkish territory, accounts for a specific runoff of alluviates amounting to 516 t km^{-2} year^{-1}. In the mountain regions the annual silt runoff is much greater and reaches a few thousand tons per km^2. For this reason, high rates of deposition have buried fertile alluvial soil and blocked irrigation systems.

Under such conditions any interference with the balance of nature, has had serious consequences, and thus the history of the peoples living on this territory, like those inhabiting the basin of the Yellow River, is marked by the fight against erosion or catastrophes resulting from the man-made acceleration of erosion processes. The consequences of the deforestation of *Lebanon,* of the destruction of permanent vegetation by goats, sheep, and fire, and of deforestation in the basin of the *Euphrates* and *Tigris* rivers, etc., are all well known. During four thousand to five thousand years of "civilization", extensive wasteland scourged by erosion and drought has been created. As evidence of the intense erosion that has taken place in this period, it is pointed out that the delta of the Euphrates and Tigris rivers has advanced about 200 km (about 50 m annually) into the Persian Gulf over the whole of this period.

Very extreme conditions prevail in Iran which forms a gigantic closed basin surrounded by mountain ranges (Fig. 167). Here salt is washed out of Miocene layers in great quantities, giving origin to a large *salt desert* (Fig. 168) which includes the *Dashti-e-Kevir, Dashti-e-Lut* and other desert areas. Erosion reaches serious levels in the mountains and on both the borders and central areas of the plateau. In addition, there is considerable underground erosion and wind erosion, the latter occurring even on loamy soils. Erosion is generally widespread, especially in *Baluchistan*. Furon (1947) links the desolateness of Seistan in the centre of the Iranian plateau (about 500,000 km^2 area) with the Mongolian invasions which occurred between 1200 and 1222 under the Mongol conqueror Genghis Khan. Historical research has shown that where there now exists an extensive area destroyed by wind erosion, there was fertile land before the Mongolian invasion.

Severe soil erosion also occurs in Afghanistan and Pakistan where deserts, semideserts, and areas of mountain erosion can be found. The *Dashte Margo* and *Dashte Tahlab deserts* form closed basins with larger or smaller influxes of water into swamp areas or eroded, mostly saline land. In the alluvial territory of the Indus, there are sand dunes stretching as far as India. In all three countries (Iran, Afghanistan and Pakistan) the characteristic climate of desert, semidesert and steppe prevails, and extends in fact over a large zone straddling the African and Asian continents where it gives rise to severe erosion.

Rates of water erosion are given in a report by Starmans (1970), who found that in 17 *Indo-Pakistani* rivers the specific flow of silt ranged from 277 to 6,596 t km^{-2} year^{-1}. This range of values is consistent with flow rates occurring under similar conditions elsewhere. The highest rates of removal were recorded in the Ravi River which deposits sediment averaging 53.5 million tons over an area of some 1,200 km^2. After analyzing erosion in the basins of Asian and African rivers, Starmans concluded that there is a close relationship between aridity and rates of erosion, the connecting factor being the growth of a protective cover of vegetation (Fig. 169).

The degree of soil damage caused by erosion can also be assessed from the proportion of the land that is covered by forests and the proportion of cultivated soil in these countries. Thus according to statistical data, the proportions of forested land in *Afghanistan, Pakistan, Iran, Iraq, Jordan* and *Israel* are 1.7% (1.2%), 2.8% (2.4%), 11.6% (9.8%), 3.5% (2.0%), 0.8%, and 0.25%, respectively. (The figures in brackets represent proportions of accessible forest.) There are relatively high proportions of forest land in *Syria* (7.4%), *Turkey* (13.8%), and *Cyprus* (18.5%). However the areas of productive forest in Syria, Turkey and Cyprus represent only 0.3, 1.4 and 14.8%, respectively. Other forested areas are damaged by grazing and erosion. Protective afforestation in these countries is an urgent requirement; for example in Turkey alone, about 10.5 million ha of severely eroded soil are in urgent need of afforestation.

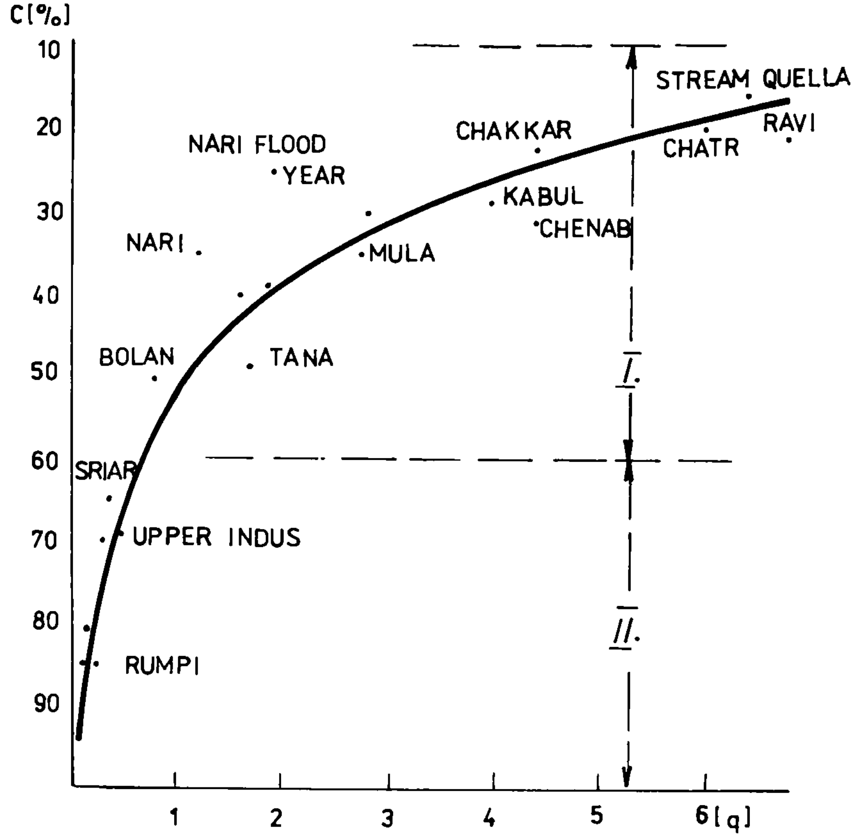


Fig. 169. Relationship between cover, C (C – percentage of catchment area adequately protected), and soil erosion (1,000 t km^{-2} year^{-1}), derived from the characteristics of rivers in Africa and Asia.

Thus a large proportion of the soil in these countries is damaged by erosion. In Pakistan wind erosion and water erosion together spoil over 75% of the total soil surface, and in Iran more than 80% of the soil is in a state of deterioration because of erosion and sedimentation, the soil on 15.0 million ha of land being entirely destroyed. A similar situation prevails in Iraq, 1.6 million ha being affected by water erosion, and 2.4 million ha by wind erosion (Buringh 1960). Large areas are impaired by irrigation erosion and sedimentation. It is estimated that in Iran accelerated erosion has been going on for 5 to 6 thousand years. 37% of the land in Israel is either eroded or so far unused.

In general, the rate of erosion, and more especially the extent to which erosion has progressed in these countries are very considerable as a consequence of the long history of erosion in each case. All forms of erosion occur from typical mountains forms, to irrigation forms, underground erosion and wind erosion. The overall period of continuous erosion over a large part of the Asian continent is four to five times longer than in Europe (with the exception of the *Mediterranean region*), and therefore completely exposed rock formations, badlands, sandy deserts, rocks, salt and takirs, deposits several tens of metres deep, can all be found on the Asian continent bearing testimony to the acceleration of erosion by man's activities. Many forms of anthropogenic erosion can now be distinguished from natural erosion only with difficulty. The fact remains that erosion and drought in this part of the world are the main causes of low soil fertility and poverty among the people, and they can be controlled only by the systematic and long-term application of effective measures.

5.3 Africa

Africa is the second largest continent. It forms part of the Old World, and of all the continents it is the most affected by soil erosion as a result of much of the land mass being equatorial. Ascending air currents in the equatorial belt give rise to a high rainfall, whereas in the tropics descending air currents and high pressures result in a lack of rainfall. Huge *deserts* and *semideserts* are constantly extending their margins in the latter regions. In the northern zone around the tropic of Cancer the largest desert in the world, the *Sahara* (Fig. 170), covers about 7,780,000 km^2, and in the south around the tropic of Capricorn there are the Namib and Kalahari deserts. Of Africa's total land area of 29,820,000 km^2 (the total area of the islands of the continent is 620,000 km^2) deserts occupy 40%; the equatorial forests occupy about 7% (2.0 million km^2), about one third of Africa on either side of the equator is accounted for by the savanna with its long drought spells, and the remainder consists of steppe and semidesert country and the subtropical belts in the north and south.

Under these conditions erosion inevitably spreads rapidly in almost all regions, assisted by increasingly extensive land utilization. Much of the evidence for the acceleration of soil erosion comes from *North Africa* and recently from *South Africa* also. The most dangerous practices include the extensive rearing of sheep, goats and other domestic animals, such as camels, which graze the poor vegetation to ground level. There is an estimated stock of 125 million sheep, 100 million cattle and 6 million camels. Further damage is caused by the burning of the vegetation and ploughing of the land. Because of these factors, the area of desert and

Fig. 170. Sahara territory modelled by wind and water erosion on the route from El Getch to In Salah. Catastrophic incidental flooding causes greater loss of lives than drought. (Photo J. Haleš.)

semidesert and that of the sahel regions increase, soil being laid waste at a very rapid rate. It is estimated that the Sahara desert advances southward at the rate of about 1 km per annum (Furon 1947). Some FAO investigators have established that the 6,000 km long Sahara border has been shifting southward by 1.5 to 10 km year^{-1} during the last 50 years owing to increasing cultivation of the land in the border regions.

The restoration of soil fertility and the control of erosion present complex problems because of the extreme natural conditions, and many attempts to solve these problems onesidedly have failed. Thus 57% of the uncultivated soil in Africa is impaired by erosion, according to the FAO.

Fig. 171. Bedrock completely denuded by erosion on slopes once covered by forests. (Photo D. Zachar.)

The difficulties of erosion and drought are perhaps most pronounced in *Egypt, Lybia, Tunisia, Algeria, Morocco,* and in the lower lying areas of *Northern Sudan, Chad, Nigeria, Mali* and *Mauritania.* Only in small parts of these territories there is some possibility of improving ecological conditions by irrigation from rivers of which the upper part of the catchment area receives abundant precipitation. The possibility of irrigation schemes arises mainly in the Sudan, Egypt, parts of Algeria, Morocco, Mali and Mauritania. However, soil irrigation tends to cause salination in these regions with an ensuing decrease in the fertility of the land, which then proceeds to be eroded more vigorously than ever. In addition water reservoirs and irrigation systems tend to become choked by the products of erosion, the construc-

Fig. 172. View of severely eroded territory on the southern slopes of the Atlas Mountains, and a valley of the Sous river partially used for the cultivation of agricultural crops becoming filled with gravel deposits (Morocco). (Photo D. Zachar.)

tion of dams alters the underground water regime, enrichment by fertilizing substances diminishes, and therefore projects aimed at the control of drought and erosion need to be carefully considered from all angles.

The problems of irrigation in Egypt, Lybia, and other countries are well known. Furon reports that in *Algeria,* 10 large water reservoirs with a capacity of 700 million m^3 were constructed for the irrigation of 100,000 ha of land. Some reservoirs become choked by deposits after 20 years, while others had a serviceable life of 50 years. The average turbidity of small rivers called "oued" was 1 m^3 of silt per 50 m^3 of water, but this increased during floods to 1 m^3 of silt per 25 m^3 of water, i.e. from 80 to 160 kg m^{-3}; these are high values for such small watercourses. In the Sig reservoir alone over 800,000 m^3 of silt were deposited annually. The main problem in Algeria, however, is wind erosion in the desert and semidesert regions; in the border zones of these areas wind erosion is growing in severity.

There are similar problems in *Morocco*. Thus, for example, Margat (1954) observed that in the Tafilelt enclosed area of palm groves (2,500 ha), the annual volume of deposits loosened by erosion in the basin of the Zis river was about 1.0 million m^3, representing a specific silt runoff of 125 t km^{-2} $year^{-1}$. (Precipitation in this region, which is adjacent to the Sahara, is 100 to 200 mm $year^{-1}$; the basin of the Zis river occupies an area of 8,000 km^2.) The average turbidity of the water was 50 kg m^{-3}, and the annual increment in the amount of alluvial deposits in the palm groves was found to be about 100 m^3 ha^{-1}. Very severe erosion occurs in the Atlas Mountains where there are examples of total soil destruction caused by accelerated erosion (Figs. 171, 172).

Tixeront and Berkaloff (1954) observed that the most intense precipitation erosion in *Tunisia* occurs in regions in which the annual precipitation is between 300 and 700 mm, and where the specific runoff of silt is about 1,500 t km^{-2} $year^{-1}$ (Figs. 173, 174). However in some regions of Central Africa and the Atlas Mountains, the specific runoff of silt exceeds 3,000 t km^{-2} $year^{-1}$. Although figures for the intensity of precipitation erosion in Africa vary a great deal, Africa has the most intensely modified landscape of all the continents.

