Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Introduction to Geomorphology

Lecture ONE

by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Introduction to the Geomorphology

The word geomorphology derives from three Greek words: gew (the Earth), morfh (form), and logo (discourse). Geomorphology is therefore 'a discourse on Earth forms'. The term was coined sometime in the 1870s and 1880s to describe the morphology of the Earth's surface.

Geomorphology is the science concerned with the form of the land-surface and the processes which create it. Today, geomorphology is the study of Earth's physical land-surface features, its land forms – rivers, hills, plains, beaches, sand dunes, and many others. It is extended by some to include the study of submarine features, and with the advent of planetary exploration must now incorporate the landscapes of the major solid bodies of the Solar System. One focus for geomorphic research is the relationship between landforms and the processes currently acting on them. But many landforms cannot be fully explained by the nature and intensity of geomorphic processes now operating so it is also necessary to consider past events that may have helped shape the landscape. To a significant extent, then, geomorphology is a historical science. Since the land-surface is located at interlace of the Earth's lithosphere, atmosphere, hydrosphere and biosphere, geomorphology is closely related to a wide range of other disciplines (Table 1.1). While having a central interest in landforms, geomorphologists must, none the less, be aware of those aspects of allied disciplines that bear on their subject. Equally, geomorphology has a potential, as yet only partially realized, of making significant contributions to these other areas of knowledge.

Table 1.1 Examples of relationships between geomorphology and allied disciplines

DISCIPLINE	EXAMPLE OF CONTRIBUTION TO GEOMORPHOLOGY	EXAMPLE OF CONTRIBUTION FROM GEOMORPHOLOGY
Geophysics	Mechanisms and rates of uplift	Erosional response of land-surface to uplift
Sedimentology	Reconstruction of past erosional events from a sedimentary sequence	Form of alluvial channels in interpretation of fluvial sediments
Geochemistry	Rate and nature of chemical reactions in rock weathering	Mobilization of elements in earth surface environments
Hydrology	Frequency and intensity of flooding	Sediment concentration in streams
Climatology	Effect of climatic elements on rate and nature of geomorphic processes	Effect of surface deposits and morphology on climatic variables
Pedology	Effect of soil properties on slope stability	Topographic control over soil-forming processes
Biology	Role of vegetation cover in affecting rates of erosion	Topographic control over micro-environments of plant growth
Engineering	Techniques for analysis of slope instability	Identification of morphological features indicative of slope instability
Space science	Context for understanding special characteristics of landform-creating environment on the Earth	Interpretation of planetary landscapes by analogy with terrestrial landforms

Geomorphology investigates landforms and the processes that fashion them. Form, process, and the interrelationships between them are central to understanding the origin and development of landforms. In geomorphology, form or morphology has three facets, ❶constitution (chemical and physical properties described by material property variables), ❷configuration (size and form described by geometry variables), and ❸mass flow (rates of flow described by such mass flow variables as discharge, precipitation rate, and evaporation rate). These form variables contrast with **dynamic variables** (chemical and mechanical properties representing the expenditure of energy and the doing of work) associated with geomorphic processes; they include power, energy flux, force, stress, and momentum. Take the case of a beach. Constitutional properties include the degree of sorting of grains, mean diameter of grains, grain shape, and moisture content of the beach; configurational properties include such measures of beach geometry as slope angle, beach profile form, and water depth; mass-flow variables include rates of erosion, transport, and deposition.

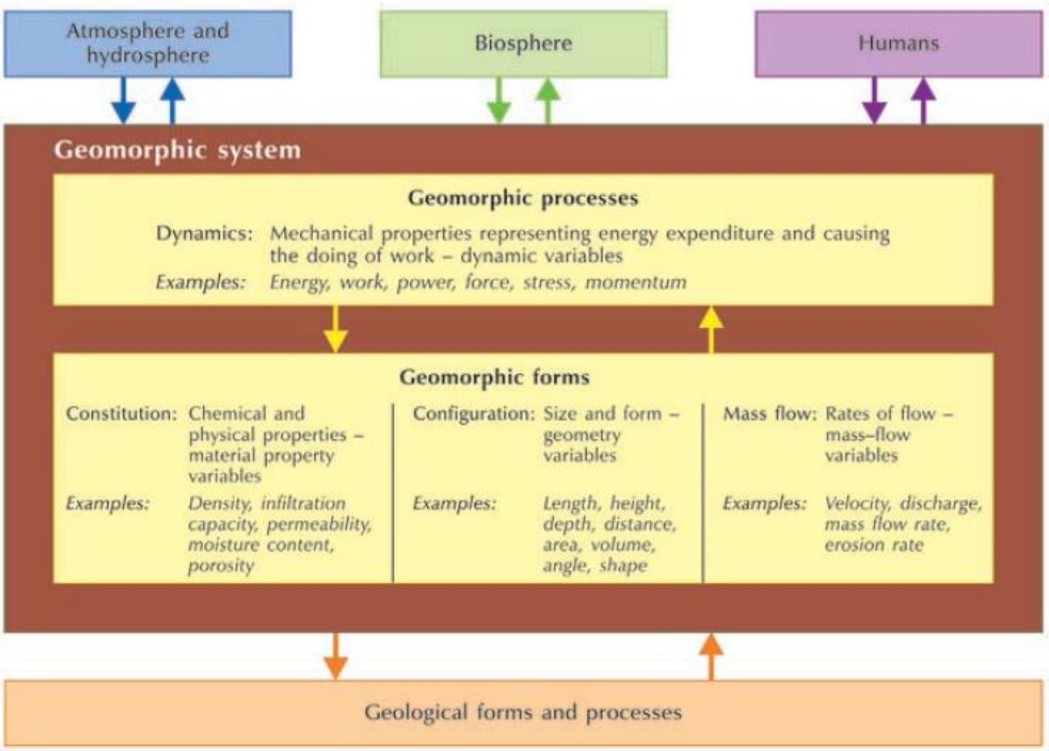

Figure 1.2 Process–form interactions – the core of geomorphology.

