1768

The New Encyclopædia Britannica

Volume 20

MACROPÆDIA

Knowledge in Depth

FOUNDED 1768
15 TH EDITION

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First Edition	1768–1771
Second Edition	1777–1784
Third Edition	1788–1797
Supplement	1801
Fourth Edition	1801–1809
Fifth Edition	1815
Sixth Edition	1820–1823
Supplement	1815–1824
Seventh Edition	1830–1842
Eighth Edition	1852–1860
Ninth Edition	1875–1889
Tenth Edition	1902–1903

Printed in U.S.A.

Library of Congress Catalog Card Number: 91-75907
International Standard Book Number: 0-85229-571-5

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Geomorphic Processes

Geomorphic processes are all those physical and chemical processes that affect the surface features of the Earth. They include tectonic activity and surficial earth movements such as rockfalls and landslides. Geomorphic processes also involve weathering and the erosion and deposition of the resultant rock debris by streams, glaciers, and wind.

This article treats the mechanics and dynamics of these various natural geomorphic agents and processes. In addition, it discusses the ways in which humans promote their performance in altering landforms and near-surface features. The direct effects of human action on the physical environment are covered as well. For further information about the origins and evolution of landforms, see the articles CONTINENTAL LANDFORMS; PLATE TECTONICS; EARTHQUAKES; RIVERS; OCEANS; and ICE AND ICE FORMATIONS: *Glaciers*. (Ed.)

This article is divided into the following sections:

Physiographic effects of tectonism

It has long been known that tectonism has profoundly influenced the physiography of the Earth. The Russian scientist M.V. Lomonosov stated in 1763 that the Earth's relief is formed by shaking of the Earth (earthquakes) or long-term sinking and raising of its surface. The German geographer Alexander von Humboldt suggested that mountain ridges were formed by upwelling of molten lava (volcanism) from the Earth's interior. This idea was derived from observation of volcanic activity in Central America. The French scientist Élie de Beaumont, however, thought that the Earth's relief features were principally attributable to gradual compression or shrinking of the Earth's crust because of slow cooling since its formation. In this view, tangential pressures in the crust caused folding and general tectonic movements.

The Austrian geologist Eduard Suess noted that rock deformation was particularly intensive in mountainous areas, and he came to the conclusion that a close link existed between mountain ranges and crustal uplift and between oceanic and other basins and tectonic sinking or depression. In general, scientists today discriminate between tectonic movements that are geographically restricted and involved in mountain building (orogenic movements) and those movements that are broader, less localized, and associated with the uplift or depression of plateaus and basins (epeirogenic movements). Both types of tectonism are responsible for the Earth's present relief features or physiography. Modern investigations have led to the theory that all tectonism is produced by the movement of great blocks of the Earth's crust relative to each other (see PLATE TECTONICS).

RECENT TECTONIC MOVEMENTS

Seismic movements. Direct observations and theoretical generalizations suggest a close connection between earthquakes, which periodically occur in different parts of the globe, and recent tectonic movements. Such movements are considered to be mechanical dislocations or faults within the crust. When faulting occurs, earthquakes are generated. The detection, recording, and analysis of the resulting seismic waves provides information on the intensity of an earthquake and its location—both where it occurred on the surface of the Earth and at what depth.

Tectonic movements of a seismic character are often called rapid because the physical consequences occur swiftly. The reasons for these tectonic movements cannot be directly observed because they occur at depths of hundreds of kilometres, but strong earthquakes generally manifest themselves in repeated disturbances of the Earth's crust through time and thus produce discernible physiographic effects at the surface.

Rapid tectonic movements frequently occur in the Soviet Union, where intensive earthquakes of destructive character take place several times every year; less intensive earthquakes are even more common. The earthquake epicentres are geographically localized and this provides the opportunity to carry out regional seismic investigations and to forecast the locations and the intensity of possible earthquakes. The greatest frequency of intensive, rapid tectonic movements has been established as characteristic of mountain regions—from North Africa, the Alps, Balkans, and Caucasus to the Tien Shan, Pamirs, and the Himalayas and around the margins of the Pacific Ocean (Alaska, the Aleutians, Coast Ranges, Andes, mountains of New Zealand and Indonesia, and the Japanese islands,

Mountain regions of intensive seismic activity the Kurils, and Kamchatka). There are reasons to suppose that the main sources of rapid tectonic movements are the zones of deep faults that extend from the crust into the upper mantle (zone beneath the crust) of the Earth.

Slow tectonic movements. Slow tectonic movements of the Earth's surface were detected first in Scandinavia (the Baltic or Fennoscandian Shield, an area of ancient crystalline rocks) and subsequently in the Canadian Shield. In both regions dome-shaped, elevated, ancient shorelines and deposits of former lakes and seas were discovered. These phenomena have been explained in terms of a rather broad upward arching during the last 10,000 to 12,000 years because of constant melting of thick glacial ice cover and consequent rising caused by lightening of weight. The maximum uplift of Fennoscandia for this period was about 250–275 metres (820–900 feet) in the northern part of the Gulf of Bothnia, according to some authorities; uplift in the Hudson Bay region of Canada has attained nearly 300 metres (1,000 feet). Simple calculation shows that average rate of such tectonic movements reaches two to three centimetres (0.8 to 1.2 inches) per year in the tops of arches, gradually decreasing toward their peripheral regions.

Slow tectonic uplift and subsidence with such characteristics are common almost everywhere. These slow tectonic movements were detected by precise geodetic measurements, and their average velocities ranged from tenths of millimetres to several millimetres per year for plains areas and 10 millimetres (0.4 inch) or more per year in mountain regions. Unfortunately, data are available for no more than 100 years. For this reason it is not yet possible to define the general trends (or periodicity) of such movements, as is possible in regions of postglacial isostatic uplifts, where there are records or available data. Nevertheless, the main conclusion about recent tectonic mobility, supported by seismic phenomena, may at present be considered proved by numerous geodetic, oceanographic, and geomorphological investigations.

Recent horizontal movements of separate parts of the Earth's surface also have been detected with the help of precise geodetic measurements and geomorphic observations. The features of horizontal displacement along the Talass-Fergana Fault in Soviet Central Asia or more complex movements of separate points along the San Andreas Fault in California are examples. In both regions single horizontal displacements of several metres have been observed immediately after strong earthquakes.

Recent tectonic mobility must be considered proved, although not all aspects are yet understood. The evidence consists of orogenic zones that coincide with seismic phenomena, shield areas in which isostatic movements occur, and regional lines of horizontal dislocations. It is not yet possible, however, to make an evaluation of tectonic mobility through time and, in particular, to define the periodicity, if any, that is manifested. Further measurements and observations are needed for a full understanding of these phenomena.

EFFECTS ON THE EARTH'S SURFACE

Uplift and denudation. Modern concepts of the origins of the Earth's surface features are based upon acceptance of the existence of a dynamic equilibrium between vertical tectonic movements (uplift and subsidence) and denudational processes, which encompass the erosion of rocks and accumulation of the resulting sedimentary debris.

Rates of continental denudation

The German geographer Albrecht Penck considered data on the volume of transported river sediments and concluded in 1894 that a layer 0.8 millimetre (0.03 inch) thick is annually transported to the sea by rivers. More recent calculations have shown that the solid sediment transported by rivers is alone sufficient to account for reduction of the Earth's surface by 0.09 millimetre (0.004 inch) per year. Other values, given by various scientists, range from 0.01 to 0.02 millimetre (0.0004 to 0.0008 inch) in lowlands to 0.6 to 0.8 millimetre (0.02 to 0.03 inch) per year in mountains. Values as great as 0.16 millimetre per year have been given for the United States.

It should be noted that the sediments transported by rivers make up only part of the general volume of all sediments that are subject to continental denudation. Some sediment is not carried to the sea at all but is retained within the area being eroded—on terraces, hillslopes, in local closed depressions, and elsewhere (see below). This material, which is not transported to the sea, is not included in the calculations on which the foregoing data are based. Hence, values of continental denudation are greater than stated.

The general correlation between recent tectonic movements and continental denudation may be considered an indication of dynamic stability of the Earth's surface. The stability is characterized by the following process: slow tectonic uplift of certain areas stimulates erosion and denudation, whereas other parts of the Earth's surface undergo tectonic subsidence and the eroded material tends to accumulate in place, although part of it is transported to the sea in one form or another.

Data on the rate of development of a normal soil profile (sequence of vertical layers) under a natural vegetation cover are instructive in this regard. The use of radiometric measurements for defining the absolute age of humus in the recent soils has shown that a time period ranging from several hundred to 1,000 to 3,000 years (depending on the genetic type of a soil) is required for the formation of a normally developed soil. This means that the natural "growth" of soil thickness ranges from one to 10 millimetres per year. It shows that under the conditions of natural denudation, a well-regulated dynamic system exists. The three basic processes of the system—tectonic movements, soil formation, and continental denudation—tend essentially toward a state of balance. The middle component of the system, the soil-vegetation cover, serves as its main regulator.

Equilibrium is lost when the natural soil-vegetation cover is disturbed, whether by human activities or natural causes. Rather widespread phenomena of natural denudation are evident in mudflows, which form in mountains during rainstorms when there is an exceedingly high surface runoff. They are characteristic of many high mountain zones, where thick accumulations of loose sediments are present, and in arid regions, where the vegetation cover is sparse. Catastrophic displacements of ground masses on slopes (avalanches, landslides, and mudflows) are quite common in these areas. They are usually caused by earthquakes; and many interruptions of the soil-vegetation cover on mountain slopes are characteristic elements of the natural landscapes and must be considered a reflection of disturbed equilibrium.

Mudflows as an example of natural denudation

Relief features. The state of equilibrium between tectonic movements and the processes of denudation should provide for a stable, level character of the Earth's surface; it is clear, however, that such is not the case. Aside from extraordinary disturbance of the equilibrium by catastrophic movements of sediment, more gradual tectonic movements and denudation are not balanced in the strict sense and may be progressively altered and directed over time. Consideration of this balance—or lack of balance—has led to the distinction among three main categories of elements in the structure of the Earth's surface. The largest category consists of first-order relief features, which are elements of the so-called geotecture. They include crystalline massifs, continental platforms, oceanic hollows, and large mountain systems.

Positive and negative elements of lesser magnitude make up the second-order relief features that complicate the surface of continents and ocean floors; these elements are called the morphostructure. This group includes plateaus, uplands, lowlands, ridges, and similar features. Morphostructures may be described chiefly as large forms of relief that emerge as a result of the interaction of tectonic and denudational forces. Tectonic movements play the leading role in this interaction.