In Somalia it is estimated that about 90% of the agricultural land is threatened by erosion, mainly gully erosion. Dubreuil and Vuillaume (1970) observed that in the tropical semiarid region of Nigeria, precipitation erosion begins on gradients of as little as 1%. The following values for the specific runoff of silt from an area of 16 km^2 were obtained (average rainfall 400 mm $year^{-1}$; average annual temperature 28.5°C):

Soil type	Gradient	Crop	Specific runoff of silt [t km^{-2} $year^{-1}$]
Sand	1	Millet	650 to 215
Clay	3	Sparse grass	575 to 710
Loam	12	Savanna, thornbush	750 to 1,230
Clay	3	75 % tilled land	1,210 to 1,320
Loam	12	15 % tilled land	1,450 to 1,950

These data refer to small catchment areas and represent a transitional situation between measurements obtained from small plots and data relating to larger basins. In any case, they provide evidence of the high potential soil erosion in this

Fig. 173. Severely eroded territory in the semiarid region in Tunis. (Photo D. Zachar.)

"southern-sahel" region. It can be seen from the above figures that ploughing 75% of land with a gentle gradient of 3% and ploughing 15% of land on steeper gradients of around 12% both result in an approximate doubling of the rate of soil removal, this observation being relevant to catchment areas which are little protected by vegetation.

Goujon (1968) prepared a survey of precipitation erosion in Africa and *Madagascar,* and reported that in Nanisan (Madagascar), sheet erosion of the soil ranged from 2.7 to 26.5 mm. Measurements of rates of soil removal in Senegal (expressed in t ha^{-1} $year^{-1}$) were as follows:

Fig. 174. Antierosive rock terraces in the semiarid region in Tunis. (Photo D. Zachar.)

Gradient	Crop under cultivation		
	Peanuts	Rice	Sorghum
1 %	3.1	6.4	7.0
1.5 %	4.3	9.5	14.2

These are relatively high values, nevertheless at gradients of 3 to 4%, rates of erosion increased sharply to as much as 50 t ha^{-1} $year^{-1}$ on land under sown crops.

In Adiopadoumé (the lower part of the Republic of the Ivory Coast), the mean annual removal of soil from barren ground on a moderately steep slope was found to be about 130 t ha^{-1}. And finally, near the Alaotra lake on Madagascar, an average rate of soil removal of 59 t ha^{-1} $year^{-1}$ was observed on cropped land with a gradient of 7%. Again these data provide evidence of the very high levels of potential precipitation erosion.

The calculations of Goujon relating to erosion in some parts of Madagascar, show that the precipitation index, R, ranges from 288 to 509, and the soil index, K, ranges from 0.05 on slightly eroded soil to 0.6 on very severely eroded soil. For a moderately eroded soil with a K index of between 0.1 and 0.2, the potential erosion on a 9% slope 22.12 m long would range from 28.8 to 101.8 t ha^{-1} $year^{-1}$. In an extreme case the potential erosion would reach 305 t ha^{-1} $year^{-1}$, a very high value indeed.

High levels of potential erosion are also implied in data obtained in *South Africa* by Hudson (1971), who reported that soil removal from within a tobacco crop growing on a 6% gradient was 18 t ha^{-1} in the first year, 45 t ha^{-1} in the second year, and about 70 t ha^{-1} in the third year (Fig. 175). On barren land the annual removal of soil over a ten-year period was 126.57 t ha^{-1}.

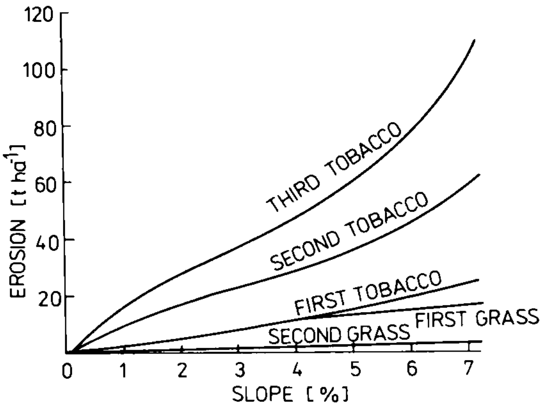


Fig. 175. Influence of continuous crop cultivation at different angles of slope inclination (according to Hudson, 1971); with successive cropping of root and tuber plants erosion increases, whereas with grass, erosion decreases.

Fournier (1972) mentions an average annual soil removal of 28.42 t ha^{-1} from barren land in the Glen locality of South Africa; this was observed to take place on a gradient of 5%, and therefore seems to indicate a relatively lower level of soil erodibility.

From the data presented thus far on the intensity and distribution of precipitation erosion on the Africa continent, it is clear that *soil erosion is a serious problem in this part of the world.* Where erosion losses are small it is usually because already severely eroded soils are present with a protective *layer of gravel or stone* (e.g. the *hamada* and *serir* desert formations, the rock soils of mountains and plateaus, or the salt-stabilized soils of the dried out dunes of *shotts* and *takyrs,* etc.) (Fig. 176).

Fig. 176. The wind-eroded floor of a shott is extremely unfavourable for the existence of any plant life (Chott al Tadjlat, Tunisia). (Photo D. Zachar.)

Exceptionally high rates of wind erosion are attained in regions of shifting sands (Fig. 177) which drift tens of metres per annum in the direction of the prevailing wind with finer soil particles being blown over much larger distances. Almost every year the *fall out* of sand from Africa is recorded in central Europe.

Considerable damage to soil is also caused by river erosion; African rivers with their large volumes of water flow have relatively large amounts of energy which can be diverted into erosion activity.

Fig. 177. The binding of sand dunes in the Great Eastern Erg in Tunisia is necessary for the continued existence of an oasis (Taghit). (Photo D. Zachar.)

5.4 Australia and the islands of the Pacific

Australia is the youngest of the continents and is one of the most severely affected by erosion. According to FAO data of 1960, the lands of the Pacific region have the largest proportion of unused soil, agricultural land representing 11%, forest land 5%, and other types of land 84%. The world average figure for unused land is 42%. The greater part of the continent is covered by scrub, heath, mulga, semiarid and desert steppe, desert and savanna. The arid region represents 43% and the semiarid region 20% of Australian territory.

In barely two centuries the few million inhabitants who settled on the continent managed to accelerate erosion in some places to catastrophic levels. The main cause of this was grazing and the burning and removal of vegetation; the breading of about 100 million sheep and 10 million cattle, and the growth in the populations of rabbits and kangaroos gave rise to serious degradation of the soil in a short time. The rabbits were mainly responsible for accelerating *wind erosion,* whereas de-forestation brought about an acceleration of surface runoff, a more rapid formation of *gullies,* and faster *desiccation of the soil* with greater accumulation of *salt* in the surface layers followed by the *underground washing of the soil.* There is therefore much evidence of man's acceleration of erosion in Australia, and plenty of indication, too, that it can be controlled effectively as well.

Tunnel erosion occurs mainly in the soils of southeastern Australia, these soils having a "pathod zone" of whitish clay which forms the relic of a leached horizon of fossil soil originating under a partially tropical climate. These formations occur mostly in New South Wales and Victoria, where tunnel erosion develops into gully erosion and badlands develop in some places. The intensity of sheet erosion is so high in some regions that the whole of the upper horizon may be removed in one rainstorm. Water erosion is also important in the more arid areas where storms are apt to be sudden and torrential. Wind and water erosion accelerate both once the "basal cover" of vegetation has been removed by grazing animals.

The severity of soil erosion in Australia is shown on an erosion map (Fig. 178) prepared by Herriot while working for the FAO in 1953. The map has been modified by the author into an easily reproducible form, and several adjustments have been made by incorporating data from the Soil Conservation Authority, which also provided the author with a generous amount of photographic documentation.

The general map gives an account of actual water erosion and wind erosion, and the degree of destruction of the basal cover as the cause of accelerated erosion. In general, the intensity of erosion decreases on the edge of the continent owing to effective soil protection, while land is deteriorating in the interior.

Vegetation is more severely damaged in the western regions of Australia where there are excessive numbers of wild donkeys and hill kangaroos. Heavy destruction of plant life also occurs in *northwestern Australia* in the Victoria basin. In *central Australia* erosion is apparently more severe than is indicated on the map; according to the information of the Soil Conservation Authority, intense forms of water and wind erosion occur in this district.

There is also severe destruction of vegetation in *Queensland, New South Wales,* where 25% of the land is damaged by erosion, and in *southern Australia* where wind erosion predominates. In *western Australia* in the wheat belt basin, and over a part of the Margin Plateau, water erosion prevails. Water erosion also occurs in severe forms in the eastern parts of Australia, namely the coastal belt, the mountains and tablelands, the slopes and the plains. Severe water erosion takes place in the coastal belt including *Hunter Valley,* the mountains and tablelands of *New England,* the *Central Uplands,* the *Southern Tablelands,* the *Australian Alps,* the *Central Victorian Hills* and *Tasmania.* In the tablelands and hills which account for a considerable part of the terrain, erosion affects 40% of the tableland area and 52% of the hill area, the northern slopes in both cases receiving the most damage, as might be expected in the southern hemisphere. On the plains, 40% of the

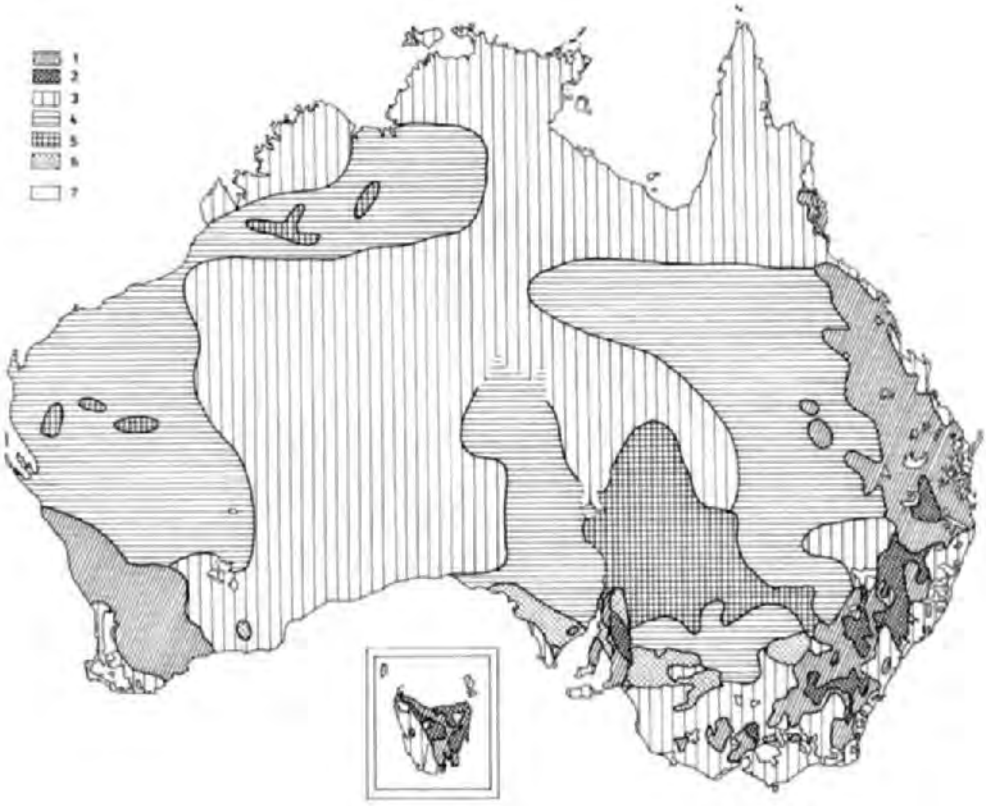


Fig. 178. Erosion in Australia. 1 – slight water erosion, 2 – severe water erosion, 3 – terrain with slightly disturbed vegetation, but with clear erosion forms, 4 – terrain with moderately disturbed vegetation, 5 – terrain with severely disturbed vegetation, 6 – wind erosion, 7 – land protected by forest.

territory is denuded by precipitation erosion and 30% by wind erosion. In 54 basins a major part of the land is endangered by severe or very severe erosion.

The most intense wind erosion occurs in the *Great Sandy Desert,* the *Gibson Desert,* the *Victoria Desert,* the *Simpson Desert,* and in the adjacent regions with predominantly sandy soils in the southwestern, northern and central regions of Australia.

The problems of soil erosion are vividly illustrated in Figs. 19 and 20 (the consequences of the burning of forest vegetation), Figs. 25 and 122 (aggressive

Fig. 179. Severe soil erosion caused by large-scale ploughing of the soil, and not influenced by the characteristics of the relief in the case (Daling Downs, Queensland). (Photo M. Roberts.) ▶

Fig. 180. Contoured banks used to prevent sheet and gully erosion on cultivated land (north-central Victoria). (By courtesy of Soil Conservation Authority, A.S.)