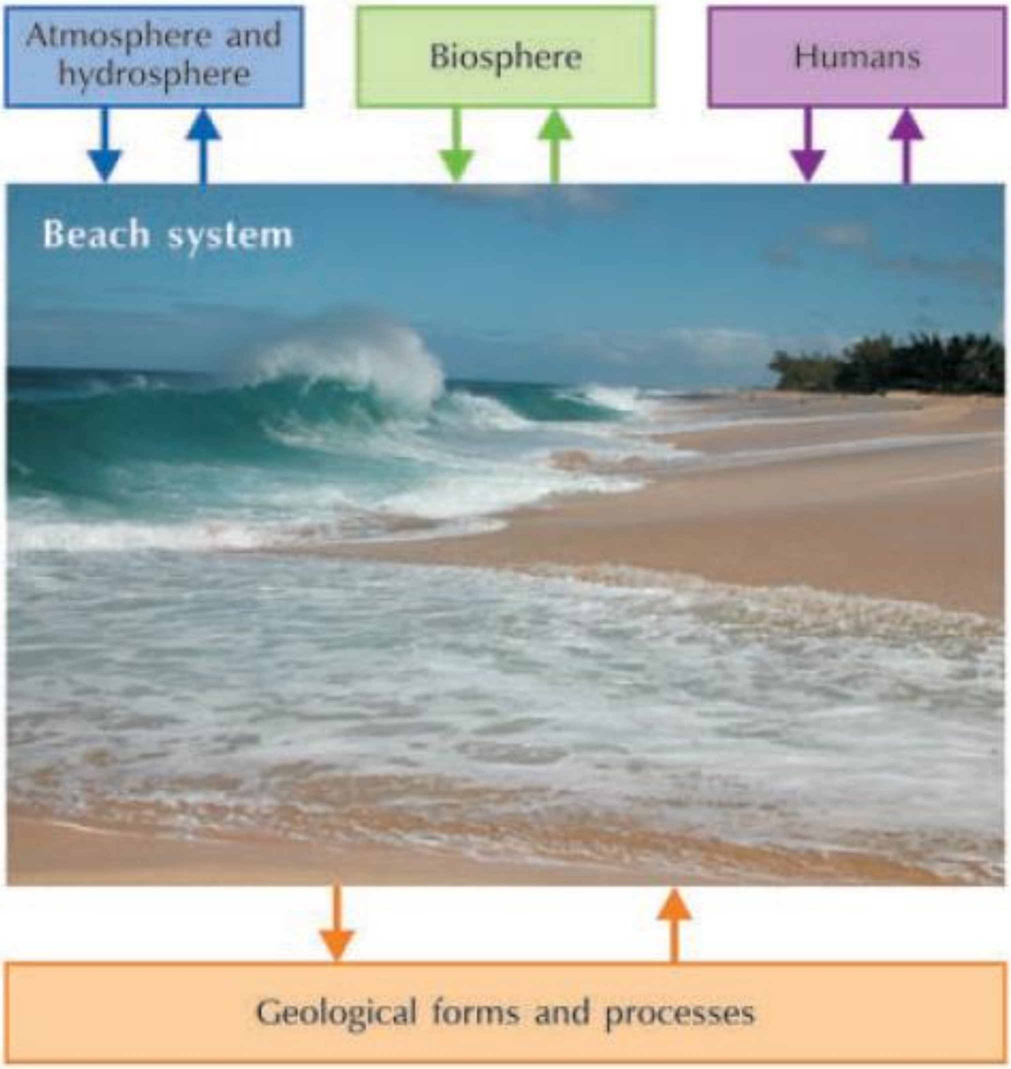

Figure 1.3 Process–form interactions for a beach. (*Photograph by Andy Short*)

Dynamic variables include drag stresses set up by water currents associated with waves (and modulated by tides), possibly by channelled water flowing over the beach, and by wind, and also include forces created by burrowing animals and humans digging beach material. Geomorphic processes are the multifarious chemical and physical means by which the Earth's surface undergoes modification. They are driven by geological forces emanating from inside the Earth (**endogenic** or **endogene processes**), by forces originating at or near the Earth's surface and in the atmosphere (**exogenic** or **exogene processes**), and by forces coming from outside the Earth (**extraterrestrial processes**, such as asteroid impacts). They include processes of transformation and transfer associated with weathering, gravity, water, wind, and ice. Mutual interactions between form and process are the core of geomorphic

investigation – form affects process and process affects form. In a wider setting, atmospheric processes, ecological processes, and geological processes influence, and in turn are influenced by, geomorphic process – form interactions. The nature of the mutual connection between Earth surface process and Earth surface form has lain at the heart of geomorphological discourse. Currently, there are at least *four approaches* used by geomorphologists in studying landforms:

1. A process–response (process–form) or functional approach that builds upon chemistry and physics and utilizes a systems methodology.
2. A landform evolution approach that has its roots in historical geological science (geo-history) and that explores the important historical dimension of many landforms.
3. An approach that focuses on characterizing landforms and landform systems and that stems from geographical spatial science.
4. An environmentally sensitive approach to landforms, systems of landforms, and landscapes at regional to global scales.

The age of Hutton and Lyell

In 1785 James Hutton presented a paper in which he argued that the land surface had been shaped by the slow, unremitting erosive action of water rather than by the catastrophic events advocated by biblical scholars; to the history of the Earth. Hutton saw '***no vestige of a beginning, no prospect of an end***'. His ideas disseminated only slowly until in 1802, five years after his death, John Playfair, his friend and Professor of Mathematics at the University of Edinburgh, restated and elaborated his views with an elegance and clarity that has rarely been matched in scientific writing. In his *Illustrations of the Huttonian Theory of the Earth*, Playfair provided the first detailed and closely reasoned account of several important aspects of landform genesis, most notably the relationship between rivers and their valleys:

"Every river appears to consist of a main trunk, fed from a variety of branches, each running in a valley pro-portioned to its size, and all of them together forming a system of valleys, communicating with one another, and having such a nice adjustment of their declivities, that none of them join the principal valley, either on too high or too low a level, a circumstance which would be infinitely improbable if each of these valleys were not the work of the stream which flows in it".