Small elements of relief of the third order include such features as river valleys, lake basins, and dune or karst (cavern) forms. They are sometimes called elements of morphosculpture. The origin of these morphosculptural elements is largely dependent upon denudational processes.

It has been suggested that all tectonic movements that are expressed in present-day relief should be called neotectonics. In papers by Soviet scientists this connotation

Neotectonics as the basis of present-day relief

emerges from considerations of the general amplitude and character of the most recent tectonic movements. A review map of these movements in the Soviet Union served as a model for international compilations of this type. The main basis for the work was study of the thickness of Pliocene–Quaternary deposits (younger than 7,000,000 years) that accumulated in tectonic subsidence areas (depressions and basins) and of the amplitude of uplift of ancient surfaces in mountain ridges. In these instances, the obtained values of uplift and deposition amounted to hundreds of metres of relief over distances of several kilometres. The role of neotectonics in forming the present relief features is acknowledged by all authorities. (I.P.G.)

Weathering

In general terms, weathering may be defined as the disintegration or alteration *in situ* of rocks at and near the Earth's surface, and within the range of temperatures that occur there. The distinction drawn between disintegration and alteration highlights the difference between physical and chemical processes. Disintegration involves a breakdown of the rock into its constituent minerals or particles, with no decay of any of the rock-forming minerals. Chemical weathering, on the other hand, implies the alteration of one or more of these minerals. The phrase *in situ* is not meant to suggest that there is no translocation of material within the rock being weathered. Redistribution and reorganization of various minerals are clearly at work in many vertical soil profiles, and these processes lead to the development of distinct horizons or zones and the lateral movement of minerals. The rock mass as a whole remains in place, however. Finally, the location of weathering processes at or near the ground surface, plus the range of temperatures indicated, differentiates these processes from metamorphism, which takes place either deep in the crust or at higher temperatures than are characteristic of weathering reactions.

eathering ıd ndforms

Weathering is an essential precursor to erosion and transportation. Weathering reduces rocks to particles and to a condition suitable for transportation. Hence, it is a key phase in the eventual production of new strata, as well as in the formation of the alluvial lowlands. Weathering also plays an important part in shaping scenery. It is not too much to say that weathering, by exploiting various weaknesses in the Earth's crust and thus preparing the way for erosional agencies such as rivers, glaciers, and wind-driven waves, plays a major role in determining the form of the land surface over much of the continental area of the globe. Upland and plain, ridge and valley, hill and lowland, are in large part a reflection of structural contrasts brought about by weathering. In addition, pedogenic accumulations of minerals can reduce infiltration of water, cause heavy and rapid runoff, and thus induce floods. Iron pans in the soils of Exmoor, in southwestern England, were an important factor in the flooding that partially destroyed Lynmouth in 1952. On a large scale, silcrete and laterite, for instance, form the caprock in many plateaus and mesas in arid Australia. Even travertine deposited in river beds may protect the landscape to such an extent that the river bed becomes more resistant than the surrounding areas and eventually comes to stand above the rest of the land surface. An example of such inversion of relief, with a winding, travertine-capped old river course now forming a distinct ridge, has been described from Arabia. Finally, numerous minor forms, including elegant flares, caverns, pits and etchings, are a result of weathering.

PROCESSES INVOLVED IN WEATHERING

Physical processes. Weathering processes are complex and several processes generally act together to achieve rock disintegration and decay. The processes may for convenience be labelled physical (or mechanical), chemical, or biological, but such treatment tends to disguise the essential complexity and interconnection of weathering activities. Very few weathering processes can be observed in action; scientists must be content to see the results and try to infer from them what has taken place. Many variables affect weathering reactions, and laboratory experiments fail to reproduce either the complexity or the immense duration of geological time, which together provide the only correct context in which to view these processes. As a result, accounts of weathering have been dominated by speculation and theorization, much of which, though it seems logical enough, fails to accord with the evidence of nature.

Thermal expansion and contraction

Thermal expansion and contraction, for example, has long been cited as a cause of rock disintegration, particularly in the tropical deserts where great extremes of temperature are experienced. Man-made fires were used as an aid in quarrying in ancient Egypt and in India; bush or forest fires certainly cause superficial rock flaking; and some stones subjected to the intense, though local, heat of camp fires are rapidly split. Reports of loud cracking noises in tropical regions have been attributed to the expansion or contraction of rocks on heating or cooling. Thus many types of weathering, including granular disintegration, spheroidal weathering, onion (or onionskin) weathering, and sheet structure, have been explained in terms of thermal expansion and cooling as described below.

Two related processes were involved. First, it was argued that because many rocks consist of minerals that expand by differing amounts when subjected to the same temperature change, they would, on heating, expand at varied rates. As a result, daily heating and cooling would eventually loosen the cohesion between the rock-forming minerals and would thus cause the breakdown of the rock into small particles. The process described is termed granular disintegration. Second, because rocks are poor conductors of heat, those parts of a rock mass exposed at the surface would expand and contract as they are heated and cooled, while those parts below the surface should remain at a constant temperature and volume. It was supposed that stresses would be set up between the outer and inner zones, the former eventually becoming separated due to the development of fractures, and repetition of the process would cause the development of concentric layers or sheets of rock. This process resulted in spheroidal, onion weathering, or sheet structure, depending on the thickness of the detached rock masses.

It now appears, however, that most field evidence points to contrary conclusions. Granular disintegration, spheroidal and onionskin weathering, and sheet structure have all been found deep beneath the Earth's surface, well beyond the range of the Sun's heat; and even in the tropical deserts, which should be most suitable for solar-heating effects, the evidence is contrary to that required by the insolation hypothesis. Observations in the Egyptian deserts indicate that rock surfaces known to have been exposed to the Sun's rays for 42 centuries display no detectable sign of disintegration, whereas the same rock types buried nearby beneath the desert sand, where there is just a little moisture, show clear signs of decay. Furthermore, laboratory experiments suggest that heating and cooling alone either achieve little or work slowly, whereas heating and cooling in the presence of moisture produce almost immediate effects. Apart from the local effects of the ephemeral high temperatures achieved in bush fires, little weathering is today attributed to thermal expansion and contraction. The forms once related to this process now are generally attributed to contact with moisture.

Pressure-release hypothesis

Similar doubts relate to the interpretation of sheet structure, the massive (up to nine metres) slabs of rock that commonly are developed in crystalline rocks such as granite but that also occur in sedimentary strata. Though once considered an insolational (derived from the Sun's heat) effect, they have for many years been accepted as a manifestation of offloading or pressure release. In fact, they often are called offloading joints. The pressure-release hypothesis is based on the following argument. Granites crystallize deep within the Earth's crust under conditions of high hydrostatic pressure, so that the very fact that granite is now exposed at the Earth's surface in itself proves that erosion of a considerable thickness of overlying material has occurred and that there has been a drop in vertical loading. During the erosional unloading, the granite tends to expand in response to the decreasing pressure. Such expansion is radial and upward, in the direction of least stress; in detail, it is perpendicular to the land surface. The radial stress is relieved by the development of fractures that are tangential to the stress—that is, roughly parallel to the land surface: hence, the occurrence of arched or gently dipping joints that subdivide the granite into the massive slabs known as sheet structure. An essential feature of the hypothesis is that the form of the land surface predates and, in a broad sense, determines the geometry of the sheeting joints.

All joints are in a sense due to pressure release. All are near-surface features and presumably all disappear at depth. Moreover, many surfaces recently exposed from beneath ice masses display numerous joints that parallel the glacially eroded surfaces, and it is difficult to look beyond pressure release in explanation of these forms, which are, however, essentially superficial. But deep-seated fractures are another matter; sheet structure extends to at least 90 metres. Considerable evidence suggests that the offloading hypothesis is not everywhere applicable and that it should not be used unquestioningly. The very residuals, frequently inselbergs (island mountains), in which sheet structure is commonly displayed, are preserved because they are under compression, not under tension as demanded by the expansive pressure-release hypothesis. In some areas the dip of the sheet structures is opposed to the slope of the land, rather than parallel to it; sheet structure affects sedimentary and volcanic sequences that have never been deeply buried, and in some areas the ages of land surfaces and the sheet structure are the reverse of that implied by offloading. Faulting and lateral compression in the crust, residual from past or continuing earth movements, appear to offer an alternative explanation of sheet structure and to account more satisfactorily for the field evidence.

Thermal expansion and contraction, and the pressure-release hypothesis, have been considered at some length because they demonstrate how a too-ready acceptance of seemingly reasonable ideas can lead to incorrect or incomplete interpretations. Most accounts of weathering processes are similarly oversimplified, if not fallacious. Although some are more soundly based than others, none is completely understood, and it is better to work from evidence of what has taken place rather than from what presumably should happen in given circumstances.

Freeze–thaw cycles and effects

In many subarctic regions, outcrops of finely bedded rocks carry a veneer of platy debris. In these areas, where the vegetation blanket is thin and discontinuous, the common oscillations of temperature around the freezing point affect the superficial layers of rock. Water lodged in numerous cracks is frozen and expands, exerting a pressure sufficient to widen the fissures in which it rests. On thawing, the water rests lower in the now wider crack, until it again expands on further freezing. Such repeated expansion and contraction of contained water causes many rocks, especially those that are naturally well bedded, to break down into slabs and plates. Many close observations of strata, in regions where the temperature fluctuates about the freezing point, indicate the effectiveness of the freeze–thaw mechanism.

Crystal-growth pressures

Crystallization of such salts as sodium chloride and gypsum is also cited as a cause of rock disintegration, particularly in arid regions. There seems little doubt that the pressures exerted by crystal growth can rupture weakly cohesive rocks. Roofing tiles on the eastern (windward) side of Port Phillip Bay (Victoria, Australia) commonly are disintegrated by the crystallization of salts blown in by spray, and clays also can be broken up and disturbed. But whether salt crystallization can shatter such strongly cohesive rocks as fresh granite is problematical. Certainly salts are a common product of granite weathering and in arid climates efflorescences of sodium chloride can be seen on sheltered rock surfaces far distant from the coast. It seems likely that such expansion may at least contribute to the total weathering process. Some writers, however, considering the problem of weathered crystalline rocks in Antarctica, where, at present, there is never any moisture because of the consistent extreme cold, suggest that salt crystallization is responsible for the shattering of the rocks. It is, however, pertinent to ask what the source from which the salts crystallize is, if no moisture is available, and to point out that other common difficulties facing weathering studies seem to be involved: did the weathering observed occur in relation to the present climate or in a slightly different set of climatic circumstances? Is the weathering observed in Antarctica taking place now, or is it, as it were, inherited from a former, warmer and moister climate (in which case, freeze–thaw action could be invoked)?