Fig. 181. Comprehensive soil conservation has resulted in maximal productivity and permanent protection of this land (Inverell, New South Wales). (By courtesy of Soil Conservation Authority, A. S.)

washing of the soil), Fig. 36 (the development of pronounced underground erosion into gully erosion), Fig. 40 (the intensity of wind erosion of the soil) and Fig. 44 (the past effects of wind corrasion). In Figs. 134 and 136 control measures against wind erosion are shown. Also illustrated are accelerated erosion as the result of large-scale ploughing (Fig. 179), erosion control by contour ploughing, retention ditches and the stabilization of collectors by a grass stand both in moderately undulating terrain (Fig. 180) and in more steeply undulating terrain (Fig. 181), a system of measures against sheet and gully erosion (Fig. 181), and finally, the stabilization of a waterway with periodical runoff by means of a grass mattress and the channelling of gullies into a watercourse (Fig. 182).

Oceania, comprising about 10,000 islands and accounting for about 2% of the world's dry land, is dispersed over an area equivalent to about one third of that of the globe, and thus includes within its bounds a great variety of natural conditions. In a similar way the influence of man shows great variety in the region and has differing histories, so that both complete nature reserves and totally devastated and derelict islands are to be found here.

According to Egler (1956) the vegetation has been destroyed right from the earliest settlement by natives who practised extensive burning; in fact the burning of forests and shrub vegetation is still carried out today on some islands. In a second developmental stage, the vegetation was further destroyed by the intro-

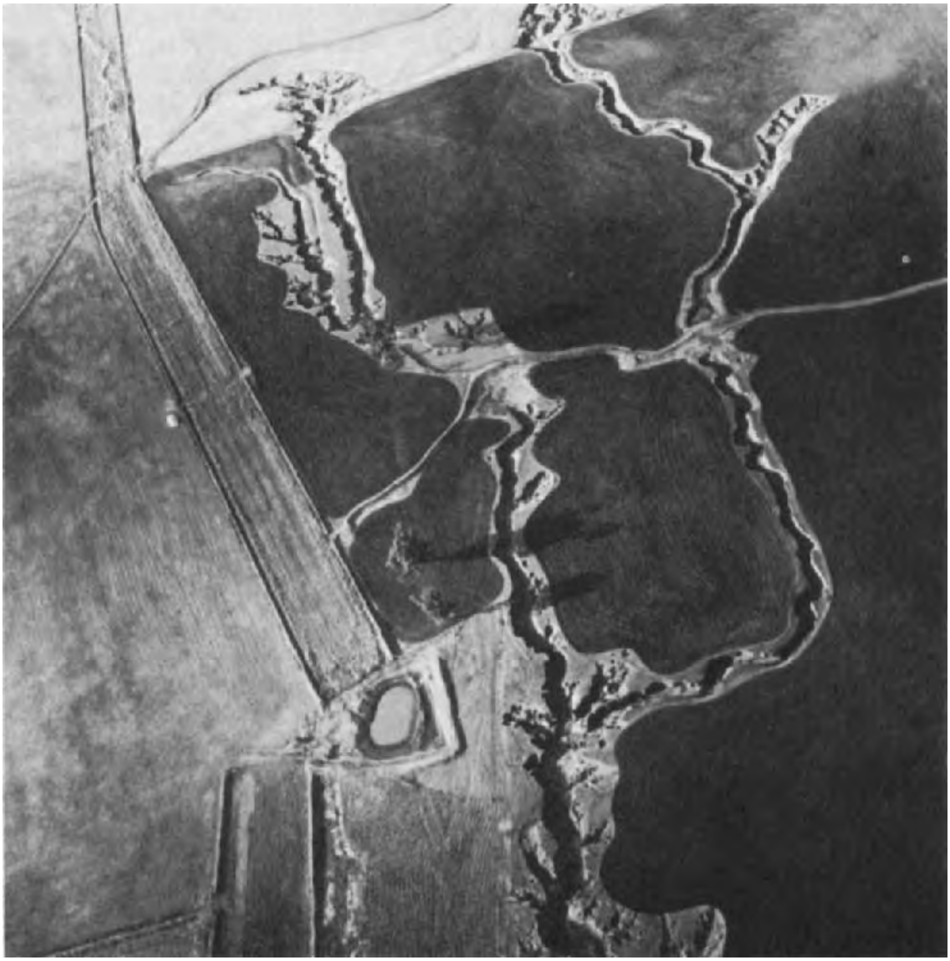


Fig. 182. Comprehensive measures taken against sheet and gully erosion, including contoured banks, grassed water verges and dams. The picture also shows gully growth owing to tunnel erosion (northern Victoria). (By courtesy of Soil Conservation Authority, A.S.)

duction of domestic animals, mainly cattle, sheep, goats and horses. Egler mentions a *Hawaiian island* on which rabbits have completely destroyed the plant communities; in 1903 this island was covered by lush vegetation, but after the introduction and growth of the rabbit population, it was turned into a "desert" and the rabbits died out because of lack of food. The soils of many of the islands were also destroyed by white men who laid forest fires, established plantations, and when the soil was devastated by erosion, left the island.

Classical examples of accelerated erosion occur in *New Zealand,* where the acceleration has been caused by burning and destruction of the plant life over large areas. Enrican and Holloway (1956) reported that during the last one hundred years 16 to 18 million acres (6.5 to 7.3 million ha) of forests in New Zealand have been destroyed, mostly by burning, so that the 30 million acres (12.1 million ha) of forests existing in 1850 were reduced to less than a half of this area, and the consequences were no less than catastrophic. The initially lush growth of the pastures gave way to meagre plant associations which gave insufficient protection to the soil against the forces of erosion. The consequences of such interference with the ecosystem may be seen from the erosion processes portrayed in Figs. 14–16. In recent years far-reaching erosion control has been put into practice on the island.

5.5 North and Central America

The extensive dry lands of the western hemisphere stretch over the two continents of *North* and *South America*. These land masses are connected by a strip of dry land, *Central America,* and are host to a great variety of erosion phenomena. The total land area of the American continent is 42.4 million km^2, North America accounting for 24.5 million km^2 and South America for 18 million km^2. It is the only continent which comprises nearly all the climatic zones of the globe occurring between the two poles. The *Central-American continental bridge* (the *Isthmus of Panama*) has a width of less than 50 km. Whereas North America has a very ragged landscape, that of South America is the least ragged of all.

Natural conditions and the degree of exploitation of natural resources together determine the proportions of forest, agricultural, and other types of land. According to FAO statistical data published in 1960, South America has the highest percentage of forests (54%) and the smallest proportion of agricultural land (23%) of all the continents while Central America has the least amount of forested land (27%) and the most agricultural land (40%). Other types of land account for 33% of the landscape in both North and Central America, and 23% in South America. The proportions of these land categories are, of course, very different in the various countries.

Canada has a larger area than Europe, a small population density, and the largest percentage of forests of any one country. In spite of this the proportion of eroded land, which is increasing northwards mainly on agricultural soils, is relatively high. According to Ropley et al. (1961), areas with different degrees of damage caused by erosion are represented as follows:

Degree of erodedness of the soil	Reduction of crop yields	Proportion of total land area
I	Up to 10%	72.6%
II	From 10% to 35%	27.3%
III	More than 35%	0.1%

Fig. 183. Stabilization of gullies by sowing grass on river terraces. The picture was taken shortly after heavy rains. (By courtesy of Australian News and Information Bureau.)

Erosion is widely distributed on land that is put to intensive agricultural use and is also a feature of cold regions and mountain districts. For example, the territory of the Great Whale River (above the Hudson Bay) is an area affected by wind erosion (Fig. 183), and the glacial valley in the Province of Alberta which is floored with fluvioglacial deposits (Fig. 184) contains very eroded soils of the badland type.

Fig. 184. Territory affected by wind erosion in the Great Whale River basin (above Hudson Bay, Canada). (Photo A. Jahn.)

Much more severe soil erosion occurs in the *United States of America.* During 150 years of colonization, 113.9 million ha of soil have been damaged or totally destroyed by erosion in the USA (Bennett 1955), and there are 313.1 million ha of land that are moderately damaged or endangered by erosion. Thus the amount of eroded agricultural and forest land was estimated to be 427 million ha. Besides this, another 58.5 million ha of unusable land in the mountain regions, deserts, saline and other areas of the USA are affected by natural erosion. Only 283.6 million ha, i.e. about 38% of the total land area of the USA (769.216 million ha), are not affected by accelerated erosion. Out of 427 million ha of erodible land, about 300 million ha are impaired by actual erosion. About 4 million ha of land are excluded from economic utilization every year because of deposition and flooding. Wind erosion damages about 44 million ha of land annually (Stallings 1957).

The distribution of soil erosion according to the *Soil Conservation Service* is shown on the schematic map in Fig. 185.

The greatest intensities of erosion and the largest expanses of eroded land are found on cultivated land, followed in sequence by pastures, then forested land. Stallings (1957) gives the results of detailed investigations organized by the Soil Conservation Service in 1948 in the states of *Arkansas, Louisiana, Oklahoma* and *Texas.* Five classes of eroded soil were distinguished, and it appeared that 49% of

Fig. 185. Fluvioglacial sediments under Saskatchewan Glacier, to the southeast of Mount Athabaska in the Province of Alberta, Canada. (Photo H. E. Malde, U.S. Geological Survey.)

arable land, 38% of the prairies and 20% of forested land were placed within the latter four of these classes. Erosion on forest soil was found to occur mainly on burned areas, clear-felled areas, and in particular, on grazed land supporting a thin and degraded forest stand. 141,637,300 ha of forest have been damaged by grazing in the USA (Papánek 1948), and the area of wasteland increases annually by about 500,000 acres (202,000 ha).

Severe levels of erosion are associated with heavy losses of nutrients into the rivers; according to Bennett (1955), losses of silt exceed 3×10^9 tons annually and the amount of the five principal nutrients contained in this silt is 92,172,300 tons on the basis of the following proportions: 1.55% potassium, 0.15% phosphoric acid, 0.1% nitrogen, 1.56% calcium, and 0.84% magnesium oxide. The phosphorus, potassium and calcium together represent 43,361,000 tons annually. Stallings (1957) calculates a loss of 4×10^9 tons annually, and points out that if the lost nutrients were to be replaced by fertilizers, this would represent (at 1947 prices) 4.3 milliard dollars for the nitrogen and phosphorus alone. To include potassium would increase this amount to 7.75 milliard dollars. In addition, the cost of damage in the form of soil deterioration amounts to about 750 million dollars on arable land, 180 million dollars on pasture land and 25 million dollars on forest land, totalling almost 1 milliard dollars annually. Further losses amounting to some millions of dollars are incurred because of crop destruction, floods, and the costs of soil conservation schemes, etc. For example, the damage caused by floods and deposits alone is estimated to be 557 million dollars However these large financial sums surely cannot express the long-term consequences of erosion, in particular the permanent loss of soil and the effects of reduced fertility which will be felt by many generations to come.

An estimate of the specific runoff of silt in the USA will provide a basis for comparing the intensity of erosion with that in other countries. Taking the amount of material carried away as 4×10^9 t year^{-1}, and the land area of the USA as 7,692,160 km^2, a value of 520 t km^{-2} is obtained. The average yearly removal from eroded land is, of course, much higher. According to Bennett and Lowdermilk (1938), the average level of soil removal is about 61 t ha^{-1} (115 t ha^{-1} maize crops, 15 t ha^{-1} for wheat crops in a rotation, and 55 t ha^{-1} for cotton). Although these data are approximations, they give a good picture of intensities of erosion in the USA.

Exceptionally high levels of erosion are confirmed by the very turbid waters of some *North American rivers,* which are not far behind China's Yellow River in the league of turbidity values. Parde (1954) reported some extraordinarily high values, e.g., 77.6 kg m^{-3} for the *Little Colorado,* and 144 kg m^{-3} for the *Rio Puerco* (a tributary of the *Rio Grande*), the maximum turbidity of the latter reaching 680 kg m^{-3}.

Figures for soil wash are available which illustrate rates of soil removal in the USA. At the La Grosse Experimental Station in the southwestern part of *Wisconsin,* measurements were made on silt loam of the Fayette type on a gradient of 16% (9°05′), and values for the average annual soil removal were derived as follows: 0.22 t ha^{-1} year^{-1} under a herbaceous stand, 62.27 t ha^{-1} year^{-1} under an agricultural crop rotation, 250.20 t ha^{-1} year^{-1} under maize, and 427.84 t ha^{-1} year^{-1} on fallow land. Taking each of these situations in the above sequence, an 18 cm deep layer of soil would be carried away in 10,000 years, 36 years, 9 years, and 5 years, respectively. These results, according to Bennett, are typical of the southwestern part of Wisconsin, the southern part of *Minnesota,* the northern part of *Illinois* and the whole of *Iowa.*

Very high rates of wind erosion in some of the states of the *Great Plains* are reported by Lyles (1975). Supposed soil losses due to wind erosion on unprotected fields range from 38 to 310 t acre^{-1} year^{-1} according to the degree of soil resistance or susceptibility to erosion. Erosion of this severity produces marked reductions in harvest yields.

The following example points out how damaging losses of extreme proportions occur. One of the hurricanes of 1934 blew away over 500 million tons of soil in the states of Nebraska and Dakota, the soil being eroded in some places down to a dept of 70 cm. Large movements of soil are also well known in the states of *Arizona, Idaho, Nevada, New Mexico* and *Utah* (a combined territory about three times as large as France) where desert formations are gradually on the increase and wind erosion is a serious problem.