Hutton's methodology, founded on the belief that the slow but continuous operation of processes observable at the present day provided a sufficient basis for explaining the present configuration of the Earth's surface, was taken up and developed by Charles Lyell in his idea of uniformity. Through his highly significant work, *Principles of Geology* (1830-33), Lyell became the 'great high priest' of what became known as the

principle of uniformitarianism, a concept frequently (but inadequately) summarized by the phrase 'the present is the key to the past'. Lyell's idea of uniformity was far more complex than is often appreciated by many earth scientists and this has led to much confusion as writers have failed to distinguish between its various meanings. In fact four distinct meanings can be identified in Lyell's *Principles:*

1. ***Uniformity of law***: this is the assumption that natural laws are constant in time and space.
2. ***Uniformity of process***: this is the suggestion that if past events can be explained as the consequence of processes now known to be operating then additional unknown causes should not be used. In essence, this is the principle of simplicity adopted in all scientific explanation; if known processes are capable of explaining natural phenomena additional 'exotic' mechanisms should not be introduced
3. ***Uniformity of rate (gradualism)***: this is the proposition that changes on the Earth's surface are usually slow, steady and gradual. Although that major events, such as floods and earthquakes, do take place, but such phenomena are local in extent and that they occurred in the past with the same average frequency as they do today.
4. ***Uniformity of state***: this is the idea that, although change occurs, it is directionless; that is the Earth always looked and behaved much as it does at the present time.

The modern era

A lack of experimental evidence as to the nature and rate of landscape change through time, coupled with a poor level of understanding of the processes responsible for landform genesis, led to increasing doubts among many geomorphologists as to the viability of historical explanation in geomorphology. Foreshadowed by R. E. Horton's remarkable synthesis of drainage basin hydrology published in 1945, the following decades witnessed a growing emphasis, especially in the UK and North America, on both the quantitative analysis of landform morphology (landform morphometry or geomorphometry) and on the field measurement of geomorphic processes. These developments were not so evident in Continental Europe where the earlier tradition of geomorphology founded on the relationship between landform characteristics and climatic zones was strengthened after the Second World War.

The 1960s and 1970s saw a major reorientation of geomorphology in the UK and USA towards the development of predictive models of short-term landform change. These were based on a much improved knowledge of

geomorphic processes founded on a greater understanding of the basic physical principles involved. Indeed, these models often reflected a significant input from research by engineering geologists, particularly with respect to slope stability, the flow of water in river channels and the entrainment and transport of sediment. During this period there was also a rapid growth in applied geomorphology with predictive models being used to assess the likely response of the landscape to changing conditions brought about by human activities, such as land use changes and dam construction.

HISTORICAL GEOMORPHOLOGY

All landforms have a history. Such landforms as ripples on beaches and in riverbeds and terracettes (a **terracette** is a type of landform, a ridge on a hillside formed when saturated soil particles expand, then contract as they dry, causing them to move slowly downhill) on hillslopes tend to be short-lived, so that their history will pass unrecorded unless burial by sediments ensures their survival in the stratigraphic (rock) record. For this reason, geomorphologists with a prime interest in long-term changes usually deal with relatively more persistent landforms at scales ranging from coastal features, landslides, and river terraces, through plains and plateaux, to regional and continental drainage systems. Nonetheless, ripple marks and other small-scale sedimentary features that do manage to survive can provide clues to past processes and events.

Historical geomorphology is the study of landform evolution or changes in landforms over medium and long timescales, usually timescales well beyond the span of an individual human's experience – centuries, millennia, millions and hundreds of millions of years. It brings in the historical dimension of the subject with all its attendant assumptions and methods, and relies mainly on the form of the land surface and on the sedimentary record for its databases.

The foundations of historical geomorphology

Traditionally, historical geomorphologists tried to work out landscape history by mapping morphological (form) and sedimentary features. Their golden rule was the dictum that *'**the present is a key to the past**'*. This was a warrant to assume that the effects of geomorphic processes seen in action today may be legitimately used to infer the causes of assumed landscape changes in the past. Before reliable dating techniques were available, such studies were difficult and largely educated guess work. However, the brilliant successes of early historical geomorphologists should not be overlooked.

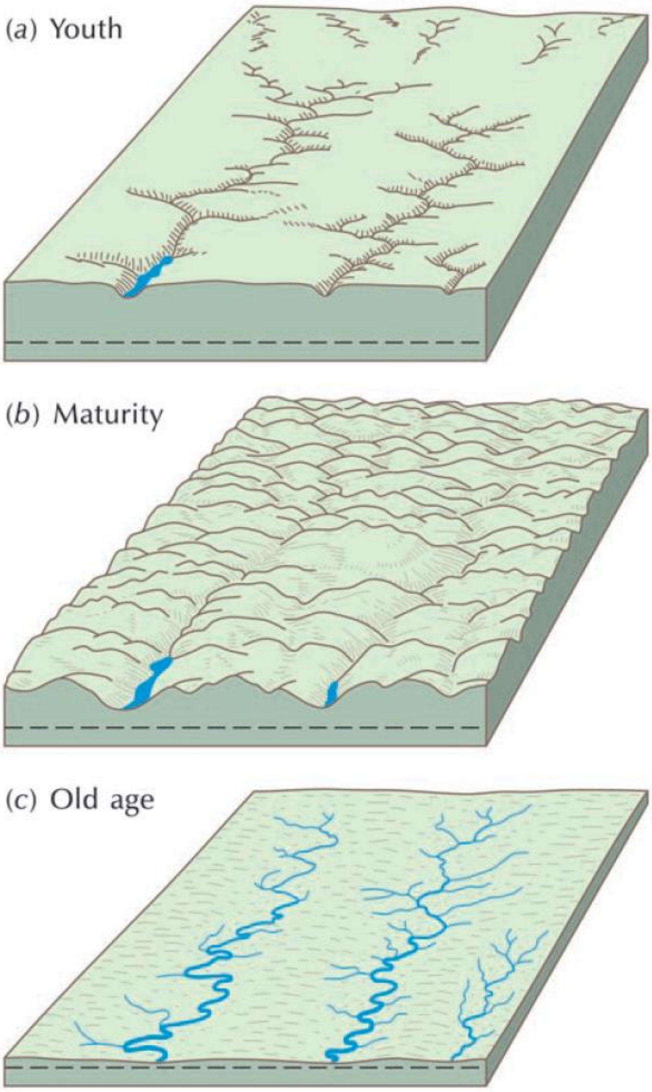

Figure: William Morris Davis's idealized 'geographical cycle' in which a landscape evolves through 'life-stages' to produce a peneplain.

Modern historical geomorphology

Historical studies tend to fall into two groups: Quaternary geomorphology and long-term geomorphology.