The role c plant growth an organisms

There is no doubt that tree roots can force aside considerable blocks of rock and widen pre-existing joints during growth. Even the tiny roots, or hyphae, of lichens can penetrate along crystal boundaries and cleavages and can accomplish some physical disintegration. Burrowing animals such as rabbits and termites open up avenues for other agencies, particularly moisture. The burrows of earthworms long have been recognized as a significant element in the soils; worms penetrate about 120 centimetres (four feet) below the surface and they pass about four tons of soil per hectare, on average.

Chemical processes. The understanding of physical weathering processes is beset with difficulties, but the chemical reactions at work in the regolith are even more complex. Although it is possible to set down plausible reaction formulas, the fact is that most of these are so oversimplified as to be misleading. For purposes of exposition, however, it is convenient to isolate certain processes and their results.

Solution effects

No mineral is chemically inert, and many are to an appreciable extent soluble in water. Some, like rock salt, gypsum, and limestone, react strongly with water and either go into solution or form products that are soluble. Even quartz is to some extent soluble in water. Some minerals are more soluble in salt water than in fresh water: orthoclase, one of the feldspars and a common constituent of crystalline rocks, is 14 times more soluble. It is likely that solution is in many instances the first stage of chemical weathering. Solution also produces many distinct forms such as pits and fretting patterns, and its widespread significance is indicated by the vast quantities of material carried in solution by rivers.

Materials in solution are translocated vertically within the weathering profile (vertical section that exhibits weathering alterations) or may be carried great lateral distances and be precipitated far from their place of origin. Because of the translocation of minerals in solution (as well as of fine particles in solid state) within profiles, distinct horizons or pans rich in iron oxides, lime, silica, or gypsum may be developed. In reality, lateral accession is indistinguishable from vertical relocation, but whatever the precise source of material, salts are taken in solution from one part of the weathering profile and concentrated in another. Nodular or sporadic accumulations develop first, but these coalesce to form continuous, frequently massive, sheets. Because they tend to form tough, impermeable layers, they are called duricrusts. Bauxite is an example of such an accumulation; it essentially is an alumina-rich crust.

The cause of the localized precipitation of such minerals is not clear. In some instances, reaction with groundwater with contrasted chemical properties may be involved. In the case of calcrete, a lime-rich crust, it has been suggested that evaporation of groundwater may give rise to precipitation of lime at the upper fringe of the water table and near the land surface. On the other hand, plants markedly accumulate some minerals, and it has been suggested that silcrete, a silica-rich crust, for example, is fixed by plants and then added to the regolith when the plants die and decay. Whatever the mechanism involved in their formation, great sheets of laterite, calcrete, and silcrete have accumulated in various parts of the world.

Hydration and hydrolysis

Water and contained radicals and gases combine with various minerals to form new minerals, sometimes of volumes significantly different from those of the original. These processes are known as hydration (the addition of water) and hydrolysis (the addition of hydroxyl [OH^- ions]). Thus, iron readily combines with water and oxygen to form various hydrated iron oxides that are responsible for the yellow and red coloration of many weathered profiles. Orthoclase, a common constituent of acid crystalline rocks, as well as of some sediments, reacts with water and carbon dioxide to produce a clay, typically kaolin, and a soluble salt and silica. All common rock-forming minerals except quartz are converted to clay minerals by chemical weathering, principally by hydration and hydrolysis. Thus mica is hydrated to hydrobiotite and vermiculite, and eventually to chlorite, but kaolinite and gibbsite are other common products of hydration. Hydration is important on its own account, but it also prepares mineral surfaces for oxidation and carbonation and generally enables ionic transfer to occur more readily.

Oxidation, or the formation of oxides, occurs in the aerated zone of soils, probably by interaction of the oxygen dissolved in water. Oxides are a common constituent of the regolith, and much takes place through the agency of bacteria that derive energy from the oxidation of iron and other elements. In waterlogged anaerobic sites bacteria can bring about chemical reduction (loss of oxygen). Thus, sulfates are reduced to sulfides, and organic material is reduced by fermenting bacteria.

Carbonation, or the reaction of carbonate or bicarbonate ions with minerals, is an important intermediate step in the weathering of such minerals as feldspars, and carbonic acid, though weak, is a potent solvent in nature.

Silicification and desilicification can cause the conversion of one type of clay to another. Thus, in tropical lands desilicification of micas gives rise to kaolin and iron oxide, or, if taken further, to bauxite (gibbsite) deposits.

As is the case with physical processes, considerable chemical weathering is attained with the aid of organic agencies. Humic acids, produced by the decay of organic matter, promote weathering generally. The hyphae of lichens can apparently extract certain radicals from minerals. Green algae in tidal pools raise the pH (acidity–alkalinity index) of the seawater during the day, and the emission of carbon dioxide (CO_2) by algae and invertebrates at night brings about the solution of carbonate. Humus in general helps conserve moisture in soils and hence aids weathering in a number of ways.

Although numerous individual processes have been proposed, the details of many are obscure; most are complex, and many open the way for the activities of others. Hence the complex of processes embraced by the term weathering is more effective than the total of its individual components. Water, acting both directly and indirectly as a solute and as a vehicle for various radicals, is undoubtedly the most important single factor in weathering, though biological agencies have increasingly been recognized as of great significance. But however achieved in detail, weathering produces a relatively thick mantle of debris.

CONTROLS AND RATES OF WEATHERING

Several factors control the type and rate of rock weathering.

Mineralogical composition. Of the common rock-forming minerals, the order of susceptibility to chemical attack is the same as the order of crystallization from magma (silicate melt that yields igneous rocks on cooling): olivine is most vulnerable, followed by plagioclase, biotite, potash feldspar, muscovite, and quartz. Thus a rock like quartzite, composed overwhelmingly of quartz particles and cement, will weather only slowly, while a basalt, on the other hand, rich in ferromagnesian minerals and plagioclase, will suffer rapid alteration.

Texture. In fine-grained rocks the total surface area of the constituent minerals is very great. These surfaces are quite prone to chemical attack, but such minerals tend to be closely interlocked and do not offer avenues for physical attack. Thus the coarse-grained rocks are most susceptible to such processes.

Fracture pattern. Fractures such as joints and faults are, if open, avenues of weathering readily exploitable by various environmental agencies, but particularly by water. Thus greatly shattered and fractured rock masses are much more rapidly weathered than are essentially monolithic masses.

Climate. From what has been said so far it is clear that many weathering processes achieve their optimal activity in specific climatic zones. Thus, freeze–thaw occurs and is effective in subarctic regions, but not in arctic areas where temperatures are too consistently low. Insolational heating and cooling effects are greatest in arid tropics. And chemical weathering is most important in the humid tropics, where moisture and humic acids are abundant and where temperatures are consistently high. Here the rate of chemical reactions is three times greater than in temperate regions, although a decrease in the viscosity of capillary water and its more rapid circulation in the regolith compensates for this to some extent.

Erosion and topography. Erosion may, by the removal of weathered debris, expose new rocks to weathering (renewal of weathering). On the other hand, stable conditions may encourage the development of deep weathering and thick regoliths. Steep slopes and high peaks are well drained, but valley floors, and particularly enclosed depressions, receive water and solubles. Hence, in poorly drained areas, waterlogging and reduction may be characteristic, or there may be marked precipitation of dissolved salts, depending on the prevailing climate.

Time. Prolonged weathering of rocks produces minerals different from those that result from brief reactions and removal of the debris. Reaction series resulting from continued hydration, silicification, or desilicification, for instance, are known and some of them have already been outlined.

Man. Quite apart from man-induced erosion, quarrying, and excavations of various types, man has stimulated weathering by his pollution of the air. The industrial release of sulfur has produced sulfuric acids in the air, which has affected not only man-made buildings but also natural rock exposures (see also below *Physiographic effects of man*).

Rates of weathering

Turning to rates of weathering, various isolated estimates have been made. Limestone tombstones in northwestern England are said to require 250 to 500 years to weather to a depth of 2.5 centimetres (one inch). After 45 years the ash falls associated with the 1883 eruption of Krakatoa displayed strong leaching of alkalis and some of silica. At Soufrière in the West Indies, soil development and reforestation on volcanic ash were "normal" after only 30 years. On the banks of the Murray River, Australia, weathering of a sandy limestone has varied between one and 30 centimetres (12 inches) per century.

Such estimates, though of interest, are all meaningless in a geological sense. So many factors are involved that great variations in the rate of weathering occur in the same area; aspect, rock type, exposure to erosion, and other factors have their effect. Even deep-weathering profiles, the initiation of which can be dated (as, for example, on lava flows of known age), are of little help. Is the weathering going on now, or did the whole profile develop in the past? Was the weathering achieved during a short time span when climatic conditions were especially suitable, or has it been going on steadily over a long period? It is not possible, at present, to be confident on these points, so general and imperfect is knowledge of Earth history and weathering processes. (C.R.T./Ed.)

Slope movements

Very few landscapes are totally flat. Nearly all plains contain isolated hills rising from the main land surface, and most plateaus have valley-side slopes. The form of hillslopes, using the term for the sloping surface of both hills and valleys, consitutes a distinctive landscape feature. Of major importance in shaping hill-slopes are earth movements that occur under the influence of gravity when the stability of a slope is disturbed either by natural forces or by human interference. Earth movements of the most varied nature depend on the interaction of a number of factors, including the angle of slope, nature of materials, and time.

The great diversity of slope movements can be classified according to the mode and rate of movement, form of the surface of sliding, and the type of earth material moved. The most common downhill movement, almost imperceptible because of its small rate, is called creep, a category that comprises mass movements of a very wide scale, ranging from the creep of slope debris through outward bulging to long-term gravitational slides on mountainous slopes.

Another large group of slope movements includes landslides (landslips), which are rapid movements of earth materials separated from the underlying stationary part of the slope by a definite surface. When the movements occur along a predisposed surface where there is jointing or bedding, the slide is designated as a glide. If a free fall is involved in the movement of blocks of solid rock, the phenomenon is called a rockfall. Slides along newly formed curved surfaces are called slumps, and mass movements involving high-water content are called earthflows and mudflows. On steep mountainous slopes, torrential rains may produce debris avalanches. A special case of flowage and slipping is solifluction; *i.e.*, the movement of a thawed surface layer on the frozen substratum.