Because of the ever increasing requirements of the human population for food, the problem of soil erosion in the USA increases in magnitude as it becomes necessary to extend the area of arable land in use. According to Heady and Timmons (1975) the proportions of land in the USA are as follows:

Use of land	Million of acres	% of total land
Arable land	472	20.9
Grassland, pasture and prairie	604	26.9
Forest land	723	31.9
Special uses of land	178	7.8
Other land uses	287	12.7
Total land	2,264	100.0

In future the acreage of arable land is expected to be increased by 264 million acres to 736 million acres at the expense of forests (124 million acres), deserts and swamps, etc. (117 million acres), and pasture land and prairies (23 million acres), thus greatly increasing the danger of erosion.

The consequences of this extension of the area of arable land may be seen from the effects of erosion on the soil of newly ploughed virgin land (Grant 1975). These results were obtained by the Soil Conservation Service in 1973–1974 and show that on new arable land which previously comprised 8.9 million acres of pasture, woodland, and other types of land, 4 million acres of this suffered losses greater than 4 t acre^{-1} year^{-1}. In the season 1973–1974 60 million tons of soil were removed altogether (13 million tons by wind erosion and 47 million tons by water erosion) from the entire area of 8.9 million acres. In some localities rates of water erosion and wind erosion of up to 140 t acre^{-1} year^{-1} and 40 t acre^{-1} year^{-1} were observed, respectively. The average soil removal on 1.6 million ha of tilled land is 37 t ha^{-1} year^{-1} – a rapid rate of erosion leading to an early deterioration in soil fertility and the disintegration of the land.

The consequences of erosion in the USA are therefore of a serious nature involving not only soil and soil fertility, but also a large number of water reservoirs which, like watercourses in the USA, tend to become silted up with large amounts of deposits. Thus the problems of irrigation and other activities involving the use of water are closely linked with erosion. In arid and semiarid regions especially, soil degradation is combined with salination, desiccation, underground erosion, and a reduction in the protective cover of vegetation, all of which finally results in devastation of the soil mantle.

According to Glymph and Storay, 1.23×10^9 m^3 year^{-1} (about 1.85×10^9 t year^{-1}) of deposits settle in the reservoirs of the USA, and this, according to

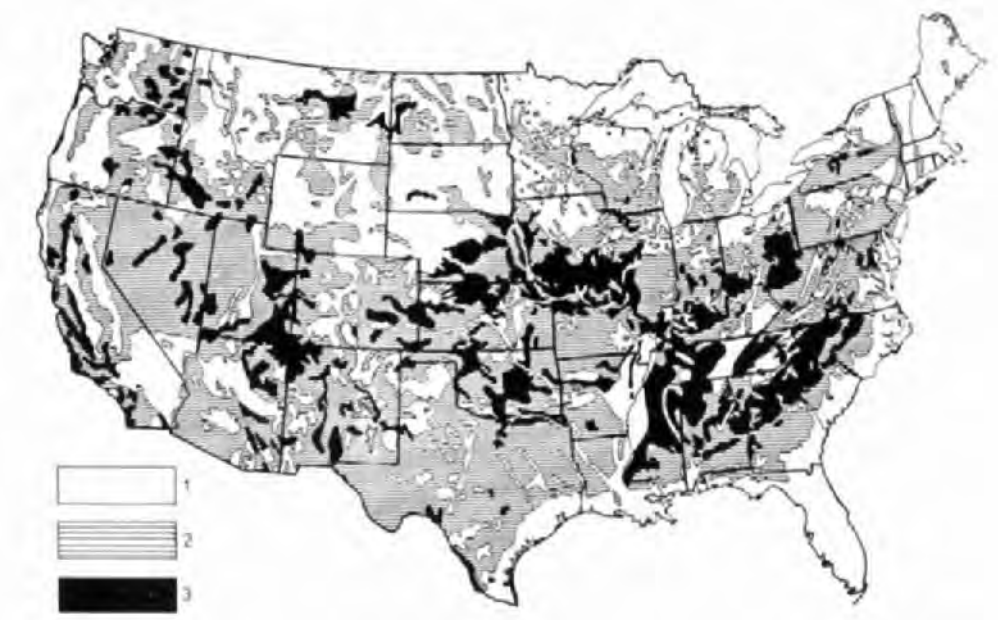
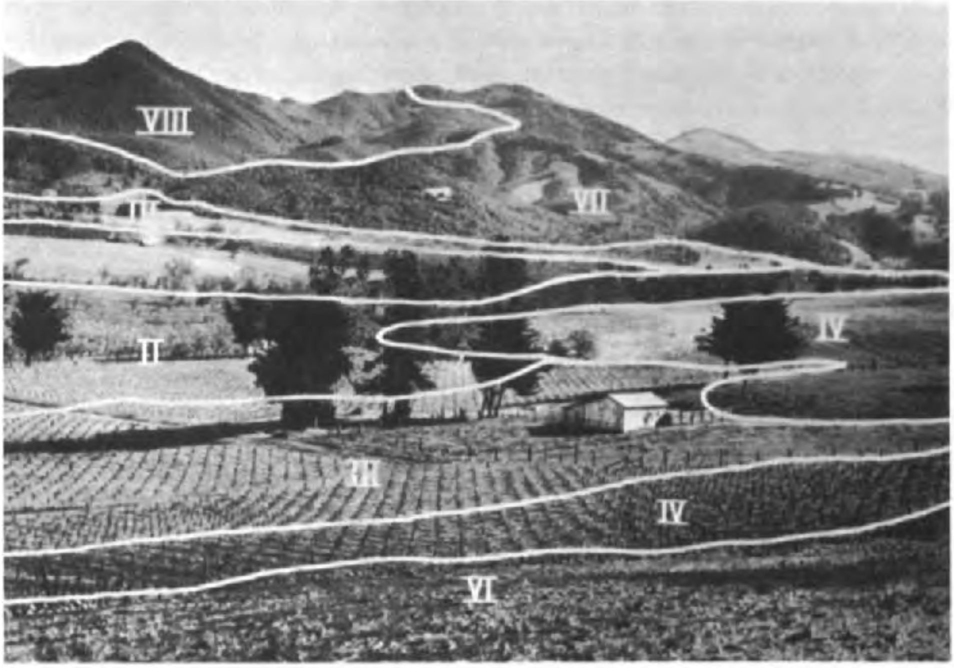


Fig. 186. General map of the distribution of erosion on cultivated soil in the USA. A – slight or none, B – moderate – 25 to 75% of topsoil lost, may have some gullies, C – severe – more than 75% of topsoil lost, may have numerous or deep gullies. Includes severe geological erosion in parts of low rainfall area. Many small areas could not be shown at this scale.

1,948 calculations, represents a financial loss of about 50 million dollars as the cost of the reduction in volume of the reservoirs. Dendy (1968) published the results of an investigation into the reduced capacities of 968 state reservoirs in the USA, and came to the conclusion that the original capacity of the reservoirs (75.54×10^9 m^3) was reduced by 2.97×10^9 m^3 (3.94%) per year. The smaller the reservoir, the faster was the rate of silting up and the shorter was its serviceable life. The average decrease in reservoir capacity was 0.72% and according to the size of reservoir the decrease in capacity ranged from 0.1% for reservoirs exceeding 1.23×10^9 m^3 in volume to 2.2% for reservoirs up to 12,335 m^3 volume. The record time for the complete silting up of a reservoir was 16.1 years.

Fig. 187. Land utilization categories. Suitability for cultivation: I – good soil-management sufficient to maintain soil quality, II – moderate conservation practices required, III – intensive conservation required with infrequent cultivation, IV – perennial vegetation required with infrequent cultivation. Land unsuitable for cultivation (pasture, grassland for hay-making, woodland, wildlife reserves): V – no restriction in use, VI – moderate restrictions in use, VII – severe restrictions in use, VIII – best suited for wildlife conservation and recreation. ▶

Fig. 188. Contoured hill cultivation with the spreading of fertilizer in progress. (By courtesy of Soil Conservation Service, A.S.)


VIII
VII
II
IV
III
IV
VI

Because of the situation outlined in the foregoing, considerable attention is given to soil erosion in the USA, most of the burden of this falling on the Soil Conservation Service. Basic principles of soil conservation are detailed for eight different categories of land use (Fig. 186), and universal equations are used for the calculation of water and wind erosion, as mentioned in previous chapters. It appears that the level set for tolerable erosion represent a rate of soil removal greater in magnitude than the rate of formation of soil and supply of nutrients in the form of fertilizers, no account being taken of the consumption of nutrients by crops and other forms of soil deterioration such as occur under the influence of snow water, chemical erosion, and arable erosion. Nevertheless, large sums are spent on soil conservation, and yet it seems that the measures taken have not been sufficient to reduce erosion losses to satisfactory levels; total soil losses due to water and wind erosion show a continuing tendency to increase in some regions.

The most favoured conservation measures are crop rotation using crops of differing resistance to erosion, contour ploughing, alternation of crops in narrow strips following the direction of the contours, hill cropping (Fig. 187), construction of retention ditches, terracing, mulching, the ploughing-in of organic material and waste, fertilization, furrowing on pastures, protective afforestation, and a number of other measures aimed at controlling water erosion and precipitation erosion. Recently more radical methods of soil protection, including the levelling of gullies and eroded land with heavy machinery, have come into use (Fig. 188).

In the *Latin American countries* also erosion is a serious problem. Schultze (1952) reported that of the total area of 56.6 million ha of agricultural land in 12 republics of this region, 18.2 million ha were affected by serious erosion. A further 10.1 million ha are excluded from productive use.

The largest country in *Central America* (total area about 2.5 million km^2 including about 22% forest) is Mexico which with an area of 1.97 million km^2 (including 13.1% forest), has relatively unfavourable natural conditions from the point of view of erosion. The effects of human intervention are also particularly damaging, and tend to increase with the increasing size of the population. It is estimated that about 25 million ha are occupied by cultivated land, the remaining area being made up by devastated grazing land and wasteland. Kunkel (1963) reported that devastation of the countryside proceeded rapidly and if agricultural policy was not changed, the entire landscape would be degraded in a period of hundred years. In recent times about 3 million ha of land have been devastated by grazing, and other forms of misuse.

Less serious is the situation in the countries to the South where there are more favourable climatic conditions for the preservation of the natural vegetation and the maintenance of its protective effect. *Guatemala* has 51.2%, *El Salvador* 76.8%, *Honduras* 43.8%, *Nicaragua* 47.1%, *Costa Rica* 78.3% and *Panama* 69.8%, of the land area forested (Tseplaev 1961). Although the larger part of the territory in these countries is mountainous, the plant life is favoured by high precipitation and warm temperatures, and the resulting rich growth protects the soil well. Furon (1947) described the formation of mud flows in Guatemala and Costa Rica which the author here refers to as aquatic soil flow, or aquasolifluction. This phenomenon was observed on loamy and clay soils where the rainfall was from 2,000 to 5,000 mm. An exception within this group of countries is El Salvador where conditions allow the cultivation of crops which give little protection to the soil, and where erosion is therefore fairly widespread.

As in other regions, intense erosion occurs also on the islands of Central America. There are serious soil erosion problems in *Cuba* which in the past was almost entirely covered by forests. In particular, erosion causes difficulties with the cultivation of sugar cane, root and tuber crops, specialized crops and pasture land.

Forns (1957) observed that the mean annual removal of soil from tilled land (gradient 4%, precipitation 600 to 1,250 mm year^{-1}) varied from 31 to 36 t ha^{-1}, and was 16 t ha^{-1} under maize, 6 t ha^{-1} under rye, and 0.25 t ha^{-1} where grass was sown.

In *Jamaica* the larger part of the land is eroded. Catastrophic levels of erosion are known to occur in the *Dominican Republic,* but perhaps the most harmful effects of erosion are to be found in *Puerto Rico* where in 1940, 41.8% of the land were observed to be severely eroded, 26.8% were moderately eroded, and only 19.1% of the land suffered losses of less than 25% of the topsoil. In 1960 the proportion of severely eroded land reached 48% of the island's total area (Kunkel 1963).

5.6 South America

As already mentioned, *South America* is the continent with the largest proportion of forested land (54%); it also has a relatively small proportion of unused land (23%), and a small area of agricultural land (23%) (FAO, 1960 in Dregne 1978). Even under these conditions, however, the specific runoff of silt exceeds the world average. According to Lopatin (1950), 148 t km^{-2} year^{-1} are transported off this continent, whereas the world average is 134 t km^{-2} year^{-1}. The probable reason for this is that although large areas of the territory are still inaccessible and the population of the continent is still relatively small, the present large-scale management methods are causing a gradual acceleration of erosion which is reaching dangerous proportions in some regions. As on other continents, extremes of landscape type can be found, e.g. *French Guiana* has forests covering 96% of its territory, while *Uruguay* has only 3% of forested land. Thus there are great differences in erosion activity in these countries.