Quaternary geomorphology

The discovery in this branch was that landscape changes over periods of 1,000 to 100,000 years display consistent patterns largely forced by the interplay of ❶climatic changes, ❷sea level changes, ❸uplift, and ❹subsidence. Originally, most Quaternary geomorphologists concerned themselves with local and regional changes, usually confining their enquiries to Holocene and Late Pleistocene, so to roughly the last 18,000 years of the 2.6-million-year-long Quaternary. In doing so, they collaborated with other Earth scientists to produce palaeogeographical reconstructions of particular areas at specific times and to build

postdictive or retrodictive models (that is, models that predict in retrospect), so contributing to a revival of historical geomorphology.

Long-term geomorphology

Studies of landforms and landscapes older than the Quaternary, or even late Quaternary, have come to be called **long-term geomorphology.** They include investigations of Cenozoic, Mesozoic, and even Palaeozoic landforms.

PROCESS GEOMORPHOLOGY:

Process geomorphology is the study of the processes responsible for landform development.

Measuring geomorphic processes:

Some geomorphic processes have a long record of measurement. The oldest year-by-year record is the flood levels of the River Nile in Lower Egypt. Yearly readings at Cairo are available from the time of the Prophet Muhammad, Peace be upon him. The amount of sediment annually carried down the Mississippi River was gauged during the 1840s, and the rates of modern denudation in some of the world's major rivers were estimated in the 1860s. The first efforts to measure weathering rates were made in the late nineteenth century. Measurements of the dissolved load of rivers enabled estimates of chemical denudation rates to be made in the first half of the twentieth century, and patchy efforts were made to widen the range of processes measured in the field. But it was the quantitative revolution in geomorphology, started in the 1940s, that was largely responsible for the measuring of process rates in different environments. Since about 1950, the attempts to quantify geomorphic processes in the field have grown fast.

Modelling geomorphic processes

Since the 1960s and 1970s, geomorphologists have tended to direct process studies towards the construction of models for predicting short-term changes in landforms, that is, changes happening over human timescales. Such models have drawn heavily on soil engineering, for example in the case of slope stability, and hydraulic engineering in the cases of flow and sediment entrainment and deposition in rivers. The spur to these advances in landscape modelling was huge advances in computational technology, coupled with the establishment of a set of process equations designated 'geomorphic transport laws' where some of the geomorphologist put it. Figure below shows the output from a

hillslope evolution model; landscape evolution models will be discussed the next lectures.

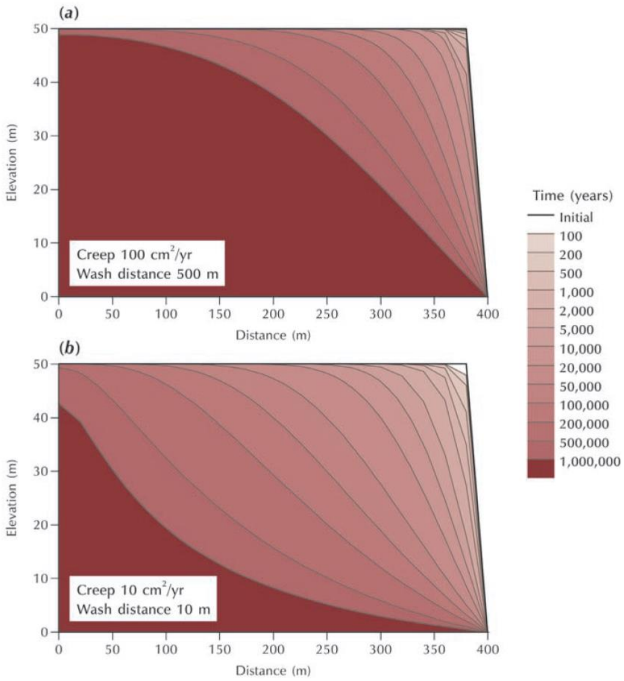

Figure 1.7 Example of a geomorphic model: the predicted evolution of a scarp bounding a plateau according to assumptions made about slope processes using a numerical model of hillslope evolution built by Mike Kirkby. (a) Slope evolution with creep processes running at 100 cm^2/year and no wash processes. (b) Slope evolution with wash process dominating.

Process studies and global environmental change

With the current craze for taking a global view, process geomorphology has found natural links with other Earth and life sciences. Main thrusts of research investigate ❶energy and mass fluxes and ❷the response of landforms to climate, hydrology, tectonics, and land use. The focus on mass and energy fluxes explores the short-term links between land-surface systems and climate that are forged through the storages and movements of energy, water, biogeochemicals, and sediments. Longer-term and broader-scale interconnections between landforms and climate, water budgets, vegetation cover, tectonics, and human activity are a focus for process geomorphologists who take a historical perspective and investigate the causes and effects of changing processes regimes during the Quaternary. The developments in geomorphology partly parallel developments in the new field of **biogeoscience**. This rapidly evolving interdisciplinary subject investigates the interactions between the biological, chemical, and physical processes in life (the biosphere) with the atmosphere, hydrosphere, pedosphere, and geosphere (the solid Earth). It has its own journal – *Biogeosciences* – that started in 2001. Moreover, the American Geophysical Union now has a biogeoscience

section that focuses upon biogeochemistry, biophysics, and planetary ecosystems.

OTHER GEOMORPHOLOGIES

Although process and historical studies dominate much modern geomorphological enquiry, particularly in English-speaking nations, other types of study exist. For example, structural geomorphologists, who were once a very influential group, argued that underlying geological structures are the key to understanding many landforms. Today, other geomorphologies include applied geo - morphology, tectonic geo morphology, submarine geomorphology, climatic geomorphology, and planetary geomorphology.

Applied geomorphology

Applied geomorphology, which is largely an extension of process geomorphology, tackles the manner in which geomorphic processes affect, and are affected by, human activities. Process geomorphologists, armed with their models, have contributed to the investigation of worrying problems associated with the human impacts on landscapes. They have studied coastal erosion and beach management, soil erosion, the weathering of buildings, landslide protection, river management and river channel restoration, and the planning and design of landfill sites. Other process geomorphologists have tackled general applied issues. ***Geomorphology in Environmental Planning*****,** for example, considered the interaction between geomorphology and public policies, with contributions on rural land-use and soil erosion, urban land-use, slope management, river management, coastal management, and policy formulation. *Geomorphology in Environmental Management*, as its title suggests, looked at the role played by geo morphology in management aspects of the environment. *Geomorphology and Land Management in a Changing Environment* focused upon problems of managing land against a background of environmental change. The conservation of ancient and modern landforms is an expanding aspect of applied geomorphology. Three aspects of applied geomorphology have been brought into a sharp focus by the impending environmental change associated with global warming and illustrate the value of geomorphological expertise. First, applied geomorphologists are ideally placed to work on the mitigation of natural hazards of geomorphic origin, which may well increase in magnitude and frequency during the twenty-first century and beyond. Landslides and debris flows may become more common, soil erosion