Significance to man

Slope movements, which may become a serious economic problem in their extent and recurrence because they often cause great damage to the property and life of men, can be an insurmountable hindrance to human activity. Many disastrous landslides and rockfalls are known to have destroyed whole towns and caused hundreds of deaths. Extensive depreciation of agricultural land, as well as of wooded areas, may also be caused by slope movements. Major slides result in the complete removal or extinction of forest growth; trees are uprooted or become dry. Highways and railways traversing areas susceptible to sliding are not infrequently interrupted by landslides, particularly when the stability of slopes is disturbed during construction. There are cases in which railway lines have had to be abandoned because of permanent danger of sliding movements and consequent high maintenance costs. Slope movements frequently produce serious difficulty during major engineering construction projects, such as tunnels and dams.

Adverse indirect effects of earth movements on slopes include clogging of valleys by landslides that impound temporary lakes that endanger the downstream reaches by flooding after the natural dam has collapsed. Sudden landslides and rockfalls along sea shores also have very disastrous indirect effects. In the Norwegian fjords, for example, landslides often provoke high-water swells up to several tens of metres that are detrimental to the inhabited coast. (Q.Z./Ed.)

FACTORS PRODUCING SLOPE MOVEMENTS

The variety of landslide types reflects the diversity of factors that are responsible for their origin. Such factors include the character and structure of rocks, the angle of slope, the soil or debris cover, climatic and groundwater conditions, and time.

Debris cover on slopes is generally liable to downslope movements. As physical and chemical weathering disturb the cohesion of newly exposed rocks, further material for sliding is continuously supplied. The stability of rocks is also impaired by chemical changes induced by percolating water.

Slopes composed of resistant permeable beds that are underlain by weak, incompetent, impermeable rocks, such as clays, are quite prone to sliding. The underlying clays become saturated with water and are squeezed out by the weight of the hard rocks above. In stratified rocks the contributing factor is the downslope dip of beds. If this slope is undercut by river erosion, then the stability of beds is disturbed, and slipping takes place.

The vegetation cover is also important. The roots of trees maintain stability by their mechanical effects and contribute to the drying of slopes by absorbing part of the groundwater. Deforestation of slopes impairs the water regime in the surface layers and facilitates erosion. The grass mat is gradually worn away so that weathering proceeds more intensively and produces free debris.

Earth movements on slopes are frequently induced by an increase of slope angle. This may be caused by natural or artificial interference; for example, by the undermining of the foot of a slope by stream erosion or by excavation. Exceptionally, the angle of a slope becomes steeper as a consequence of processes such as subsidence or uplift of the Earth's crust. The increase in slope gradient increases the shear stress within the rock mass, and this disturbs the equilibrium.

Effects of earthquakes, groundwater, and climate

Tremors produced by earthquakes also affect the equilibrium of slopes by evoking a temporary change of stress. Some disastrous rockfalls in high-mountain areas (*e.g.*, Peru in 1970) are known to have been caused by earthquakes. In loess (fine-grained silts) and loose sands, shocks can disturb the intergranular bond and thus lower the shear strength. In water-saturated fine sands—those in which water occupies the pore space between all sand grains—and some kinds of clays (quick clays), shocks may result in a displacement or rotation of grains leading to a sudden liquefaction of soil.

The conditions of slopes are greatly affected by groundwater. Flowing groundwater exerts pressure on soil particles that impairs the stability of slopes. In fine sand and silt, groundwater washes out fine particles, and the strength of the slope is weakened by the cavities that are formed. Moreover, soluble cement may be removed and, consequently, the cohesion of rock and the shear strength decreased. If the groundwater is under pressure it acts to uplift the overlying impermeable beds, thus decreasing the stability of slope.

The factors listed above often combine with the influence of climatic conditions, particularly the amount of precipitation and frost activity. Rain and ice meltwater, for example, penetrate into joints (rock fractures) producing hydrostatic pressure, and the increase in pore-water pressure in soils induces a decrease of shear strength.

It has been observed that, under certain climatic conditions, slope movements occur repeatedly in extremely humid years; measurements of rainfall amounts have confirmed that there is a direct relationship between precipitation and frequency of landslides. In Czechoslovakia the monthly rainfall for 70 years of record shows a striking coincidence with the recurrence of slope movements. Systematic examination of such records makes it possible to predict the renewal of slope movements in areas liable to recurrent sliding and to warn of imminent danger.

In clayey rocks, the deleterious effects of atmospheric water are heightened when the rainfall comes after a long dry period: clayey soils are desiccated and shrunken, and water easily penetrates deep into the fissures. The disturbance occurs on the lubricated layer in which the water accumulates.

Among the numerous factors inducing earth movements on slopes, that of time must not be omitted. As the agents change in the course of time, several phases of development occur. These range from the first signs of disturbance of the equilibrium to general loosening of the mass, which is then propelled into motion, travels downslope, and is gradually deposited.

TYPES OF SLOPE MOVEMENTS

Rockfalls. Rockfall refers to the abrupt movement of loosened blocks or complexes of solid rocks detached from rock walls. Rockfalls are distinguished by a very high velocity resulting from the free fall. Their size ranges from isolated stones to enormous masses of rock. The stones and blocks that fall build up talus, or debris fans, at the foot of mountain slopes that in some places coalesce into extensive aprons. The slopes of these fanshaped deposits range from 25° to 40°, depending on shape and form of stone fragments. Floods often carry away the loose material of these fans and deposit it farther downslope.

Origin and occurrence of rockfalls

The origin of rockfalls depends on the morphology of the slopes and on the jointing and fracturing of rocks. Rockfalls are frequent in mountainous areas, particularly in valleys that have been overdeepened by glaciers. Factors contributing to the loosening of blocks are climatic conditions, chiefly weathering, wedging effects of freezing water in joints, hydrostatic pressure of water in open fissures, and pressure of growing roots. The movement can be triggered by the undercutting of steep slopes by erosion or excavation, by earthquakes, or, exceptionally, by thunderbolts.

Hundreds of rockfalls have been recorded from young mountain ranges, such as the Alps, Carpathians, Himalayas, Andes, and Rocky Mountains. One of the largest was the rockfall in the valley of Bartang River, in the Pamir Mountains in 1911. A rock mass of about 4,800,000,000 cubic metres (6,300,000,000 cubic yards) fell and dammed the valley, creating a lake 75 kilometres (47 miles) long and 262 metres (859 feet) deep. When large rock masses drop into a lake or fjord, dangerous huge waves flood the coast. In the Norwegian fjords numerous rockfalls have occurred, and catastrophic results were caused in part by the suddenness of the event.

If a rockfall involves an extremely large mass of rock, the freely falling body detached high from the mountain face may move downslope with a speed of up to 215 kilometres (135 miles) per hour; the blocks are shattered to rock fragments, and the mass movement takes on the character of a flow. Rock streams in the Alps and other high mountains are mostly explained in this way.

Rockfalls are also frequent on rocky shores of lakes and seas, as well as on steep concave banks of erosive river valleys. In these cases the rockfall is an important agency in modelling the cliffs and contributes to the recession of the coastline; the rocks are eroded by waves, the cliff becomes oversteepened, and the upper part of the wall collapses.

Rockfalls also can be responsible for the recession of waterfalls, particularly when hard rocks overlie less resistant beds. The existence of Niagara Falls, for instance, is threatened by large-scale rockfalls. In 1954 about 185,000 tons of rock collapsed because of undermining by water erosion.

Creep and bulging. The simplest form of creep is the slow, almost imperceptible downslope movement of soil particles and rock debris. Creeping of loose rock fragments is the result of a number of processes, particularly those related to climatic agencies. During the winter months movement is facilitated by the loosening of rock fragments and heaving by frost. Upon thawing in spring, the particles, which have been lifted at right angles to the slope, fall back vertically under the effect of gravity, so that they move a small distance downhill. The creep of slope debris can be compared to plastic deformation occurring in a snow bed on a mountain slope. The rate of movement is greatest near the surface and decreases downward. The expansion of stones by heat and shrinkage on cooling also contribute to the downslope movements, because they are not equal at the upslope and downslope sides of stones. Subordinate effects can be produced by plowing, cattle treading, burrowing by animals, and wedging by plant roots.

Clayey surface layers move slowly downhill by the action of plastic deformation. These movements do not usually develop a discrete slide surface but occur over a wider zone. They are limited to the surface layer, which does not surpass the depth of temperature and humidity effects. Though the deformations amount to only a few millimetres, during the long intervals embraced by geological time they appear as a steady creep of slope deposits.

Creep results in the bending of beds (Figure 1). Friction that is active between the creeping debris and the surface of the bedrock produces a gradual bending of the bed faces cropping out at the surface. The dragged-out and disrupted layers of the bedrock become part of slope deposits, thus increasing their thickness.

Deformation by bulging

In addition to surficial creep, a slow movement disturbing the equilibrium conditions of a slope takes place under suitable conditions at a greater depth below the surface. These earth movements are caused by the squeezing of soft rocks from beneath the more solid overlying rocks. In English literature this phenomenon is designated as

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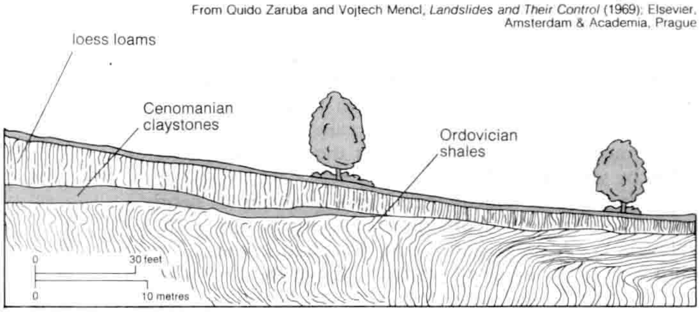

Figure 1: Creep and bending of beds in a loam pit at Prague, Czechoslovakia (see text).

bulging; in some countries the result is called a valley anticline. This process involves the plastic flowage of the underlying rocks along a system of surfaces of potential sliding. The instability of the slope is perceptible only during a longer time interval, when the minute deformations have reached measurable values. In the advanced stage these creep phenomena may grade into true landslides.