Severe erosion occurs in the northern countries of South America *(Colombia, Ecuador* and *Venezuela),* where the burning of forests, the establishment of extensive grazing grounds, and the raising of agricultural crops in plantations have

Table 105. Land areas in Peru affected by different rates of erosion (Low 1967)

Erosion [$t\ km^{-2}\ year^{-1}$]	Area affected [km^2]	Per cent of total area
0—1,000	216,040	16.88
1,000—1,500	670,620	52.38
1,500—2,000	215,540	16.85
2,000—3,000	95,400	7.45
3,000—4,000	34,580	2.70
4,000—5,000	31,700	2.48
5,000—7,000	16,120	1.26
Total	1,280,000	100.00

been widespread. Erosion is accelerating over extensive areas of *Brazil,* which is a large country (8,464,200 km^2) with only about 50 million inhabitants; a large part of the territory still remaining uncultivated. Severe erosion can be observed on the large coffee plantations where the soil quickly deteriorates. Destructive erosion, as elsewhere, increases with the dryness of the climate to produce the semidesert regions of northeastern Brazil where the erosive forces of water and wind are accelerated by the grazing of cattle. At the other extreme are the erosion phenomena of the hot, moist São Paolo region, where mountain erosion forms combined with aquatic soil flow occur (Furon 1947).

Extensive areas of agricultural land are damaged by erosion in *Peru, Bolivia, Paraguay, Chile, Argentina,* and other countries. The causes and conditions of erosion are similar in these countries and arise mainly from misuse of the land, and practices which work against natural conditions. As an example we may take the basin of the *Paraná River,* the waters of which come from southern Brazil, Paraguay, Uruguay and Argentina. Of the 800 million tons of silt which arrive in the estuary, 80% are discharged by one tributary. The sedimentation of silt produced by accelerated erosion has caused the delta of the Paraná River to advance 46 m annually during the period 1973–1979, and 84 m annually during the period 1900–1964 (Horning 1970).

An interesting feature of this part of the world is the existence of erosion control terraces built by the original Inca population which thus managed to make use of soil on very steep slopes. According to Bennett (1939), the terraces in some cases were as high as 50 feet, the spaces behind the holding walls being filled with soil manually. The terraces were irrigated with water supplied from sources which were often several kilometres away. The water flowed from terrace to terrace in stone channels and collected in basins at the bottom. Even today, thousands of hectares of soil are cultivated on Inca terraces and form the greater part of the agricultural land in many regions. In the valleys the soil used to be protected from floods by an ingenious system of walls; the bottom of the valley was divided into broad terraces

Fig. 189. Levelling of severely eroded territory as a result of cultivation with heavy machinery. a – before, b – after operations. (Photo R. J. Svacina.)

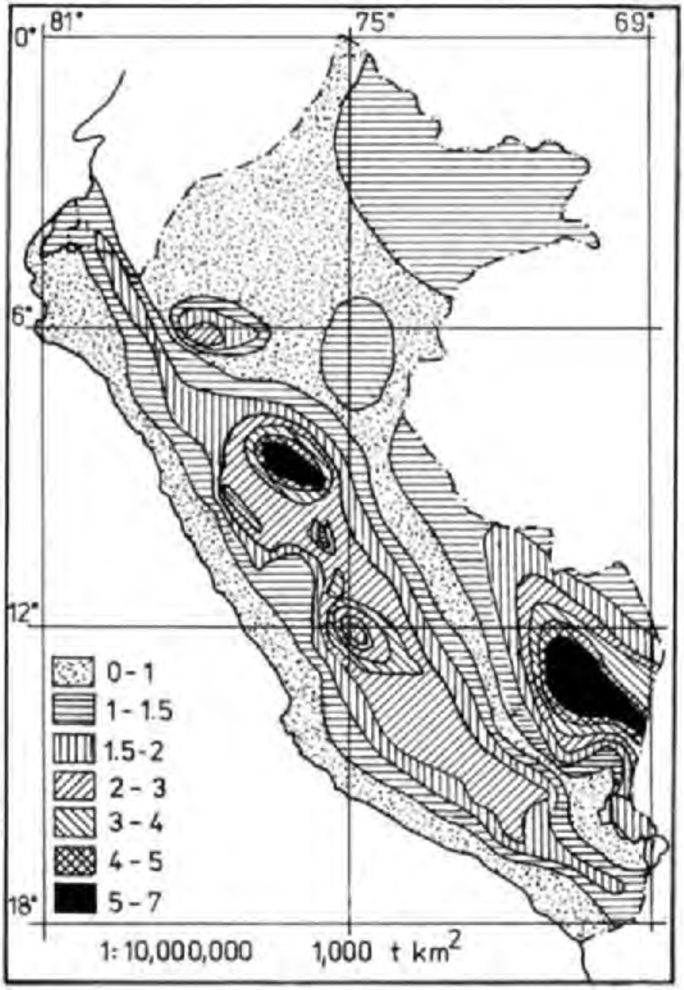


Fig. 190. Map of water erosion in Peru (according to F. K. Low, 1967).

sustained by stone walls, and the soil needed for filling in behind the walls was often brought from far afield. According to information handed down by word of mouth, soil for the Inca and Cuzco gardens was brought from Quito, about 700 miles away. This kind of agriculture dates back to pre-Incan times and its exact origin is unknown. Many of these practices were abolished by the subsequent peoples of the area.

Low (1967) established that erosion intensities in Peru may be as high as 7,000 t km^{-2} $year^{-1}$ (Table 105, Fig. 189). According to Low's calculations, the total removal of erosion products amounts to 1.9 milliard t km^{-2} and the specific runoff of these products is 1,500 t km^{-2} $year^{-1}$ – an exceptionally high rate of erosion for conditions in that country.

Fig. 191. Polygenic sediments in the alpine regions of the Andes (Argentina) are an indication of intense erosion in the basin of the Horcones River. (Photo F. Kele.)

In South America, as on other continents, erosion spreads across an ever increasing area, as the expanses of cultivated land and human interference with nature, especially with the indigenous vegetation, gradually increase. This is particularly true of regions with desert, semidesert, or steppe climatic conditions, of the coastal regions of the Pacific Ocean, of the mountain regions of the *Andes* (Fig. 190) and *Brazilian* and *Guiana Highlands*, and of the northeastern Brazilian states of *Ceará, Rio Grande do Norte, Paraíba, Pernambuco* and *Alagoas* (total area, about 540,000 km^2). In the southern regions of South America, erosion processes are further boosted by the high content of salt in the rocks and soil (Fig. 191).

5.7 Global assessment

In making a comprehensive assessment of erosion phenomena, it is necessary to take the following facts into account:

– The main parameter of the action and effect of erosion on the soil is the intensity of erosion, which expresses the soil loss from unit area of land per unit time.

– Erosion is accelerated on economically exploited territory, but from the point of view of land use in general, it is important to understand the erosion process in regions of extreme climatic conditions where the soil is unprotected by vegetation, and also in regions with favourable conditions where the natural vegetation adequately protects the soil.

– Old landscapes which have been continuously exploited for a few thousand years, the soil having long been removed so that further erosion can only be slight, should be included in the category of the most severely damaged land types.

– The latter category also should include land on which there is a high degree of erosion acceleration and a serious danger of severe land deterioration on account of erosion.

Thus it is not possible to find uniform criteria based on existing information for a comprehensive expression of soil erosion, and such an expression would not in any case be sufficiently accurate to warrant the use of these criteria in practice. Therefore this last chapter will be restricted to a short comment on the various types of erosion.

A *classification of the Earth's land surface* into geographical zones which are distinguishable on the basis of the predominant type of erosion that occurs, is given in Table 106 (after Maksimovich 1955).

Table 106. Areas of geographical zones

Zone	Surface	
	[mill. km^2]	Per cent of total land
Polar and alpine ice	16.08	10.8
Tundra	5.90	4.0
Forest	18.08	12.1
Steppe and forest steppe	32.39	21.8
Deserts and semideserts	27.21	18.3
Tropics and subtropics	21.12	14.2
Mountains	22.62	15.2
Alluvium	4.26	2.9
Continental waters (excluding the Caspian Sea)	0.97	0.7
Total	148.63	100.0

As mentioned in the previous chapter, precipitation erosion occurs predominantly in mountain regions (15.2% of the land surface), in forests (12.1%), and on susceptible soils in the tropics and subtropics (14.2%); erosion and accumulation processes may also be harmful on alluvia (3.2%). In addition, pronounced water erosion occurs in the forest steppes (about 10%) and potentially in the tundra also. On the other hand, wind erosion prevails in the deserts and semideserts (18.3%), and in the steppe regions (11.8%). Wind erosion also occurs on alluvia (1.0%), in littoral zones, and to some extent in the tundra, and other regions. In total, water erosion is a common feature of about 50% of the Earth's land surface; wind erosion is the predominant form on about 34% and fluviologlacial erosion on about 16%. Large areas are affected by several forms of erosion together.

5.7.1 Water erosion

The intensity of water erosion may be assessed by the methods outlined in Chapters 2, 3 and 4. Of the many methods on which a global assessment has been based, the principal are the hydrological and hydroclimatological methods which rely on data on the flow of silt and bedload, or data on rain erosivity and the aggressivity of the climate.

Silt flow was studied comprehensively by Lopatin (1950) and Maksimovich (1955), who gave values for mechanical erosion, chemical erosion, and degradation for the whole of the Earth's land surface (Table 107).

Thus the overall average specific runoff of suspended and dissolved matter is 134 t km^{-2} year^{-1}, the overall average turbidity of rivers is 360 g m^{-3}, the overall average rate of soil erosion is 0.09 mm year^{-1} and the total runoff is 17,564 milliard tons, including 3.7 milliard tons of chemical matter. The greatest amount of mechanical erosion occurs in Asia, the least in Europe. The overall average specific chemical erosion is 27 t km^{-2} year^{-1}, and the corresponding lowering of

Table 107. Total runoff data for mechanical erosion for each continent

Continents	Surface total area [mill. km^2]	Turbidity of of rivers [g m^{-3}]	Runoff [mil. tons]	Wash from surface [t km^{-2}]	Lowering of land surface [μ year^{-1}]	Ratio between mechanical and chemical runoff
Europe	9.67	163	725	75	50	1.8
Asia	44.89	649	9.361	208	139	7.0
Africa	29.81	291	2.152	72	48	2.4
Australia	7.96	421	345	43	29	4.5
North America	20.44	233	2.312	113	75	2.4
South America	17.98	208	2.669	148	99	2.2
World	130.75	360	17.564	134	90	4.0

the Earth's surface because of chemical erosion is 18 μ. The average specific gravity of soil was taken as 1.5 t m^{-3}.

These averages do not take account of the large movements of material that take place over relatively short distances, or the material that reaches the sea through other than the principal rivers. An example of the discrepancies that may arise because of this is given by the proportions of soil and earth transported in the Alps, where rates of removal of fluviates were found to be several times higher than those reported by Lopatin (1952) for the whole of Europe. Kukal (1964) gives the following rates of *denudation* and *degradation for the Alps:*

Agent of destruction	Weight of displaced material (10^6 t $year^{-1}$)
Water erosion	5,930
Landslides	274
Avalanches	45
Glaciers	27
Solifluction	0.75
Total	6,276.75

Thus in the Alps the annual removal of material amounts to more than 6 milliard tons, whereas all the rivers of Europe (according to Lopatin) remove only 0.725 milliard tons per year. Most mountain rivers deposit the larger part of the transported suspended matter in their river-beds, in alluvia, or in reservoirs, and only a small part of all transported material reaches the estuaries. What is more, a considerable proportion of the material that is locally loosened or displaced by erosion never reaches the watercourse. Most alpine rivers have a much higher specific flow of silt in the upper part of the catchment area than in the central and lower reaches. This is true not only of the Danube and other *Alpine rivers,* but also of such rivers as the *Mekong* (flowing through China, Laos, Thailand, Cambodia and Vietnam; specific silt runoff 1,200 t km^{-2}), the *Ganges* (flowing through India, Nepal and Bangladesh; specific silt runoff 1,040 t km^{-2}), the *Euphrates* and *Tigris* (flowing through Turkey, Syria, Iraq and Iran; specific silt runoff 690 and 1,000 t km^{-2}, respectively), and the *Irrawaddy* (flowing through Burma; specific silt runoff 850 t km^{-2}).

Much of the information that is given in the literature is out of date and inaccurate, and this unfortunately tends to foster a distorted picture of the intensity of erosion processes. For example, it is generally considered that the highest rate of fluvial erosion in a non-glacial river (1,144 m^3 km^{-2} $year^{-1}$) is to be found in the basin of the Himalayan *Kosi River,* although in fact the specific silt flow in the *Yellow River* is greater (1,850 t km^{-2} $year^{-1}$), and even exceeds 7,000 t km^{-2} $year^{-1}$ in some of the tributaries. In small torrents rates of removal may be still higher. The highest specific flow of silt ever recorded (approximately 30,000 t km^{-2} $year^{-1}$) is said to occur in the *Hidden River* in Alaska (Fig. 192), and it is probable that there are rivers in which the mean rate of removal of material is greater than that of the Hidden River.