may become more severe and the sediment load of some rivers increase, some beaches and cliffs may erode faster, coastal lowlands may become submerged, and frozen ground in the tundra environments may thaw. Applied geomorphologists can address all these potentially damaging changes. Second, a worrying aspect of global warming is its effect on natural resources – water, vegetation, crops, and so on. Applied geomorphologists, equipped with such techniques as terrain mapping, remote sensing, and geographical information systems, can contribute to environmental management programmes. Third, applied geomorphologists are able to translate the predictions of global and regional temperature rises into predictions of critical boundary changes, such as the poleward shift of the permafrost line and the tree-line, which can then guide decisions about tailoring economic activity to minimize the effects of global environmental change.

Tectonic geomorphology

This studies the interaction between tectonic and geomorphic processes in regions where the Earth's crust actively deforms. Advances in the measurement of rates and in the understanding of the physical basis of tectonic and geomorphic processes have revitalized it as a field of enquiry. It is a stimulating and highly integrative field that uses techniques and data drawn from studies of geomorphology, seismology, geochronology, structure, geodesy, and Quaternary climate change.

Submarine geomorphology

This deals with the form, origin, and development of features of the sea floor. Submarine landforms cover about 71 per cent of the Earth's surface, but are mostly less well studied than their terrestrial counterparts are. In shallow marine environments, landforms include ripples, dunes, sand waves, sand ridges, shorelines, and subsurface channels. In the continental slope transition zone are sub - marine canyons and gullies, inter-canyon areas, intra-slope basins, and slump and slide scars. The deep marine environment contains varied land - forms, including trench and basin plains, trench fans, sediment wedges, abyssal plains, distributary channels, and submarine canyons.

Planetary geomorphology

This is the study of landforms on planets and large moons with a solid crust, for example Venus, Mars, and some moons of Jupiter and Saturn. It

is a thriving branch of geomorphology. Surface processes on other planets and their satellites depend materially on their mean distance from the Sun, which dictates the annual receipt of solar energy, on their rotational period, and on the nature of the planetary atmosphere. Observed processes include weathering, Aeolian activity, fluvial activity, glacial activity, and mass movements.

Climatic geomorphology

The chief climatic geomorphologist exponents are French and German. Their arguments rest on the not universally accepted observation that each climatic zone (tropical, arid, temperate, for exam ple) engenders a distinctive suite of landforms. Climate does strongly influence geomorphic processes, but it is doubtful that the set of geomorphic processes within each climatic zone creates characteristic landforms. The current consensus is that, owing to climatic and tectonic change, the climatic factor in landform development is more complicated than climatic geomorphologists have suggested on occasions.

SUMMARY

Geomorphology is the study of landforms. Three key elements of geomorphology are land form, geomorphic process, and land-surface history. The two complementary main brands of geomorphology are historical geomorphology and process geomorphology. Other brands include applied geomorphology, tectonic geomorphology, submarine geomorphology, planetary geomorphology, and climatic geomorphology. Geomorphology has engaged in methodological debates over the extent to which the present is the key to the past and the rates of Earth surface processes.

ESSAY QUESTIONS

1. *To what extent are early ideas in geomorphology relevant today?*
2. *Explain why geomorphology encompasses a wide range of approaches.*
3. *Does geomorphology have a future?*

References:

Huggett, R. J., 2011, Fundamentals of Geomorphology, 3rd addition, Routledge Fundamentals of Physical Geography, 533 pp.

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Exogenic processes and landforms

Weathering and Associated Landforms

Lecture TWO

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

The weathering system

Material carried to the sea by rivers, or transported by glaciers or the wind, experiences some degree of chemical decomposition or physical breakdown prior to being eroded. Weathering is therefore an appropriate or suitable place to begin our look at the operation and effects of exogenic geomorphic processes. Geomorphologists are concerned with the rates at which different weathering processes operate as a function of environmental conditions, and with the nature of the weathered material that is produced. But they are also especially interested in how weathering gives rise to specific landforms.

The nature of weathering

Weathering can be divided into those processes involving chemical reactions and the formation of new minerals (chemical weathering) and those that involve only physical changes (physical weathering). Although the differences between these two types of weathering are distinct in theory, in practice they rarely operate separately; rather, the effects of one aid the operation of the other. For instance, a rock shattered through physical weathering will be more liable to chemical weathering because of the increased surface area made available for chemical reactions. Conversely, chemical weathering along micro-fractures in a rock will weaken it and help physical processes break it down more rapidly. ***Weathering*** can be defined as the adjustment of the chemical, mineralogical and physical properties of rocks in response to environmental conditions prevailing at the Earth's surface. For some igneous and metamorphic rocks formed at great depths and at high temperatures and pressures this adjustment can involve a complete transformation of their constituent minerals. Weathering occurs through complex interactions between the lithosphere, the atmosphere, the hydrosphere and the biosphere, and gives rise to three major types of product. Chemical processes lead to the release of compounds in solution and the creation of new mineral products, while physical processes cause the breakdown of the original rock into smaller particles. Dissolved material may subsequently be re-precipitated or be reincorporated into other minerals, but the great proportion is carried by rivers to the ocean.

Chemical weathering

Chemical characteristics of rock-forming minerals

There are two main types of chemical bond existing between atoms contained in the compounds constituting the Earth's rock-forming minerals - ionic bonds and covalent bonds. Atoms with eight electrons in their outermost (valence) shell are chemically stable, but all the elements with which we are concerned have either more or less than this stable number. Those with one additional electron readily lose it and become positively charged ions (cations) (for instance, K+, Na+) and those with one less tend to gain an electron and become negatively charged ions (anions) (for example, F-, Cl-). For other elements the loss or gain of two electrons similarly creates ions with a double charge (for instance, Ca2+, 0 2-). Such ions are stable on their own but only form stable compounds when their electrostatic charge is neutralized by ionic bonding. Some elements, such as oxygen, are able to form both covalent and ionic bonds. This is a simplified picture because in reality most chemical bonds in the majority of rock-forming minerals are intermediate between ionic and covalent (Fig.).