In some regions the squeezing out of soft rocks in the lower parts of the valley slopes is so widespread that it causes serious economic problems. Bulging was first described from an area of iron-ore mining near Northampton (central England), where valley sides are composed of solid Jurassic limestones and shales (136,000,000 to 190,000,000 years old) that overlie soft Lias clays. Although the beds are nearly horizontal, the near-surface beds of solid rocks are inclined into the slopes, whereas the clays at the foot of the slopes are squeezed upward and contorted. In the initial stage, the bulging appears as a slight anticlinal bend of beds; with advancing deformation the clays are bent into folds, and even small faults may form at the foot of the slope. Although the disturbances extend to several tens of metres in depth, from the geological point of view they are surficial phenomena because they disappear both downward and into the slope where the beds preserve a normal subhorizontal course. This deformation can be interpreted in terms of the upward squeezing of plastic substance from the loaded medium into the unloaded one. The stress that caused the heave of clays results from the difference in the loading of clays in the bottom of the valley and under its slopes.

Figure 2 shows an example of bulging in a river valley near Ostrava (Czechoslovakia). The valley occurs in marly shales of Cretaceous age (65,000,000 to 136,000,000 years

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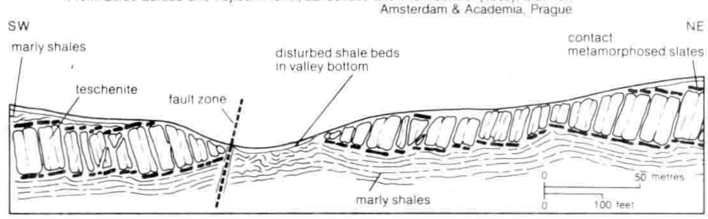

Figure 2: Squeezing-out of marly shales on the valley bottom of the Lucina River, Czechoslovakia.

old) that are pierced by sills of very firm volcanic rock called teschenite. The section shows that the teschenite body is broken into several blocks by a system of faults running roughly parallel to the valley. The soft shales that are squeezed out by the heavy rock blocks move toward the stream, which gradually carries them away. The main deformations probably occurred about 230,000 years ago, during the Pleistocene Epoch (10,000 to 2,500,000 years ago), because the steps between the blocks are filled with slope debris and loess loam.

The possible development of such deformations under present climatic conditions of the temperate zone has been the subject of much consideration. In general, squeezing of the substratum may occur whenever (1) the soft rocks within a limited area are released from the weight of the overlying rocks, and (2) this release gives rise to stresses, which, even if they do not surpass the shear resistance of plastic beds, may, if they last long enough, result in deformations.

Because the character of motion is essentially like that of creep, some long-term deformation of mountain slopes can be included in this group of earth movements. They consist of movement along planes of separation, such as planes of stratification, schistosity, or jointing.

Many analogous phenomena have been observed in young mountain ranges such as the Alps or Carpathians and occur most frequently on the slopes formed of phyllites, mica schist, and other metamorphic rocks that can be deformed by differential movements.

Movements related to creep can grade into sliding when a slide surface develops in the course of time and the movement is accelerated.

Landslides and debris slides. Landslides include a multiplicity of downslope movements, which, in contrast to creep, occur along well-developed surfaces. A landslide occurs when the stability conditions of the slopes are disturbed either by the increase of shear stress imposed on the slope or by the decrease in shear strength of the rock building up the slope. The factors and processes that may provoke the change in the state of equilibrium have been mentioned previously.

Slide characteristics

The sliding movements involve bedrock or surficial deposits, but very often both bedrock and its cover are involved. The part of the slope that moves separates from the remaining mass along the plane of least resistance (Figure 3). This slip surface is commonly formed by a bedding, joint, or fault plane. The movements are generally rapid and originate when the planes of separation dip downslope and their continuity is disturbed at the foot of the slope. In stratified rocks with smooth, even bedding planes, the dip of beds is usually the maximum inclination at which the slope is permanently stable. If the beds are undercut by stream erosion, they maintain their position only by friction, which increases with the roughness and unevenness of the bedding planes. Friction can be reduced by climatic factors: freezing and thawing of interstitial water, or by hydrostatic pressure of water in the joints if the free outflow of water is obstructed. Failure can also be provoked by the increase of slope angle, as a result of uplift of the Earth's crust.

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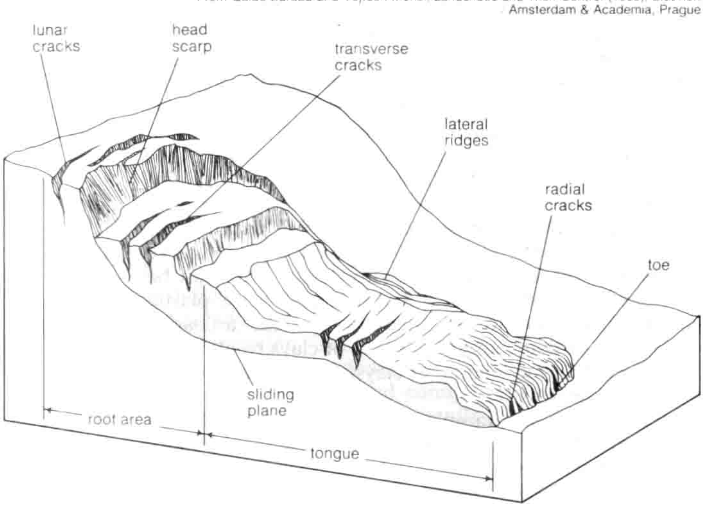

Figure 3: Principal parts of a landslide and characteristic cracks.

Rock slides on bedding planes or other surfaces of separation may be disastrous in mountain areas where, because of great height differences, the movement attains an acceleration nearly equal to that of a rockfall. In steep mountains the conditions are favourable for rock slides, because streams with steep gradients cut readily into the bedrock, and the adjustment of the slopes cannot keep pace with erosion. The stability of slopes is particularly threatened when the dip is toward the valley.

Debris slides—movements of shallow slope debris and weathering materials above the bedrock—are categorized with landslides. Debris slides involve materials of only a few metres in thickness but may cover wide areas in some regions. On disturbed slopes, various stages of slipping are observable, from initial fissuring of the weathered layer up to advanced forms with several generations piled on top of one another. Debris slides generally occur after a heavy rainfall or spring thaw. Rainwater soaking into the soil lubricates the surface bed of the unweathered rock, on which the disrupted top layer slips down. In the other case, during the freezing of the ground the surface layers of debris are enriched by water rising by capillary action toward the surface from the lower unfrozen beds. During the spring thaw the water produces slaking of the surface layer and reduces the internal friction and the stability of slope.

Slumps, earthflows, and debris avalanches. The type of slope failure known as slump is common in homogeneous, poorly consolidated clayey rocks, such as clays, marls, claystones, and clayey shales. Slide-promoting forces are increased by the undermining of the slope by erosion or by excavation or by the overloading of its upper part.

Form and occurrence of slumps

Slumps have a characteristic form. The mass usually tears off along a concave head scarp and moves down a curved slip surface to accumulate at the foot of a slope, spreading laterally. Inside the sliding rock mass some minor scarps originate, so that it is broken into blocks that are tilted toward the slope. In the depressions of the hummocky surface, water accumulates into small lakelets, thus contributing to the instability of the slope. On either side, the mass is squeezed into longitudinal ridges that are often striated. The whole body is cut by cracks of different arrangement. Slumps may gradually increase by backward caving of the head scarp. The movement generally occurs along partial cylindrical surfaces, but the resulting slip surface is somewhat distorted. The depth and shape of a slump adapt to the geological structure of the slope.

Slumps of large dimensions are frequent on river valley slopes and sea coasts. Along some valleys deep slumps occur side by side, and the interlocking alcoves enlarge the slide area laterally. Slide material that weights the foot of the slope may help to restore the equilibrium of the slope, but tongues of the slumps are generally washed away during floods or by waves, and advancing erosion includes further movements. Typical examples of these slump movements are known from the coast of England near Folkestone (Figure 4), and from the valleys of some rivers (Volga, Moskva, Dnepr, and others) in the Soviet Union.

On the lower parts of slopes, slumps frequently grade into flows, the shape of which is controlled by the topography. They move as sheets or streams that fill the erosion gullies or valleys. These movements, called earthflows, debris flows, or mudflows, according to the material and consistency of the mass, represent a separate type of slope movement, one that differs from the slumps because of the higher water content that produces motion of flow type.

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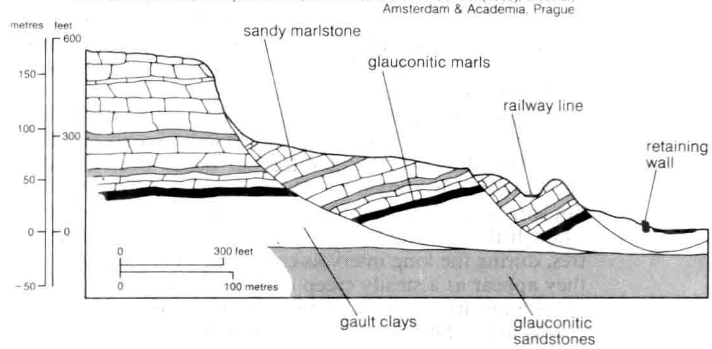

Figure 4: Landslide on the seashore near Folkestone, Kent.

Role of water saturation

Earthflows generally originate in a large basin in the upper part of the slope, where debris and weathering material have accumulated. The movement is usually triggered by heavy rainfall. Interstitial water increases the weight of the debris cover and greatly reduces its shear strength. The loosened mass flows toward the foot of the slope, forming a loaf-shaped bulge. When it flows down a valley, it moves at a rate greater than is achieved over the slope surface, because water-saturated debris packed into a narrow tongue contacts the substratum over a smaller area; consequently, the friction is smaller. When all loose material has been emptied from the source area, the earthflow gradually becomes stabilized and overgrown with vegetation. If part of the debris remains there, heavy rainfall may provoke further movement, and the bulge at the toe will be overridden by younger material.

With increasing velocity earthflows grade into debris flows. Mountainous debris flows are sometimes called avalanches or Muren, a term currently used in the Alpine countries. As a rule, they originate above the timberline in gorges filled with rock detritus. During torrential rains, debris and larger stones travel at a tremendous speed down the stream channels. The material is unsorted, and the ratio of solid particles to water is about 1:1. The rates of flow are so great that moving trains have been trapped and buried beneath debris avalanches. The Muren are very disastrous and often result from too great deforestation of mountain slopes. Scarcity of vegetation is also responsible for debris and mudflows in arid and semi-arid regions. This group of slope movements also includes volcanic mudflows. Volcanic explosions are usually accompanied by torrential rains that wash ash and ejected material downward as a mushy mass. Debris flows occasionally originate on glacier-covered dormant volcanoes. When volcanism recurs, the increase in heat before and during the eruption suddenly melts the glaciers, and the waters bring much debris downhill.