As a general rule, however, rates of erosion in glacial rivers are about four times greater than those of non-glacial rivers, mechanical erosion being the predominant form of erosion in the latter. Rates of removal of material in some *glacial rivers* are given by Kukal (1964):

River basin	[t km^{-2} $year^{-1}$]
Heilstuge (Norway)	1,400
Blanc (French Alps)	1,600
Bosson (Chamonix)	1,800
Saskatchewan (Canada)	2,000
Auserfjätur (Norway)	2,200
Jokullsá (Finland)	2,000
L-Isortok (Greenland)	2,500
Hoffelsjökul (Iceland)	3,200
Muir (Alaska)	5,000

The figures given by some authors (e.g. Holeman 1968) for the flow of silt are low. Amounts of *annual deposits* quoted by Holeman are as follows: 0.32 milliard tons for Europe, 0.23 milliard tons for Australia, 0.54 milliard tons for Africa, 1.2 milliard tons for South America, and 1.96 milliard tons for North and Central America; this gives a total amount of silt carried away annually from these continents of only 4.25 milliard tons which is the same as other estimates for the USA alone.

More acceptable data are given by Fournier (1960). On the basis of a detailed theoretical analysis of *silt flow,* Fournier constructed lines on a map connecting points at which equal volumes of material are transported by rivers (Fig. 193). By means of planimetry he obtained values for the amounts of earth loosened and transported by the rivers. Values given by Fournier for mean *annual erosion losses* are as follows:

Europe	84 t km^{-2}
Australia	273 t km^{-2}
North and Central America	491 t km^{-2}
Asia	610 t km^{-2}
South America and the Antilles	701 t km^{-2}
Africa	715 t km^{-2}

Fig. 192. Deposits of the Muddy River near its confluence with the McKinley River in Alaska bear testimony to the immense amount of erosion going on in the river basin (from Shelton, "Geology Illustrated" 1966, San Francisco and London).

These figures, with the exception of the value for Europe, are several times higher than those quoted above, and they correspond more closely with rates of soil erosion observed on smaller territorial units. After conversion to units of volume (Fournier uses a factor of 1.4), the estimated *annual soil losses* and corresponding *depths of soil removed* on the various continents come to:

Europe	60 m^3 km^{-2}, i.e. 0.060 mm
Australia	195 m^3 km^{-2}, i.e. 0.195 mm
North and Central America	350 m^3 km^{-2}, i.e. 0.350 mm
Asia	435 m^3 km^{-2}, i.e. 0.435 mm
South America	500 m^3 km^{-2}, i.e. 0.500 mm
Africa	510 m^3 km^{-2}, i.e. 0.510 mm

Absolute values for erosion losses (again according to Fournier) can be derived for the various continents as follows:

	(thousand tons)
Europe (10,050,000 km^2)	844,200
Australia (7,626,000 km^2)	2,081,898
North and Central America (23,965,000 km^2)	11,766,815

South America and the Antilles (18,140,000 km^2)	11,599,600
Africa (29,800,000 km^2)	21,664,500
Asia (44,100,000 km^2)	26,930,000
Total	76,887,213

The *total annual soil loss* of 76,887,213,000 t corresponds to a mean specific silt flow of 571 t km^{-2} year^{-1} (407 m^3 km^{-2} year^{-1}). The mean depth of soil removal for all continents is 0.407 mm (Fournier 1960). Considering that these are world average figures for erosion, and that there are large areas of the world where erosion is minimal or where deposition is taking place, the figures are very high. Moreover, these considerations do not include areas affected by wind erosion, areas which include the very extensive arid regions of Africa, Asia and Australia.

The mountain regions and the semiarid and subtropical regions of the world make the largest contributions to the overall amount of water erosion. In all other regions the amount of erosion caused by surface water is small, but its consequences with respect to land utilization may be relatively greater. There are many parts of the world in which the rate of erosion is low, but the soils have been nevertheless considerably damaged by erosion. This is the case especially in regions with shallow soils or with soils or bedrock containing large amounts of soluble matter.

From this point of view, soil erosion phenomena in karst regions are particularly distinctly expressed. The most important are *carbonate karst formations on limestone* and *dolomitic rocks, gypsum karst formations on gypsum* and *anhydrite rocks,* and *salt karst formations on halites, silvinites,* and other rocks (Maksimovich 1955).

Limestone and *dolomitic karst formations* are the most resistant to chemical weathering, the soils tending to be shallow and to become dry; these areas are frequently rocky. If the soil is allowed to become exposed, small erosion losses lead to rapid degradation. Because the upper and lower layers are highly permeable, intrasoil erosion occurs and soil is washed into the karst hollows. The area of the Earth's land surface occupied by bare and buried carbonate rocks is estimated to be 40 million km^2.

Gypsum karst is characterized by hydrated sulphate formations and is more susceptible to chemical erosion. Layers of rocks forming gypsum karst are comparatively thin and the land is less barren than that of carbonate karst. The total land surface covered by gypsum formations is estimated to be 7 million km^2.

Salt karst consists of chloride formations, and corrosion on this type of karst is the fastest one; salt karst gives rise to a high degree of mineralization of both underground and surface waters. If salt layers are located near the surface, distinct surface and underground formations arise. Wasteland occurs both at the site of dissolution and at the site of crystallization of salt; in arid regions salt deserts

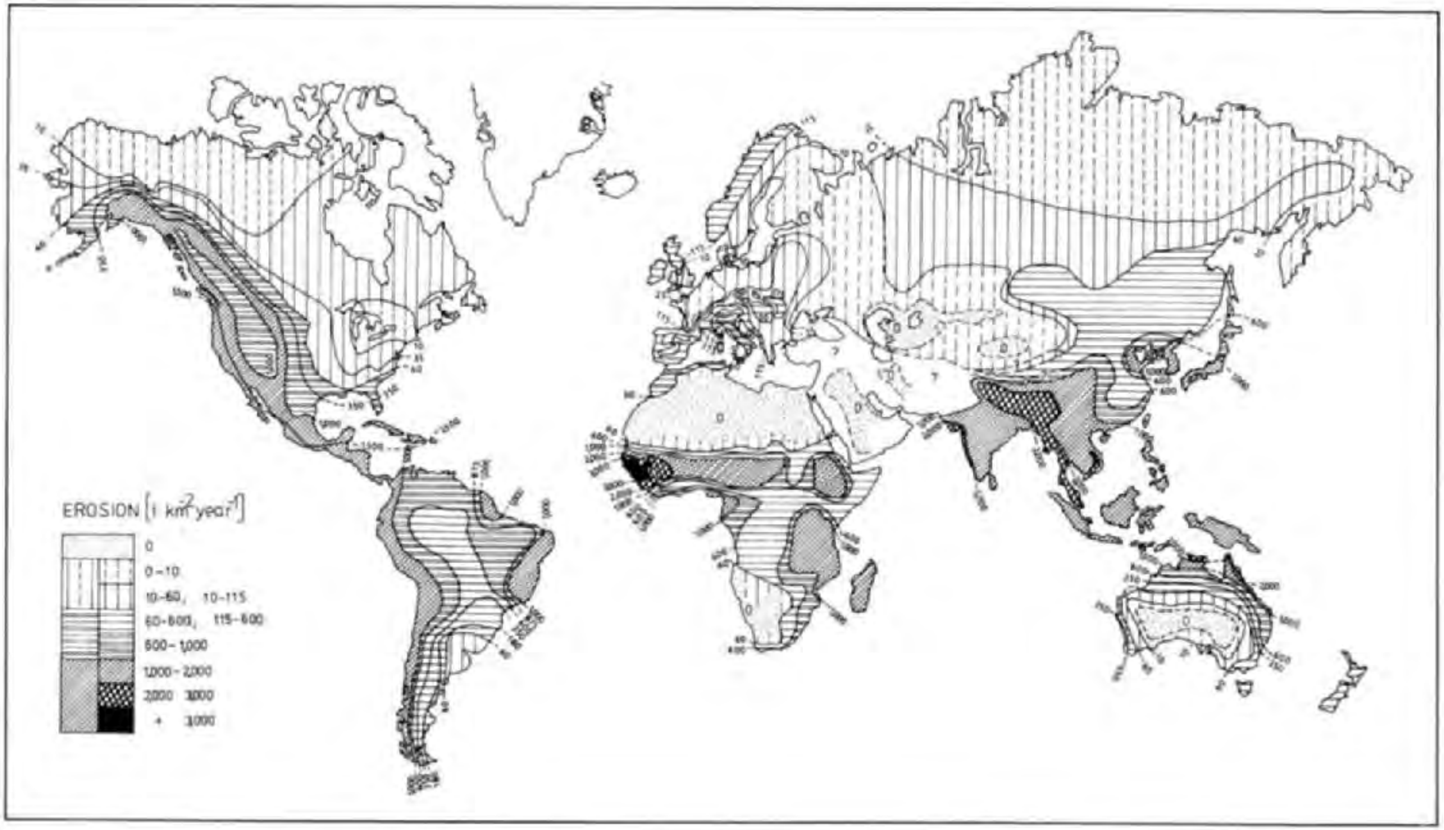


Fig. 193. General world map of the intensity of water erosion (Fournier 1960).

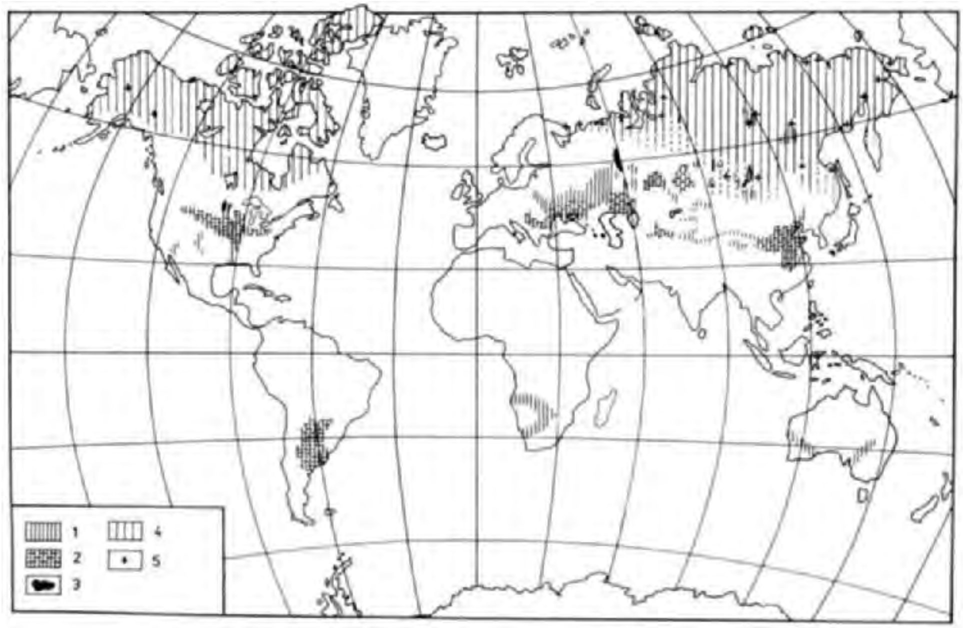


Fig. 194. General map of clastokarst and thermokarst. 1 – loess and loess sediments with potential karstification, 2 – clastokarst of loess, 3 – clastokarst of other rocks, 4 – soil with permafrost, 5 – thermokarst of permafrost soils.

develop. The total land surface occupied by bare and buried salt formations is estimated to be 4 million km^2.

Karst formations influence the processes of soil formation and soil erosion caused by surface and underground waters, and they have a considerable effect on chemical erosion in river basins.

In the colder regions of the globe, soil erosion depends on the thawing of surface ice or the upper layers of the permafrost. The proportion of frozen land on the Earth is estimated to be 20 to 25%, which includes about 20 million km^2 in the northern hemisphere. Warming of frozen areas by direct sunlight or an influx of warmer water causes depressions and other forms of *thermokarst.*

Distinct soil erosion phenomena are observed where *clastokarst* occurs, this being most widely distributed in loess and sediments. The total area of aleurolith karst (karst in loess and loess rocks) is estimated to be 30 million km^2. Karst phenomena can also be found on smaller areas comprising loams and other types of clastic rock, such as pyroclastics (tuffogenic deposits). The distribution of thermokarst and clastokarst phenomena is shown in Fig. 194; the map was prepared according to Maksimovich (1955).

5.7.2 Wind erosion

Wind erosion depends on wind force, the granular structure of the soil, the moisture content of the soil, and the density of the vegetation cover. Barren, sandy soils in arid regions are the most severely affected by wind action. The presence of solutions of salts, soda, gypstone and acid carbonate of calcium also has a strong bearing on the erosion process. The crystallization of salts on the soil surface produces a hardened layer which protects the soil against the wind, but has an adverse effect on plant growth. Therefore a combination of drought and salt results in wasteland, particularly in basins that do not drain into the sea.

In *humid* and *semihumid regions of moderate rainfall,* wind erosion occurs on exposed land (especially ploughed fields), in littoral zones where sandy shores border the sea or lake, and in the alluvia of the larger rivers. In wet regions wind erosion is contained both by frequent precipitation and by a high water table, and is thus restricted to periods of drought or to places of poor moisture retaining capacity. The main reason for accelerated erosion is the removal of vegetation and drainage of the land. Examples of areas in which wind erosion has been accelerated by the removal of vegetation are the Gascony region in the Landes, the littoral zones of the Baltic and parts of some central European countries.