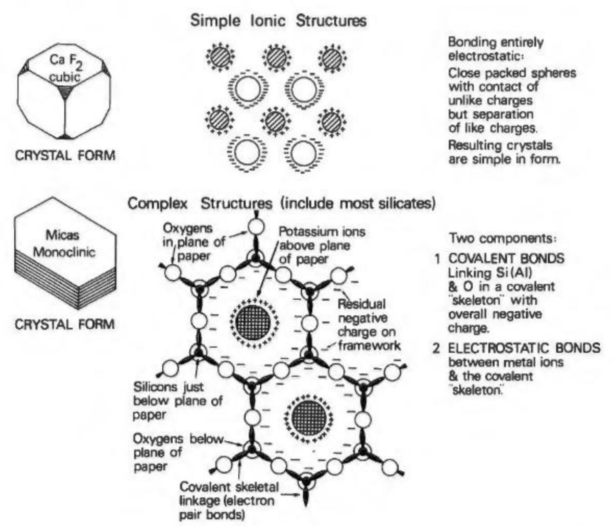


Fig. 6.2 *Ionic and covalent bonding in minerals. Silicate minerals have a basic structure consisting of silicon atoms covalently bonded between oxygen atoms to form tetrahedra with an overall negative charge. This is neutralized through electrostatic bonds with various cations, most commonly potassium, calcium, sodium and magnesium (from C. D. Curtis, (1976) in: E. Derbyshire (ed.)* Geomorphology and Climate. *Wiley, London, Fig. 2.2, p. 34.)*

Chemical reactions: thermodynamics and kinetics

Although chemical weathering reflects the tendency for new minerals to be formed which are stable under conditions prevailing at the Earth's

surface, the rate at which these stable forms are produced is often very slow. This necessitates two complementary approaches to the study of chemical weathering: ***thermodynamics*** considers the ultimately stable forms by analysing the energy changes involved in chemical reactions, while ***kinetics*** focuses on rates and mechanisms of change.

Chemical weathering processes

Solution

Solution (or dissolution) is the simplest process whereby minerals can be decomposed and involves water acting as a solvent. The dissolution of quartz (a crystalline form of silica) provides an example:

$$\underset{\text{quartz}}{SiO_2} + \underset{\text{water}}{2H_2O} \rightarrow \underset{\substack{\text{silica in solution} \\ \text{(silicic acid)}}}{Si(OH)_4^0}$$

The equilibrium solubility of a mineral represents the extent to which it will dissolve in water; it is usually expressed in ppm (parts per million by volume) or 1/mg. Some minerals, such as halite (NaCI), are highly soluble in earth surface environments but others, such as quartz, have a very low equilibrium solubility and dissolve in pure water at an exceedingly slow rate. Equilibrium solubility is affected by the temperature and pH of the environment, while the rate of throughput of water is important in controlling the rate of dissolution. This last variable is particularly significant as the film of water in direct contact with the mineral surface eventually becomes saturated with solutes, thereby hindering further dissolution. Consequently the saturated zone must be constantly flushed by under-saturated water if dissolution is to be an effective mechanism. Some minerals have a marked ability to absorb water into their crystal structure through a reversible reaction known as hydration. This can be illustrated by the hydration of iron oxide:

$$\underset{\text{iron oxide}}{2Fe_2O_3} + \underset{\text{water}}{3H_2O} \leftrightharpoons \underset{\text{hydrated iron oxide}}{2Fe_2O_3 . 3H_2O}$$

This reaction is significant in chemical weathering because it aids other chemical processes by introducing water molecules deep into crystal structures.

Hydrolysis

In addition to acting as a solvent, water may react directly with minerals through hydrolysis. This involves the replacement of metal cations (most commonly K+, Na+, Ca2+ and Mg2+) in a mineral lattice by H+ ions and the combining of these released cations with hydroxyl (OH-) ions. The effects of this process can be illustrated by crushing different minerals in pure water. The resulting abrasion pH indicates the amount of exchange between H+ ions in the water and cations in the mineral, the increase in pH being due to the abstraction of the H+ ions from the water (Table 6.2). The basic hydrolysis reaction can be illustrated by the weathering of the silicate mineral albite (a sodium-rich plagioclase feldspar) to the clay mineral kaolinite:

$$\underset{\text{albite}}{4NaAlSi_3O_8} + \underset{\text{water}}{6H_2O} \rightarrow \underset{\text{kaolinite}}{Al_4Si_4O_{10}(OH)_8} + \underset{\text{silica}}{8SiO_2} + \underset{\substack{\text{dissolved}\\\text{sodium ions}}}{4Na^+} + \underset{\text{hydroxyl}}{4OH^-}$$

Note that some of the silicon is retained in kaolinite and that sodium is removed in solution. This reaction leads to the production of hydroxyl ions and would make pore and surface waters alkaline, yet in most environments these waters are neutral to slightly acid. Consequently, we can conclude that hydrolysis as such is not a realistic weathering reaction.

Carbonation

The bicarbonate ion (HC03-) is invariably present in weathering solutions and is easily the most abundant anion in most surface waters. It is formed from the dissolution and dissociation of carbon dioxide in water in a reversible reaction:

$$\underset{\text{water}}{H_2O} + \underset{\substack{\text{carbon}\\\text{dioxide}}}{CO_2} \leftrightharpoons \underset{\text{carbonic acid}}{H_2CO_3} \leftrightharpoons \underset{\substack{\text{hydrogen}\\\text{ion}}}{H^+} + \underset{\substack{\text{bicarbonate}\\\text{ion}}}{HCO_3^-}$$

Carbon dioxide is fixed from the atmosphere by photosynthesis and enters the weathering system through the intermediary of respiration by plant roots and the breakdown of plant debris by bacteria. Carbon dioxide is consequently abundant in the atmosphere of soils, especially those characterized by high rates of organic activity; it may reach a

concentration of 10 per cent (in comparison with 0.035 per cent in the free atmosphere). The dissolution of carbon dioxide in precipitation provides an additional source of bicarbonate ions. Surface waters are in fact weak carbonic acid solutions and we can write a more realistic reaction for the weathering of albite in which the release of metal cations (in this case sodium) is matched by the production of bicarbonate ions, a reaction termed carbonation:

$$\underset{\text{albite}}{4NaAlSi_3O_8} + \underset{\text{water}}{6H_2O} + \underset{\text{carbon dioxide}}{4CO_2} \rightarrow \underset{\text{silica}}{8SiO_2} + \underset{\text{kaolinite}}{Al_4Si_4O_{10}(OH)_8} + \underset{\text{dissolved sodium}}{4Na^+} + \underset{\text{bicarbonate ions}}{4HCO_3^-}$$