Solifluction. Solifluction is a combined flow and slip movement, which involves the surface layers on slopes in the subarctic region and, to a lesser extent, high mountain regions. The deeply frozen surface layers thaw to only a small depth during a short summer. Meltwater and precipitated water saturate the soil because they cannot percolate into the frozen, impermeable substratum, and the waterlogged bed flows downslope as a dense sludge. Solifluction may occur even on a rather moderate slope, because the soil moves readily on the frozen substratum.

Another somewhat special type of flow movement involves clay sediments of marine origin and is common in Scandinavia and Canada. These sensitive clays, called quick clays in Norway and Leda Clays in Canada, cover flattish areas situated about two hundred metres above sea level. The strength of these sediments progressively decreases because of the decrease of salts in the pore water. The ground and atmospheric water impoverish the salt content by osmotic processes. The decrease in salt concentration in pore water goes hand in hand with the decrease of bond between the clay particles and the bound water, and thus with the decrease in strength. Remarkably, the drop in strength is greatest toward the end of the process. The loss of strength results in a flow movement of unusually great rapidity. The failures of quick-clay slopes are treacherous because they may affect areas that are nearly flat.

Failure of quick clays

One of the largest failures of this type occurred near Verdalen, north of Trondheim, in Norway in 1893. A layer of sensitive clay was laid bare by stream erosion. The liquefied clay, with a volume of 55,000,000 cubic metres (72,000,000 cubic yards), flowed down in 30 minutes. The dense liquid covered an area of 8.5 square kilometres (3.3 square miles) and destroyed 22 farms. The flat terrace in the broad valley of the river Veddalselva is suggestive of absolute safety today, but the monument bearing the names of 111 people who were killed in 1893 testifies to the catastrophic possibilities.

CHARACTERISTICS OF UNSTABLE SLOPES

Slopes disturbed by sliding movements show a characteristic configuration: the head scarp is arcuate with a spoon-shaped depression, and the topography on the downslope side is hummocky and irregular. In active landslides the features are clear cut, whereas in the dormant, temporarily inactive landslides, the forms are effaced by rainwash and erosion or covered by vegetation. An important characteristic is the shape of the slope in cross section. Even a very ancient landslide is recognizable from its convex bulged toe made up of the accumulated slipped mass.

The growth of trees on unstable slopes reveals the presence and age of sliding movements. Trees, which on unstable ground become tilted downslope, tend to return to a vertical position during the period of rest so that the trunks are conspicuously bent. From the younger, vertically growing trunk segments, the date of the last earth movement can be inferred.

For distinguishing the slide areas, the presence of particular plants can be of help. Horsetail (especially *Equisetum maximum*) and coltsfoot (*Tussilago farfara*) are good indicators of slopes that are prone to sliding. Horsetails contain 50–60 percent silica and 19–30 percent potash in the ash. This high content suggests the presence of potassium and hydrated silicates in the soil and explains why horsetails thrive on sliding areas formed of potassium-rich (*e.g.*, glauconite-bearing) rocks.

The earth processes and factors that cause slope movements and the characteristics noted are intensively studied because they present serious hazards. Geologists and specialists in soil and rock mechanics endeavour to decipher the reasons for slope instability and to determine how fast and how far the loosened rock mass will move. Although some slope failures will remain unpredictable for a long time, at least those disasters caused by human interference can be avoided. (Q.Z.)

Fluvial processes

Over much of the world the reduction of mountains, the building of plains, and the sculpture of the landscape is brought about to a large degree by the flow of water. As the rain falls and collects in watercourses, the process of erosion not only degrades the land but the products of erosion become themselves the tools with which the rivers carve the valleys in which they flow. The process varies over time and from place to place. Materials eroded from one location are transported and deposited in another, only to be eroded and redeposited time and again before reaching the ocean. At successive locations the riverine plain and the river channel itself are products of the interaction of the mechanics of transport by the flow and the characteristics of the sediments brought down from the drainage basin above.

The fluid in a river is not pure water. Not always visible, the load of the river may be carried in solution, in suspension, or dragged along the bed. Solutes and particulate matter are both organic and inorganic. Neither the discharge of the water nor the related rates of erosion and deposition are constant in time or in space. Steep, narrow, rock-walled canyons may be excavated by corrosion of flowing water armed with abrasive particles aided by corrosion through chemical action. Elsewhere, sediments may be deposited to form broad alluvial fans, floodplains, or river deltas in lakes along the river course.

ENTRAINMENT AND TRANSPORT OF SEDIMENTARY PARTICLES

Erosion and transport of sedimentary particles is initiated when the drag, or shear stress, exerted on the boundaries of a natural channel by the flowing water is sufficient to detach a particle from the boundary. A particle of a given size and weight will begin to move when the shear stress exceeds a component fraction of the weight of the particle under water. In general, increasing flow accompanied by increasing velocity and shear stress results in progressive entrainment of particles from the bed and higher rates of transport.

There are essentially two distinct physical modes of transport of sediment. Bed load is that portion of the material in transport that is in continuous or partial contact with other particles on the bed; thus the weight of the moving particles is supported by contact with grains in the bed. In contrast, the suspended load is born up by the fluid eddies within the flow itself. Because the turbulent eddies both near the bed and in the fluid vary in intensity from moment to moment, the motion of the individual particles is highly erratic from place to place and moment to moment. In a statistical sense, however, at any distance above the bed, an equilibrium concentration is maintained with the number of particles settling through a given level balanced by an equal number of particles thrust upward by eddies. The concentration of suspended sediment at each level above the bed is a function of the settling velocity of the particles and the intensity of shear stress or turbulent exchange in the fluid at that level. Large particles, which settle most rapidly, are found near the bed, and progressively smaller particles are carried at greater distances above the bed. Similarly, where the particles are all of the same size, larger concentrations of suspended sediment will be found closer to the bed. If the particles are very fine, they may be nearly uniformly distributed with depth.

Modes of sediment transport

The curves in Figure 5 show the way in which sediment concentration varies with depth for particles of different sizes or settling velocities at a constant condition of flow. Settling velocity is primarily a function of the size of the

Particle size and settling velocity

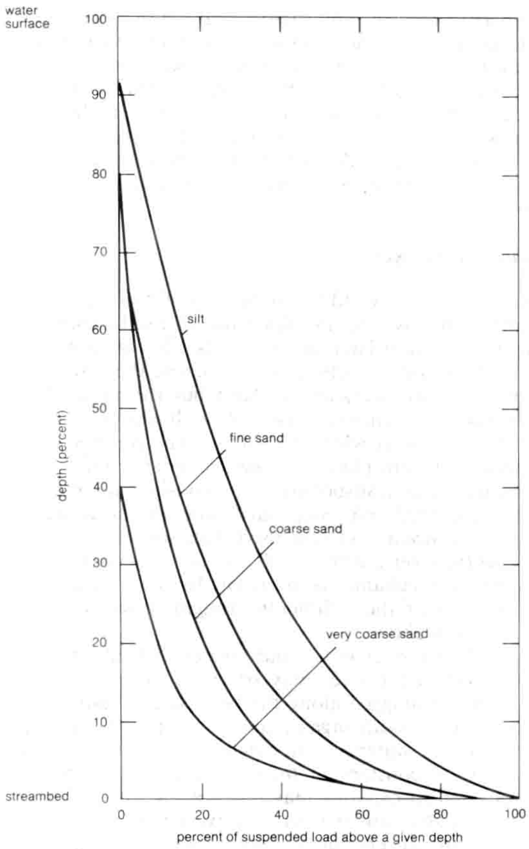

Figure 5: *Distribution of suspended load with depth for particles of different sizes.*
Coarse particles are concentrated closer to the bed, while finer particles, such as silt, are more uniformly distributed with depth.

particle. Assuming the availability of a range of particle sizes in a river, as the flow fluctuates and velocity or shear stress changes, particles transported near the bed at one time may be entrained and move in suspension at a higher flow. With diminution of the flow larger particles cannot be maintained in suspension and will settle to the bed.

Where the bed of a natural channel is comprised of noncohesive particles such as sand, as velocity and depth increase with increasing flow the particles in motion no longer behave solely as discrete individuals but instead the bed itself is deformed, resulting in the formation of sand ripples, dunes, and waves. As the rate of transport increases, these bed forms change progressively with changes in depth and velocity. The sand near the bed moves in a zone close to the dune or ripple surfaces. Usually, particles are carried up the backslope of the dune and deposited usually after sliding down on the steeply dipping face producing the downstream movement of the dune. Because deformation of the mobile boundary influences the flow distribution itself, the interaction between the flow, the boundary form, and sediment transport is complex. These interactions, and the progressive changes in bed forms observed in natural channels during passage of a flood, are discussed below, under *Erosion and deposition in natural channels.*

MATERIALS TRANSPORTED BY NATURAL RIVERS

The materials actually transported as a clastic or particulate load as well as those in solution in natural rivers are a reflection not only of physical laws governing the competence and capacity of the flow to transport material but also of the availability of materials themselves. For example, pure limestone terrain containing no insoluble materials would supply no sediment to streams draining the region. With the exception of occasional blocks broken off from limestone canyon walls, the rivers would carry only dissolved materials, and the stream beds would be devoid of sand, silt, and gravel bars. Such extremes, of course, are rare and, in general, rivers transport a mixture with coarser sediment in the bed load, finer particles in suspension, and ions in solution. Although these distinctions represent mechanisms of transport rather than sources of material, the sources are reflected in both the composition and relative abundance of each fraction.

The relative proportions of dissolved and particulate or clastic load as well as the distribution of the clastic fraction between bed load and suspended load are highly variable in nature (Table). In general, the percentage of clastic load increases as the climate becomes more arid. Less effective solution and weathering in arid regions apparently reduce the supply of dissolved solids to the stream channel system, whereas in more humid regions larger quantities of dissolved material are made available to streams. Topography, and particularly the composition of the bedrock, also can markedly influence the composition and quantity of the load. The Saline River in Kansas, for example, has a salinity of 1,000 parts per million by weight, a function of the saline deposits found in the watershed. Similarly, extreme concentrations of sediment have been measured in some rivers such as the Yellow River in China, where sediment comprised 40 percent of the total fluid, or tributaries of the Colorado River, in which 60 percent of the flow was sediment. In general, concentrations of suspended load in natural rivers vary from several hundred to 10,000 parts per million (1 percent) by weight. Some sand channels, such as the Loup River in western Nebraska, may transport 50 percent or more of the total load as bed load, but values of about 10 to 20 percent are probably more common.