Wind erosion is of much greater intensity, of course, in *semiarid* and *arid regions* where the upper layers of the weathering mantle dry more rapidly and the soil is less protected by vegetation. There is also much more intense precipitation erosion in these regions which separates and sorts the soil fractions depositing the finer particles in depressions from where they are blown away by the wind. The drier the climate, the greater is the prevalence of wind erosion. In some enclosed basins the wind may be the only destructive factor. The Libyan Desert is considered to be the driest of all deserts, and on the borders of areas surrounded by mountains enormous amounts of earth flow down the slopes during torrential rainstorms (bolsons, etc.); this material eventually being eroded by the wind.

Bare rocks are eroded more slowly by the wind as abrasion of the rock face proceeds together with removal of the loosened material. Genuine deserts develop in extremely arid regions where the rainfall is less than 200 mm year^{-1}.

Wind erosion also occurs in areas covered by volcanic ash (e.g. Iceland), and in periglacial regions (periglacial deserts).

With respect to the surface characteristics of those desert soils that are most affected by wind erosion, *rock* and *stone deserts, gravel deserts, loam* and *clay deserts,* and *salt deserts* may be distinguished.

Rock deserts are widely distributed and are referred to in the literature by the term *hamada,* the name given to this type of land in northern Africa. *Gravel deserts* are covered by smaller stones which frequently form a *desert pavement;* such deserts are called *serir* in northern Africa. The typical sand desert forming a sea of sand is called *erg* or *areg* in Africa, and *kum* in central Asia. Deserts with *dunes* and

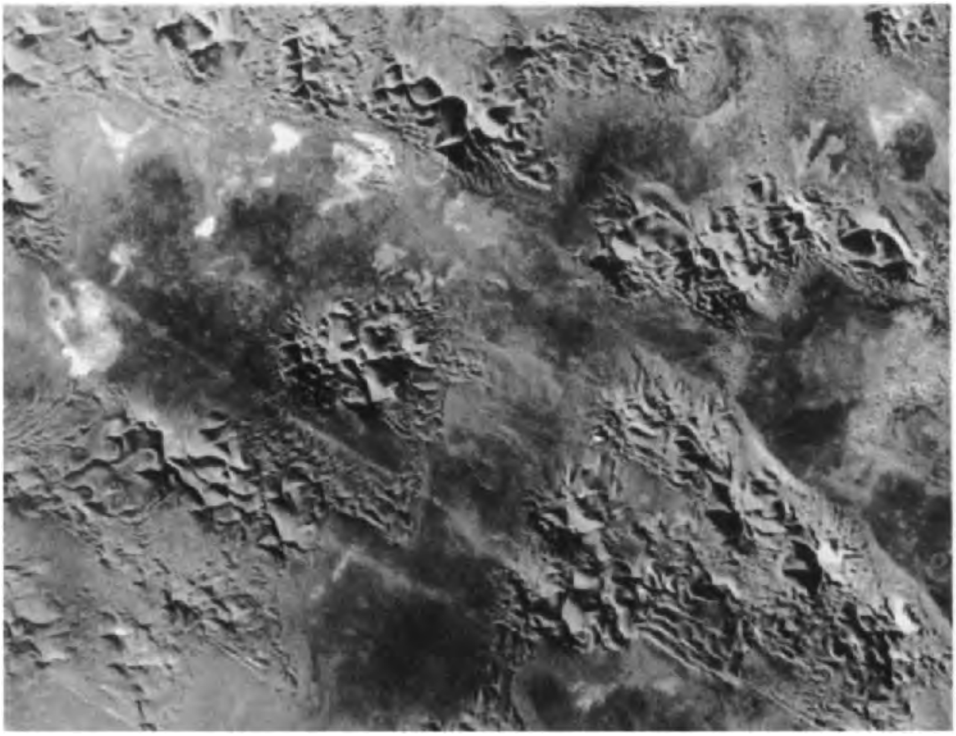


Fig. 195. Aerial view of the Sahara Desert from about 15 km altitude. (By courtesy of Institute Géographique National, Paris.)

barkhans are called ***nebka***. Loam and clay deserts which form the bottoms of temporary salt lakes are called ***shott*** in Tunisia and Algeria; desert saline soil in Turkestan are called ***takyr;*** salt deserts in Iran are called ***kavir***, etc.

The most extensive deserts occur in northern Africa and southwestern and central Asia. A wide belt of deserts spreads through both continents beginning on the Atlantic coast and ending on the eastern borders of central Asia.

The largest desert on Earth is *Sahara* (Fig. 195), which with its adjoining regions, at present extends over an area of more than 9 million km^2, which includes 6 million km^2 of genuine desert (Kettner 1955). About one ninth of the area of the Sahara consists of sandy desert *(erg, areg)* extending mostly over Algeria. The remainder is made up of ***hamada*** (occurring in almost every part of northern Africa), *serir* (mainly in the Libyan Desert) and *shott*, which are found in depressions in the sandy desert regions.

In southwestern Asia the desert belt starts in Sinai, Syria (the *Syrian Desert* is shown in Fig. 196), and Arabia *(Nefud* and *Dehma deserts)*. *Arabian deserts* have extensive areas of dunes which cover about one third of the total desert area. In Turkmenistan there is the *Kara Kum Desert,* further eastward the *Kizil Kum Desert,* and in the Tamir basin lies the *Takla Makan Desert.* Mongolia boasts the largest deserts *Gobi* or *Shamo (sea of sand),* which has an area of about 1.2 million km^2; it is composed mostly of sand and loam desert, takyr, and part gravel desert. In Iran there is the *great salt desert,* called the *Kavir Desert.* To the east of the Indus is the *Thar Desert.*

In southern Africa lies the *Namib Desert ("diamond desert")* and to the east of it in the inland of Botswana there is the *Kalahari Desert.* Adjoining these deserts are semiarid, wind-threatened regions.

The most extensive desert regions of the tropic of Capricorn are the *Great Salt Desert* (also called the *Great Sandy Desert*) and to the south of it the *Great Victoria Desert* both of which are in Australia. The adjoining semideserts and semiarid, wind-threatened regions are also fairly extensive.

In North America genuine desert areas occupy relatively small areas, and consist mostly of stone, clay and salt deserts. The largest desert in North America is the *Mohave Desert* of California which has an area of about 125,000 km^2. In southern Arizona there is the *Gila Desert* and in northern Arizona, the *Painted Desert.* To these can be added the undrained basin of northern Utah, the *Great Salt Lake Desert.*

In South America the *Atacama* salt desert of northern Chile and the sand deserts of the *Matto Grosso* Province of Brazil are well-known.

In colder regions of the globe *periglacial deserts* occur, especially where the vegetation cover has been naturally or artificially disturbed, or where wind-eroded material has accumulated. Areas which are naturally poor in vegetation are expanses of sea sand or river sand, and areas covered by volcanic ash. Periglacial deserts occur mainly in the tundras of North America and Eurasia.

The largest single area (about 12.5 million km^2) of deserts, semideserts and adjoining wind-threatened regions is that of Asia, followed in sequence by Africa (about 12.0 million km^2), Australia, North America, and South America. In Europe there are extensive areas of active littoral sand dunes in France (about 900 km^2), Spain, Italy and the Baltic and Scandinavian countries. The total area of deserts, semideserts and other wind-threatened regions in the world is estimated to be 30 million km^2, i.e. approximately 20% of the area of the continents.

It is in the desert regions of the world that the most intense erosion of all takes place, the intensity of wind erosion being potentially as high as that of precipitation erosion. As indicated in previous chapters rates of erosion during dust storms may reach an intensity of more than 1,000 t ha^{-1}, the material being transported over shorter or longer distances. It is known that a sand storm in the Sahara continuing from 9th to 12th March 1901 caused 1,960,420 tons of dust to be carried to Europe, which if covered evenly would have received a layer of average thickness 0.25 mm; the dust was transported over a distance of 3,000 to 4,000 km. Similar

Fig. 196. Aerial view of a part of the Syrian Desert on a cretaceous and limestone plateau. (By courtesy of Farey Air Surveys.)

phenomena have also been observed in New Zealand where dust arrives from Australia, and in Japan where dust originating from central China is deposited.

Since the quantities of dust in the already aeolized soils of the Sahara are now small, and since only a small proportion of it is usually carried to Europe during sand storms, the material transported in 1901 must have been of the order of milliards of cubic metres. A large part of the eroded material is carried from deserts to the sea, finds its way into depressions and watercourses, or accumulates in the lee slopes of mountain ranges. According to Richtenhofen, the deposition of aeolian dust explains the origin of loess which forms large deposits in China and in the sheltered areas of the central Asian deserts, especially the Gobi Desert. Dust deposits are often eroded by water, but in arid and semiarid regions they are subjected to further wind erosion.

The total land surface affected by both wind and water erosion is estimated to be 20 million km^2 which accounts for 14% of the total dry land area. The affected area includes land used for a variety of different purposes, and therefore the intensity of wind erosion varies greatly according to land type, ploughed and extensively grazed land being the most severely affected. The intensity of wind erosion and the area of land affected are both increasing at present, especially in the semiarid regions of the world.

In general, the world distribution of wind erosion coincides with the distribution of wasteland which is described in the following section. On the map of desertification (Fig. 197), it can be seen that wind erosion is mostly confined to desert and semidesert regions (marked on the map by hatched areas 1 and 2). Hatched areas 3 and 4 denote soil desertification caused by both wind and water erosion.

5.7.3 Devastation of the soil

Water erosion and wind erosion have a negative effect on the soil, on the ecology of the soil, on plant life and the soil-protecting function of plant life. In general, erosion causes the deterioration of the soil and the whole of the landscape. Man may accelerate or slow down this process. The more extreme the prevailing natural conditions, the stronger is the deleterious effect of human interference. Under extreme conditions *desertification* of the soil occurs; this is understood to mean impoverishment of terrestrial ecosystems by the activities of man in association with *water erosion, wind erosion, salination,* and the *deterioration* of the *micro-, meso-* and *macro-climatic conditions* over large areas.

This definition of desertification corresponds with that given by Dregne in his article Desertification: Man's abuse of the land (1978) in which he reports on the conference on desertification held in Kenya in 1977. Criteria and data on desertification are quoted in Table 108, and the map is shown in Fig. 197.

According to Dregne about 82% of the world's arid regions are in the second and third stages of devastation. In these regions about 680 million people suffer directly or indirectly from the effects of erosion. Annual wheat yield losses caused by water and wind erosion are estimated to be 35.24 million m^3 $year^{-1}$. On 13 farms in southwestern Kansas covering altogether 1.2 million acres (485,625 ha), the annual loss in yield has been estimated to be 339,000 bushels (11,946 m^3) of wheat and 543,000 bushels (19,135 m^3) of grain sorghum. Economic losses caused by desertification in arid regions are estimated by Dregne to be 15.6 milliard dollars a year. On the other hand, the running costs of erosion control schemes are relatively low and in most countries represent only a small part of the capital investment necessary for setting up these schemes. Croplands in particular are usually inadequately protected.