In general it appears that the leaching of metal cations from silicate minerals is controlled by the supply of acids; this not only involves carbonic acid, although this is the most pervasive, but also sulphuric acid and, more significantly, a range of organic acids. Carbonation plays a particularly important role in the weathering of calcareous rocks. Limestones contain a high proportion of calcium carbonate (nearly always in the form of the mineral calcite) and its weathering involves complex reversible reactions with carbon dioxide in the soil or subterranean atmosphere and carbonic acid in natural waters. In the weathering of calcite, half of the bicarbonate is derived from the calcite itself:

$$\underset{\text{calcite}}{CaCO_3} + \underset{\text{water}}{H_2O} \rightarrow \underset{\text{calcium ion}}{Ca^{2+}} + \underset{\text{bicarbonate ions}}{2HCO_3^-}$$

Oxidation and reduction

Oxidation is the process whereby an atom or ion loses an electron and thus acquires an increase in its positive charge or decrease in its negative charge. Oxygen dissolved in water is by far the most common oxidizing agent. The reaction can be reversed by reduction which involves the gaining of an electron. Oxidation acts as a weathering process in two distinct ways. Various elements, such as iron, titanium, manganese and

sulphur can be oxidized to form oxides or hydroxides. For iron this reaction is written:

$$\underset{\text{iron}}{4Fe^{2+}} + \underset{\text{oxygen}}{3O_2} \rightarrow \underset{\text{iron oxide}}{2Fe_2O_3}$$

The tendency for oxidation or reduction to occur is indicated by the redox potential (Eh) of the environment. This is measured in units of millivolts (mV), with positive values registering an oxidizing potential and negative values a reducing potential. Abundant oxygen is dissolved in most surface waters and the Eh is predominantly positive in weathering environments; that is, oxidation occurs spon concentrations taneously, though not necessarily rapidly. However, in some localities, such as below the water table and in waterlogged soils, reducing conditions can prevail. Eh also varies with pH, becoming generally lower as alkalinity increases. This relationship can be seen clearly by plotting pH and Eh values measured in various Earth surface environments (Fig. 6.9). Where a particular environment is located on such an Eh-pH diagram provides a useful indication of the nature of chemical reactions that are likely to occur.

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

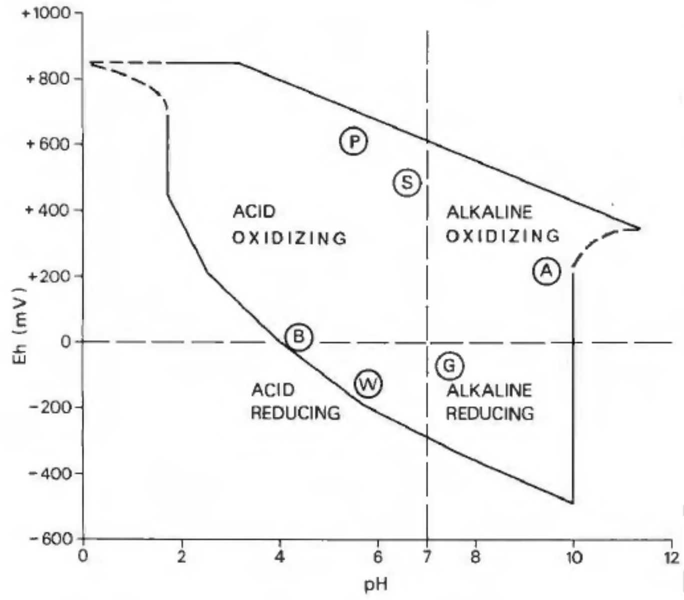


Fig. 6.9 Range of Eh–pH values encountered in the environment. The enclosed area represents the extreme limits of natural Eh–pH values as indicated by the extensive collation of data by Baas Becking et al. *(1960). The letters represent typical values for various types of environments: (P) precipitation, (S) stream water, (B) bog water, (W) waterlogged soils, (G) ground water and (A) aerated alkaline environments. (Based on R. M. Garrels and C. L. Christ (1965)* Solutions, Minerals, and Equilibria. *Freeman, Cooper, San Francisco, Fig. 11.1, p. 380 and Fig. 11.2, p. 381 after L. G. M. Baas Becking* et al. *(1960)* Journal of Geology ***68*** *Fig. 31, p. 276.)*

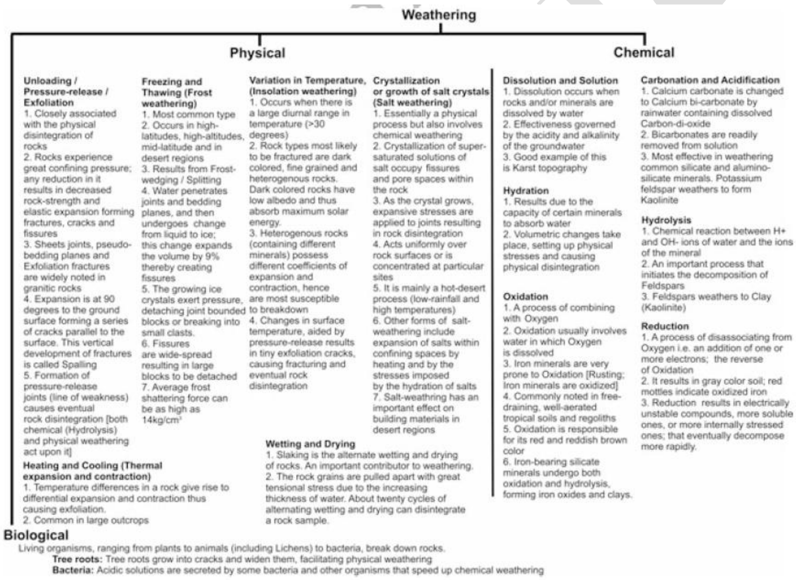
Cation exchange

Cation exchange is the substitution of one cation for another of a different element in a mineral structure. This can occur with any mineral, but it is by far the most common in clay minerals which typically have rather loosely bonded cations on their surfaces which can be readily exchanged for cations in solution.