EROSION AND DEPOSITION IN NATURAL CHANNELS

Equilibrium conditions and the migration of meanders

The supply and movement of sediment is intimately involved in the determination of the form and pattern of the river channel. A natural river flowing in sediments of its own making can maintain a stable configuration in two ways: first, by a balance of forces in which the drag of the fluid tending to erode the perimeter of the channel at any point is equalled or exceeded by the frictional or cohesive forces tending to resist the eroding force, and second, by maintenance of a rate of deposition equal to the rate of erosion. In contrast, a channel incised in rock is less free to adjust its form and pattern through deposition and erosion, but the bed of the channel will rise and fall with changes in the rate of transport of debris. The first case, in which a precise balance is maintained between the erosive force and the resistance of the boundary materials without any erosion, is relatively rare in nature. The banks of such a channel in noncohesive material are roughly parabolic and, if the flow is large, the channel cross section will consist of a wide central portion with a flat bottom and banks at each side roughly parabolic in form. Ideally, the maintenance of such an erosional equilibrium requires a

Dissolved and Suspended Load in Selected Rivers in Different Climatic Regions

river and location	drainage area (square miles)	average suspended load (000,000 tons per year)	average dissolved load (000,000 tons per year)	dissolved load as percent of total load
Canadian River near Amarillo, Texas	29,700	6.41	0.12	1.8
Green River at Green River, Utah	40,500	19	2.5	12
Mississippi River, at the mouth	1,245,000	344	123	26
Delaware River, at Trenton, N.J.	12,300	1.0	0.83	45
Juniata River, near New Port, Pa.	3,354	0.32	0.57	64
Amazon River, at mouth, Brazil	2,722,000	499	242	33
Congo River, at mouth, Congo	1,425,000	31	98	76

delicate balance between the opposing forces, a relatively uniform material, and a constant flow, conditions reproducible in the laboratory but only approximated in nature.

A natural river channel in which the rate of erosion is balanced by the rate of deposition and the outflow of sediment to the reach equals the inflow can maintain a stable form while moving laterally across the alluvial plain. A meandering river is the most common illustration of this process. Erosion takes place on the outside of each bend near the point of maximum shear stress as a result of the curvature of the flow. Deposition on the opposite bank is associated with transverse flow near the bed and with slack water eddies adjacent to the thread of the current. Because the locus of erosion is downstream from the point of maximum curvature, progressive erosion in the downstream direction is associated with progressive deposition as the entire channel bend moves downstream.

Sand and gravel bars. Where the channel boundaries are straight, either because of the nice adjustment of discharge, gradient, and sediment, as is the case in some canals, or perhaps as a result of vegetation or channellization by man, sediment and sedimentary forms such as sand and gravel bars may move downstream in a progressive and orderly fashion. The movement of sediment and the configurations of the channel bed associated with such movement may be rather arbitrarily divided into three phases. First, channel bars may be deposited along the banks at sequential positions alternately on one side of the channel and the other. This configuration of alternating bars of gravel or sand in a straight channel is not unlike that which would be observed if the bends of a meander were "pulled out" to make a straight channel. Their spacing of roughly three to five channel widths appears to be related to the discharge and to the width of the channel. Second, sand and silt may move as dunes or ripples, a mode of transport determined by particle characteristics and the interaction of boundary form and the flow. A third mode of sedimentary deposit involves successive movement and deposition of discrete particles. Larger particles may move different distances depending upon their size, shape, and specific gravity. Boulders will move less frequently and, on the average, more slowly than smaller particles, producing a differential rate of downstream migration of particles of different sizes.

Accumulation and movement of gravel and sand in bars appear to resemble the movement of what is called a kinematic wave. Discrete particles do not move independently of one another but interact or interfere with each other in much the same way as automobiles on a highway. As a result of this interaction, the particles accumulate in groups or agglomerations that move downstream as "waves." The average downstream rate of movement is then represented not by the movement of the individual particles but rather by the average rate of movement of the group of particles constituting the wave. This wave phenomenon is similar to that observed on a highway crowded with automobiles, where, when the number of automobiles is low, the automobiles interact very little and each moves at its own rate. With an increase in the number of automobiles, however, interaction takes place and groups begin to form such that there are agglomerations and openings in successive positions along the highway. The celerity or rate of movement of these waves is determined then by the density of particles (or automobiles) in a given length of stream channel (or road), by the characteristics of the particles, and by the conditions of flow.

Flow and transport variation with time

Because the flow of water in natural channels is not constant, each mode of transport in a river also varies with time. The alternation of high and low flow in most of the rivers of the world not only influences the rate of transport but also the attendant forms of the channels themselves and the deposits associated with them. In the natural world the time scale of variations in flow may be matters of minutes, days, years, decades, or millennia. Peak flows from thunderstorm rainfall may occur in a matter of minutes after the start of heavy rain in streams or in urban rivers. Storms of longer duration may produce high water lasting for days or weeks. At another time scale are the seasonal or annual variations such as the cyclical rise and fall of the Nile each spring as a result of snow melt and rains in the headwaters, a variation common to many major river systems such as the Colorado, the Rio Grande, the Ganges, and the Yukon. Lastly, successions of dry years may be followed by wet ones. Such climatic variations are less periodic in occurrence and may encompass periods from decades to thousands of years in duration.

Variations in flow rate are associated with variations in transport. The dissolved load, responding to the characteristics of the source rocks as well as the flow, often decreases in concentration as a result of dilution of the salt concentration by the direct runoff from streams. In contrast, suspended load generally increases with increasing flow. Wash load (*i.e.*, materials such as clays and fine silts) may be readily removed from the watershed, for example, when spring rains follow the melting of snow and particles of soil are detached by cycles of freezing and thawing in early spring and late winter. The first spring rains readily remove the prepared materials to the streams. The transport of bed load also increases in response to increase in velocity and shear stress accompanying the passage of higher flows or floods. As in the theoretical or laboratory condition, the change in flow in the river channel is accompanied by a change not only in the concentrations of dissolved, suspended, and bed materials but by changes in the form of the bed itself as the flow increases in depth and velocity. Changes in the form of the bed associated with the passage of a high flow in a river are shown in Figure 6, whereas Figure 7 shows concurrent changes in flow, velocity, and depth.

From *Bulletin of the U.S. Geological Survey* (1960)

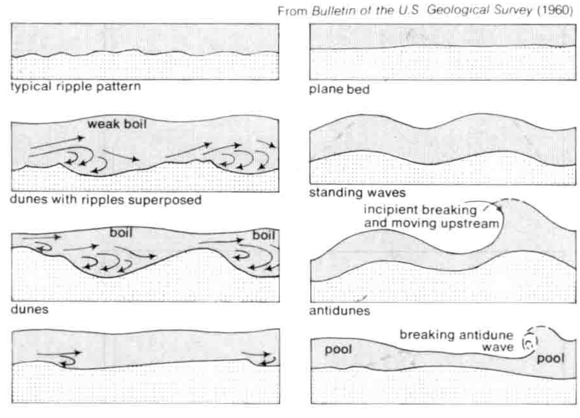

Figure 6: *Forms of bed roughness in alluvial channels.*
This sequence of forms is related to the depth and velocity of flow and to the rate of sediment transport.

Dunes

Bed forms. As noted earlier, after the initiation of movement, ripples and dunes develop on the bed. With increasing depth and velocity of flow in the channel, the dunes grow in size. Experimental and field observations indicate that the amplitude or height of the dune increases until the velocity of flow over the dune prevents further accumulation on the crest. At equilibrium, dunes may cover the bed in the same way that they do the windblown surface of a sandy desert. If velocity increases more rapidly than depth, a transition occurs, dunes begin to be erased, and all or part of the bed will be planar, with dunes covering the remainder. Beyond this transition, continuing increase in velocity leads to the formation of antidunes. Dune forms on the bed move upstream as a result of the displacement of material by scour at the upstream position while net transport continues in the downstream direction. The formation of antidunes is associated with the formation of standing waves on the water surface. Unlike dunes, the crests and troughs of antidunes are in phase with the crests and troughs of the surface waves.

With declining flow in the channel and reduction in velocity and depth, the sequence is reversed. The bed is gradually transformed from the antidunes to the mixed and planar bed, to dunes, and thence to an irregular bed of sand if the flow declines to zero. A smooth bed is rarely if ever attained inasmuch as declining flow both erodes the dunes and deposits finer sediments on the channel bed.

Thus the forms remaining in a sand channel are usually dissected dunes and ripples.

Each of the successive configurations of the bed is associated with a particular set of flow conditions, and with changing discharge a complex interrelationship exists between the geometry of the boundary as defined by the configuration of the channel bed, the concentration of sediment, and the flow parameters such as velocity and depth. In some cases the same discharge on the rising stage of a flood may be associated with a high velocity and a relatively low depth, while on the declining stage the depth may be higher and the velocity lower (Figure 7). Resistance to flow is greater on the falling than on the rising stage. This difference is presumed to be due primarily to decline in flow so rapid that there is insufficient time for the transition of the bed forms. Dunes of larger amplitude associated with the higher flow remain at the lower stage and these larger forms provide a greater resistance.

The scale of dune forms may be exceedingly large. In a large river such as the Mississippi, at depths of flow from 21 to 27 metres (70 to 90 feet), dunes or dunelike forms of major proportions have been observed, with amplitudes from 5 to 12 metres and longitudinal spacing from crest to crest of 15 to 152 metres. Dunes can be seen in the beds of many rivers and creeks carrying sand, and dunes of cobbles and gravel occasionally have been observed after great floods.