In the previous chapters we have seen how the destruction of the soil and the desertification of the landscape occur as a consequence of erosion accelerated by man's interference with the balance of nature. These processes also occur in

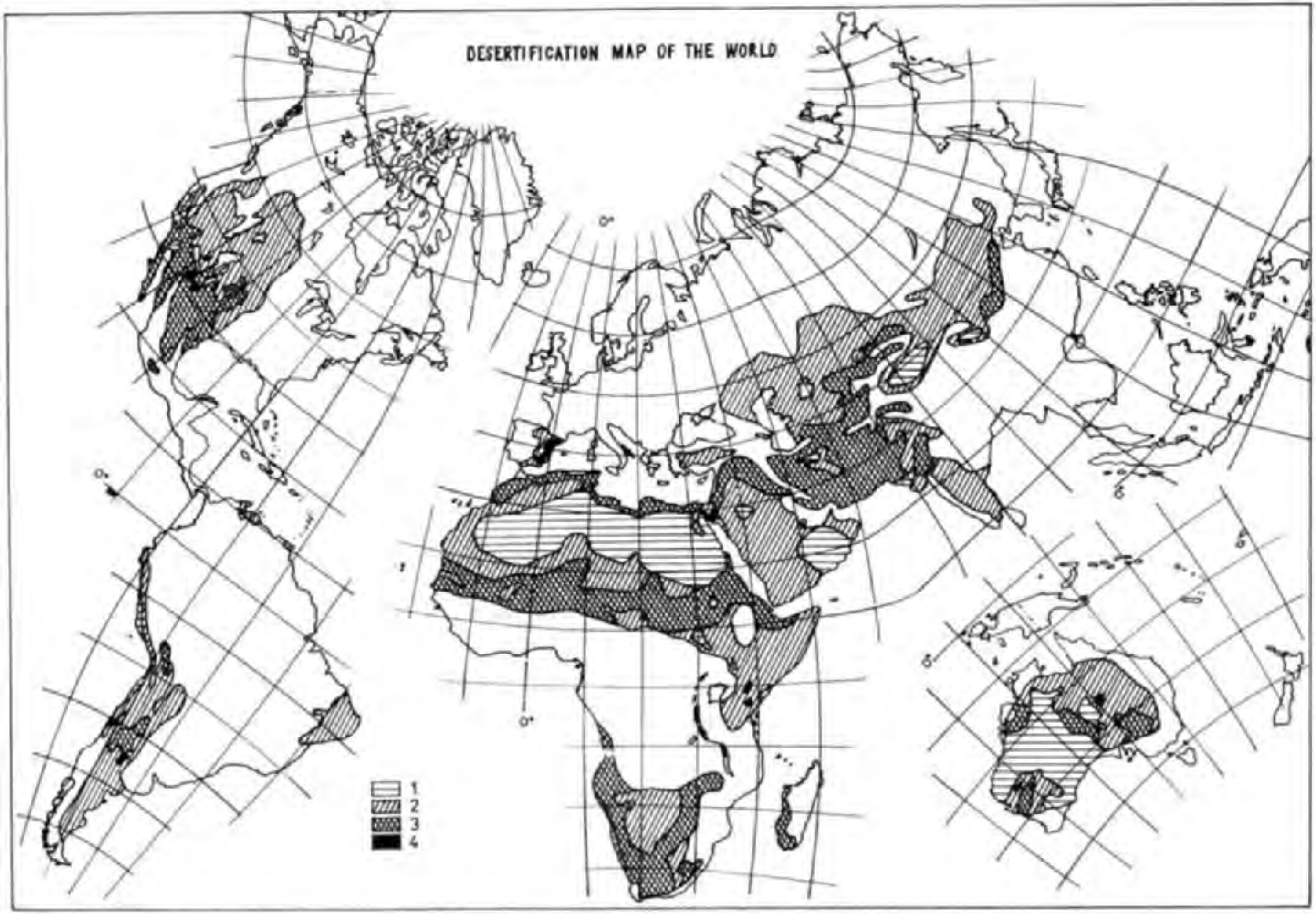


Fig. 197. Desertification map of the world (Dregne 1978).

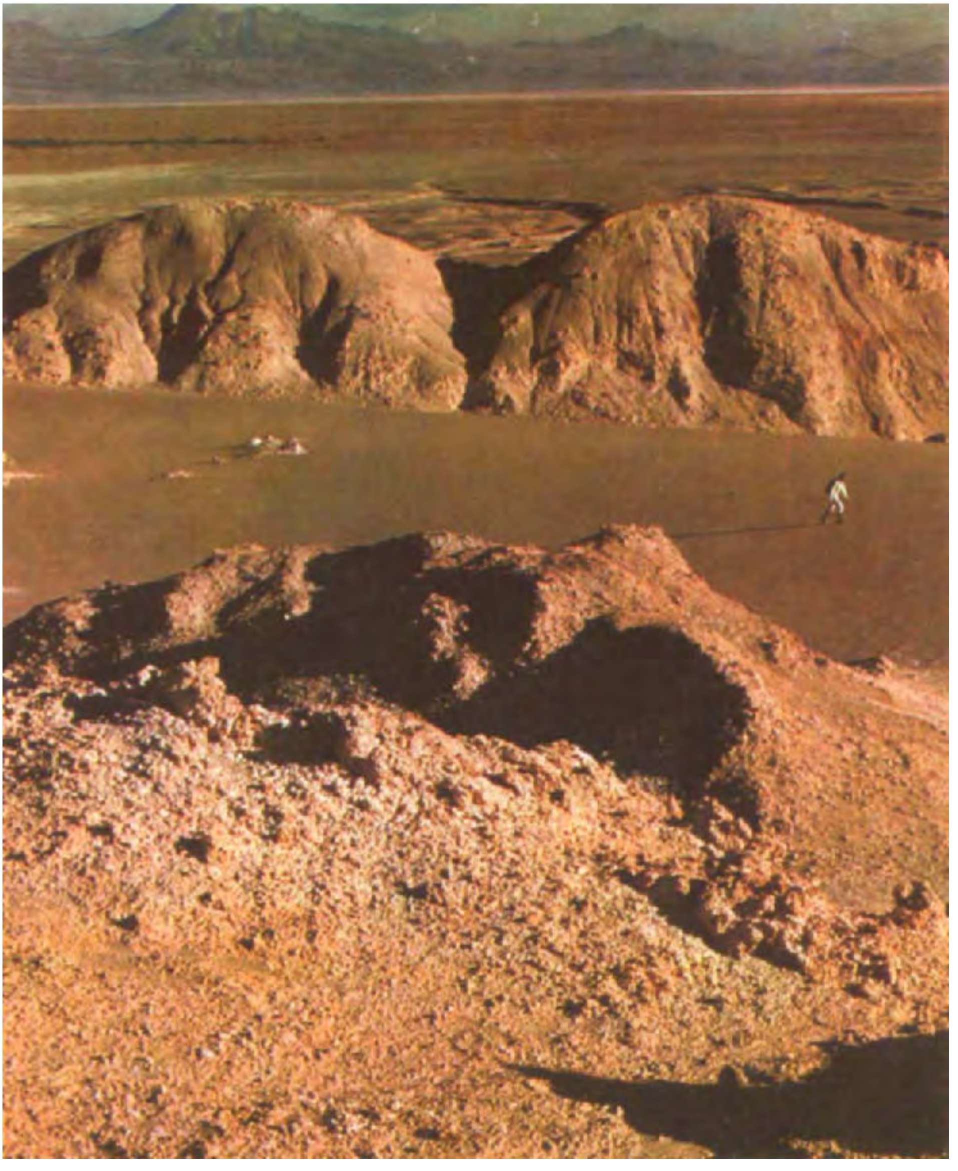


Fig. 198. The alpine salt desert Atacama in Chile. (Photo W. Bonatti.)

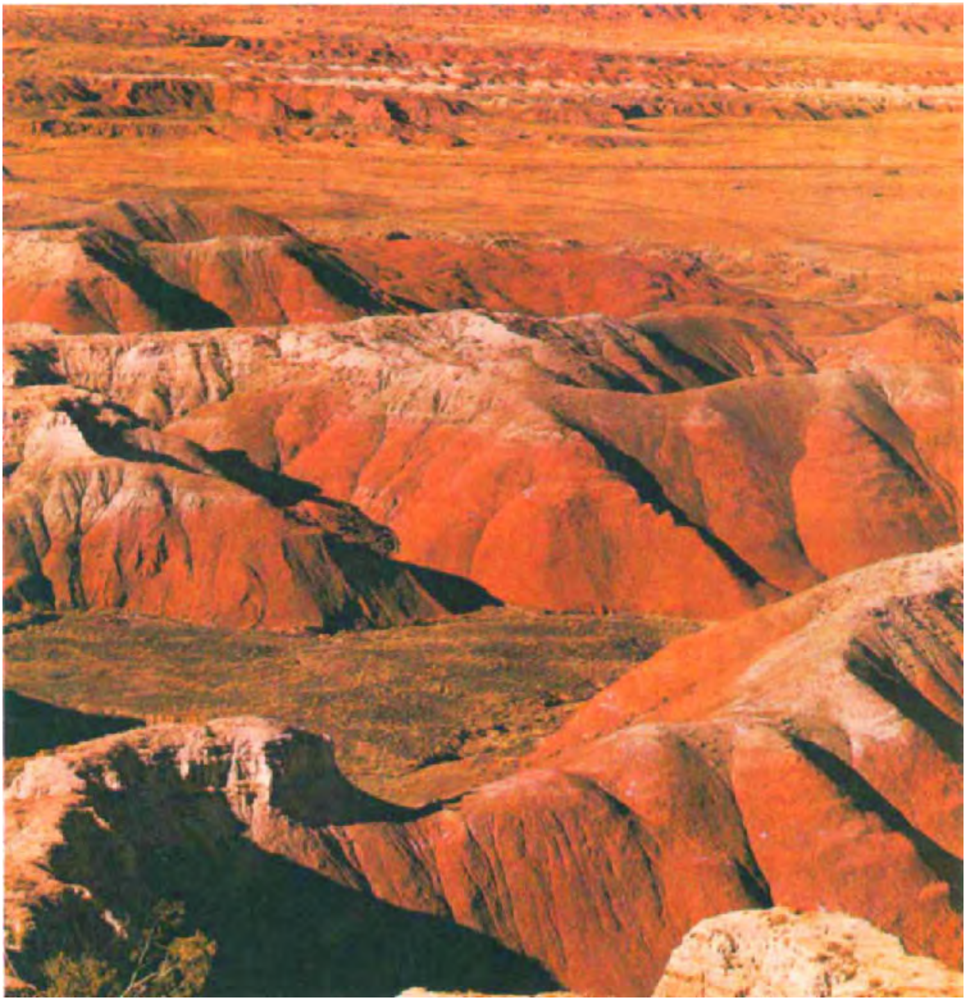


Fig. 199. Painted Desert in northern Arizona. (Photo J. Muench.)

Fig. 200. Bedrock denuded by accelerated erosion and barren land in the alpine regions of Chile. (Photo J. Čech.)

Table 108. Criteria for establishing the degree of desertification and percentage of world's arid lands affected by desertification

Degree of desertification	Plant cover*	Erosion	Salination of waterlogging (irrigated land)	Percentage of arid land
Slight	Excellent to good*	Nil to slight	Crop yields reduced by less than 10%	18.0
Moderate	Fair*	Moderate sheet erosion, shallow gullies, few hummocks	Crop yields reduced by 10—50 %	53.6
Severe	Poor*	Severe sheet erosion, gullies common, some areas affected by blow-off	Crop yields reduced by more than 50 %	28.3
Very severe	Land essentially denuded of vegetation	Severely gullied, numerous areas affected by blow-off	Heavy salt crust on near-impermeable soils	0.1

* Remarks refer to range of species and abundance of vegetation.

Fig. 201. Land in the foothills of the High Atlas covered in the past by vegetation (Morocco). (Photo D. Zachar.)

regions where the climate is harsh, and may attain catastrophic proportions. This is particularly true of mountain regions, and also of regions with a highly dissected relief and soil of low resistance to erosion (Figs. 198–201).

It may be expected that with increasing demands for food, the human population will increase its pressure on the land, leading to further expansion in the amount of cultivated land, ever greater numbers of domestic animals, more widespread irrigation of the land, and an increased input of chemical additives of all kinds into the soil.

According to the FAO (Production Yearbook, Vol. 27, 1973, Rome), out of a total of 13.4 milliard ha, arable land covers 1.47 milliard ha (11.01%), permanent meadow and pasture cover 3.01 milliard ha (22.43 %), forest and woodlands cover 3.99 milliard ha (29.78%), and other types of land cover 4.93 milliard ha (36.78%). The latter category includes so far unutilized land, built-up areas, and land destroyed by industrial activities; the requirement for land designed for further urban and industrial development will increase, unfortunately at the expense of arable land in particular which will have to expand in other directions. The consequences of any expansion in the overall area of arable land in the USA and other countries are very serious. In 1973–1974 average soil losses caused by erosion on newly ploughed soil in the USA were 37 t ha^{-1} $year^{-1}$, and there were greater losses in other countries, too.

The problems of soil erosion and soil protection are clearly going to become increasingly urgent in the future, and food production for a growing world population may only be safeguarded if this problem is successfully overcome.

5.8 Conclusion

This work gives an account of the most important aspects of soil erosion, including systems of classification, methods of research and survey, and an appraisal of the various erosion factors and their distribution. The author has endeavoured to include information on the various natural and economic factors that interact with the erosion process, and to present a relatively comprehensive view of these processes. Nevertheless, many questions remain unanswered and in many cases it has been necessary to restrict the discourse to approximate estimates and general considerations.

In spite of this the author hopes that this work will play a useful role in the further development of both the discipline of soil erosion and the techniques of soil conservation. Soil erosion is now being studied from different angles by many specialists in different fields; a large proportion of new publications deals mainly with erosion and erosion control on arable land, other publications are devoted to a particular type or form of erosion which is of special regional significance, but relatively few works take a comprehensive view of erosion.

Much work still needs to be done to clarify the ecological consequences of erosion on the soil, the water environment, the atmosphere, and the whole living environment. The reduced productivity of ecosystems and the damage caused by erosion need to be better understood, so that methods of soil protection can be improved. Protection of the land from the deposition of erosion products and from harmful agents which are released into the environment during the erosion process bring their own special problems.

It may be added in concluding that soil erosion phenomena are as complicated as the natural conditions under which they occur, as well as the different types of land involved. Therefore no theoretical work can provide practical solutions to problems of soil erosion under specific conditions, but such a work may help to throw light on the basic features of the phenomenon and indicate the general direction of a practical solution. It is to be hoped that the intensive research on erosion, which has been carried out during the last few decades in different parts of the world will provide sufficient information to make possible effective and comprehensive protection of the soil, which is one of man's most important resources. The rate of progress is determined by the application of erosion control measures in practice, and the success of these measures is the most important criterion for judging the value of current theories and planning further development.

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AUTHOR INDEX

SUBJECT INDEX

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