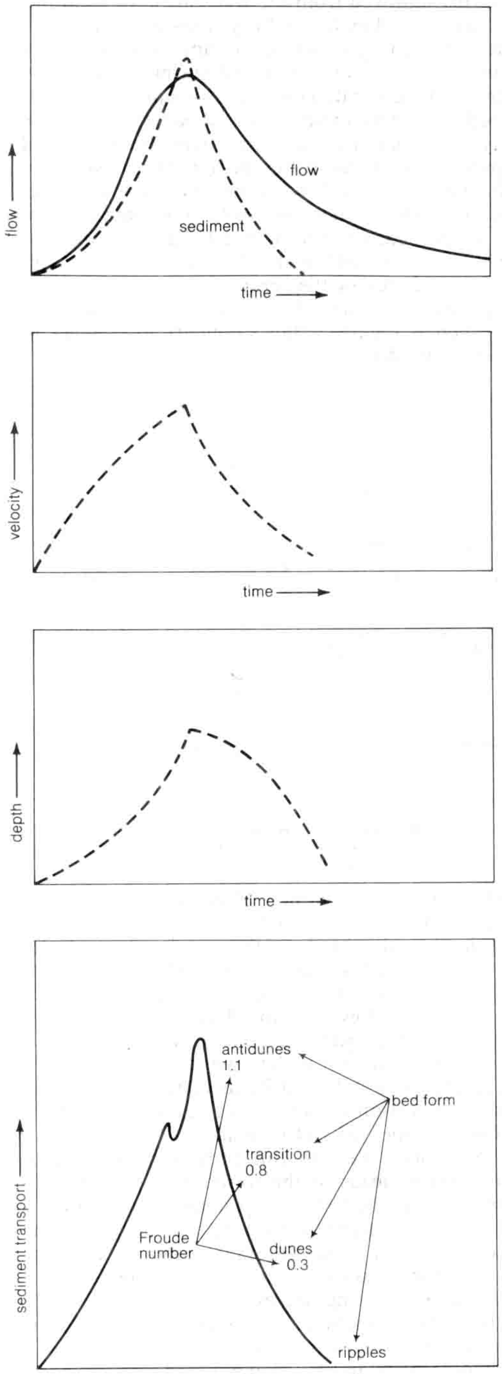

Figure 7: Sequential changes in flow (discharge), velocity, depth, and sediment transport associated with the passage of a flood in a natural channel with a sand bed. Changes in bed form and the approximate Froude number at which such changes might be observed are shown in association with changes in the rate of sediment transport.

Magnitude and frequency of flow events

Variations in flow produce variations not only in the rate of movement of particles of different sizes but also in the distances that these move in successive intervals of time. A large boulder in a small trout stream may be moved only by rare and very large floods. In contrast, material in solution is transported continuously as long as there is flow in the stream. Between these extremes lies a continuous range. Small particles are carried more often and for greater distances during each rise in the flow than are successively larger ones. Huge boulders may be moved a matter of centimetres or perhaps hundreds of metres by a flood that occurs on the average perhaps once in a hundred years. Sand may be transported continuously in a stream where the flow never declines below the point of incipient motion of the sand particles. Thus individual particles move in steps of varying lengths that depend upon the duration of the flow in which they are transported and their size, shape, and distribution.

Flood events of large magnitude but infrequent occurrence may move very large particles as well as large quantities of material but only for short periods of time. In contrast, more frequent events will move smaller sizes but over longer periods of time. In sand bed channels of ephemeral streams, such as those in the southwestern United States, most of the movement of both suspended and bed materials may occur during relatively infrequent events, which recur perhaps four or five times per year. These few floods not only mold the channel form, but they may transport 80 to 90 percent of the total annual load carried by the river. In contrast, where flow is perennial and of relatively large magnitude, more of the dissolved and suspended load may be transported by flows of modest magnitude throughout a large part of the year. Under such circumstances, the relative contribution of the large and infrequent flood event is thus of lesser significance. To the extent that generalization is possible, existing evidence suggests that the greater the variability of stream flow and the larger the particles to be carried, the more significant are the floods of large magnitude and infrequent occurrence. Similarly, where flow is less variable, dissolved load a significant proportion of the total load, and the suspended load composed of fine particles, more frequent events assume greater importance.

Scour and fill. In a given reach of channel, fluctuations in flow, accompanied by changes in the rate of transport of both suspended and bed material, produce a scouring or filling of the channel bed. In a straight channel this may amount to a few centimetres. In bends or narrow reaches, scour may be to depths of a metre or more and in rock-walled canyons as much as six to nine metres (20 to 30 feet). The cumulative effect in any given year may result in the temporary lowering of the stream bed. The bed may be lowered in some years and raised in the next or succeeding years. Over a period of time the average elevation will remain constant. A similar process of erosion and deposition characterizes the lateral or down-valley movement of the channel. In any one year erosion may exceed deposition, but over a period of years the pattern and form of the channel remain the same. Thus, the equilibrium referred to earlier constitutes an adjustment to the quantity of water and sediment delivered to the channel, an adjustment maintained within the natural variations in flow and sediment load normally experienced in a given climate or hydrologic environment. Equilibrium of the river channel then is associated with a constancy of climate viewed as an average over a period of years.

A progressive increase in the amount of rainfall on the watershed, a change in the vegetative cover, or a combination of such factors, however, may produce a progressive change in the hydrologic conditions governing the river channel at any point. Such changes might be climatic, they might result from changes in land use or, for example, they might be brought about by construction of a dam and major reservoir. These long term or progressive changes are referred to as degradation or aggradation in contrast to the processes of scour and fill that encompass fluctuations of short duration. Changes associated with man-made works, in fact, provide a good illustration of the nature of the river response to changes in climate. A reservoir, by impounding flood flows for release during periods of low natural flows, alters the frequency distribution or pattern of river flows. In addition, the reservoir becomes a sediment trap, which reduces the quantity of sediment delivered to a channel below. A river previously in equilibrium with the natural flow and sediment supply is thus subjected to a new set of controls that, in turn, result in a succession of changes in the river channel itself. Thus, where scour took place in one year to be followed by deposition in succeeding years prior to construction of the dam, after construction the elimination of floods and sediment supply produces progressive lowering of the channel bed, or degradation, rather than an alternation of scour and fill (see also below *Physiographic effects of man*).

Changes in hydrologic conditions

In contrast, progressive accumulation, or aggradation of material in channels, also results from changes in climate, from man-made works, or at the interface of land and water where sediments accumulate in deltas. In some locations, such as on the Yellow River, the river bed has risen to well above the surrounding countryside in the delta because successive floods deposited coarse sediment along the river banks and on the bed. Elsewhere, changes in climate have resulted in the delivery of large quantities of sediment to river channels, producing alluviation or filling for great distances over entire channel systems.

Rates of erosion and deposition

Rates of accumulation or degradation are highly variable. For over a thousand years the Nile has risen at an average rate of about one metre (more than three feet) per hundred years. In contrast, during the period of hydraulic gold mining in the Sierra Nevada the Sacramento River rose as much as three metres (10 feet) in 35 years. Valley fills or terraces along major rivers in the great plains of the western United States indicate that some rivers may have cut through the alluvial materials in their valleys at rates on the order of three metres or more in 100 years, although these estimates are crude at best. Below Hoover Dam on the Colorado River, the bed of the channel was lowered ten feet in five years after closure of the dam.

FLUVIAL PROCESSES IN DIFFERENT ENVIRONMENTS

The characteristics of a river and the processes associated with it are determined by a combination of the geology of a region and the climatic conditions responsible for providing the flow in the river. The term geology includes the underlying structure of the region, such as the presence or absence of mountains, as well as the composition and distribution of the bedrock. The latter, in turn, determine the quantity and often the size fractions of the materials delivered to the stream system. In contrast, although the climate may be influenced by the elevation of the region, major aspects of the atmospheric circulation will determine the amount of precipitation and seasonal variations in temperature. Indirectly these will affect the vegetation, characteristics of the soils, and the distribution of runoff or flow to the river system.

Clearly, innumerable combinations of geologic and climatic controls must exist in nature. Each segment or reach of a river will be dominated by a set of conditions of local origin as well as by a set determined from the larger area of the drainage basin upstream. Despite these obvious variations, however, some combinations of controls are sufficiently common to allow rather rough characterization of rivers in different regions. It must be emphasized, however, that the key to understanding why rivers look and behave as they do lies in a knowledge of the mutual interaction of the controlling factors such as discharge, sediment, and rock type. What might be called characteristic regional types simply represent particular combinations of these factors as well as the existence of a control in some regions or environment. Many complex factors are involved in developing a certain type of river or stream; each stream is classified according to the factors affecting it.

Trout stream. The so-called trout stream, or brook, for example, is indeed a recognizable type of river. It is usually characterized by a vigorous flow (of cool water), high gradient, and a bed of cobbles and boulders bordered by mossy banks and trees. Climate and elevation determine

Grant Heilman

Figure 8: A boulder-bed high mountain stream, Crandall Creek, Wyoming.

the availability of the flow, whereas the coarse bed material is a function of the steep gradient determined by the geologic structure. A high gradient permits the flow periodically to move some of the larger cobbles producing in time a series of pools and riffles. In some brooks, particularly those in coarse sand or cobbles, the spacing of the riffles or shallows appears to be proportional to channel width.

Arroyo. A second rather distinctive river in an opposing climatic regime is the arroyo of the Spanish-speaking world, the wadi of the Middle East and Mediterranean region. These dry washes, which experience periodic torrential flows, are otherwise dry sand channels comprised primarily of sands, perhaps dissected dune forms, some gravel, and occasional cobbles, depending on the local geology and topography. Because of the aridity and high temperature of the bed, vegetation is usually sparse. The combination of dry bed, absence of vegetation, and rapid changes in flow permit large variations in the configuration of the sand bed and in the amount of scour and fill. In addition, because such channels are often on relatively steep slopes and in noncohesive materials, they tend to be wide and shallow and subject to high velocities at shallow depths, a condition promoting the formation not only of dunes but of antidunes along with rapid rates of sand transport.

Dry washes

Channels in humid-temperate regions. In contrast to these rivers in semi-arid regions, the well-defined channels of the humid region are a response not only to the regularity of the flow but to the presence of fine sediments and vegetation that stabilize the river banks, emerging point bars, and other deposits laid down by the river. These channels, composed of silt banks stabilized with vegetation, are narrower and less mobile. Not infrequently floods that overtop the banks deposit on the adjacent plain varying amounts of silt, sand, and mud, thus creating a floodplain composed of channel and bar deposits overlain by finer "overbank" deposits. Where the channel is confined or maintains itself within a relatively narrow belt within which it meanders, the fine-grained deposits of floods may predominate in the floodplain.

The preceding descriptions suggest that in alluvial river channels flowing in sediments of their own making, there is a range of channel or river types, from the wandering channel that deposits bars over a vast shallow plain to the