Organic processes

We have already referred to the role of organic acids in chemical weathering, the anions of which may be important in the removal of metal cations from silicate minerals and in the dissolution of carbonates. However, it is the formation of chelating agents by organic processes that is of particular significance in weathering. These are able to mobilize metal cations, such as Fe3+ and A13+, which are virtually insoluble under normal Eh-pH conditions, a process known as chelation. Chelating agents can come from various sources, but most are organic compounds either secreted directly by organisms such as lichen, or formed through the decomposition of humus in the soil. Although the effects of chelation

are well known its detailed mechanisms are not. It appears that a stable ring structure is formed between complexing agents and metal cations. The structure may be subsequently broken down by microbial activity and the complexed cation precipitated. The dark staining characteristic of rock surfaces in arid regions, which is also observed to a limited extent in more humid environments, appears to be primarily a result of biogenic processes. Desert varnish (alternatively rock varnish), as it is termed, is composed of clay minerals, oxides and hydroxides of manganese and/or iron, together with detrital particles, which form a layer typically around 10-30 μm thick. It probably takes many thousands of years for a characteristic black to brown desert varnish to develop, apparently through the action of bacteria which preferentially oxidize and concentrate certain elements (especially manganese) supplied from dust and surface water.



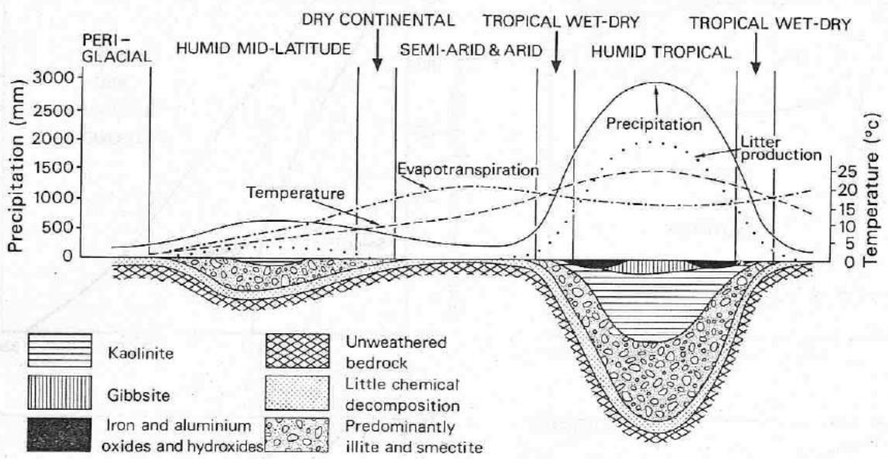


Fig. 6.18 Variation in weathering mantle depth and composition in relation to climatic and biotic variables. (Modified from N. M. Strakhov, (1967) Principles of Lithogenesis *Vol. 1. Oliver and Boyd, Edinburgh, Fig. 2, p. 6).*

Factors influencing chemical weathering

Minerals in rocks are chemically altered by this process which subsequently decompose and decay. Increasing ❶precipitation (rain) speeds up the chemical weathering process. In fact, water is an essential agent of chemical weathering. ❷Increasing temperature also accelerates chemical reactions causing minerals to degrade. Additionally, climate is another important factor affecting chemical weathering. ❸Climatic conditions control the rate of weathering that takes place by regulating the catalysts of moisture.

Physical weathering

Physical weathering encompasses a range of mechanisms, the relative effectiveness of which are not accurately known but clearly vary significantly as a function of environmental conditions. The physical breakdown of rock is always associated with some kind of volume change and it is useful to categorize the various processes into those involving an overall volumetric change in the rock mass and those related to changes in volume of material introduced into voids or fissures in the rock.

Volumetric changes of the rock mass

Rocks formed at depth or located beneath a thick overburden are under considerable internal stress. As the overlying strata are gradually removed by erosion and these rocks reach the surface they undergo expansion or dilation a process known as pressure release, which promotes the development of joints. At the smaller scale micro-fissures and incipient

joints related to the original pattern of mineral crystallization in the rock provide lines of weakness along which exfoliation (the spalling off of thin sheets of rock) and granular disintegration (the disaggregation of individual crystals or particles) can occur. In some kinds of rock, such as granite, the creation of steep, bare rock faces can lead to significant lateral expansion into the valley side as well as vertical dilation and this may give rise to exfoliation domes.

Insolation weathering, that is, the breakdown of rock as a result of volume changes arising from thermal expansion and contraction, is another potentially important physical weathering process. Fluctuations in temperature experienced by rocks at the Earth's surface due to day-time heating and night-time cooling certainly cause them to expand and contract. Nevertheless it has proved difficult to establish whether the resulting volume changes are sufficient to shatter rock, even in desert environments where diurnal temperature ranges on rock surfaces can easily exceed 30 °C. Since rock is a poor conductor of heat a thermal gradient is created when the surface is warmed. The exterior of the rock consequently expands more than the interior and stresses are set up. Data on bedrock surfaces in desert regions show that on dark rocks temperatures of 80 °C can be attained by solar radiation and diurnal ranges of 50 °C are not uncommon. Further (intergranular) stress might be expected to arise from the differential thermal expansion of individual minerals in a rock related to differences in colour, specific heat and coefficient of thermal expansion. In spite of the abundance of shattered rock in hot desert regions and elsewhere, experimental work in the 1930s cast considerable doubt on the efficacy of insolation weathering. Small rock samples subjected to repeated cycles of heating and cooling over temperature ranges well in excess of those produced naturally by solar radiation failed to shatter. The presence of moisture was found to be necessary to promote chemical weathering and weaken the rock sufficiently for disintegration to occur. Subsequently these conclusions have been challenged, it being pointed out that experiments involving the rapid heating and cooling of small individual rocks do not accurately re-create field conditions. In particular, stresses set up in such samples can be relieved in all directions whereas stresses in bedrock are much more concentrated since the rock is confined both laterally and vertically. The occurrence of scrub and forest fires provides another mechanism whereby rocks can be subjected to significant thermal expansion and contraction. It is certainly important in semi-arid regions characterized by vegetation

communities, such as chaparral (North America) and eucalyptus (Australia), which require fires for regeneration. Such areas are frequently seen to be characterized by spalling rock outcrops and shattered boulders. Chemical weathering frequently gives rise to minerals which are less dense than their precursors. The associated volume increase can exert pressure on the surrounding rock and the weathering rinds produced peel off to expose fresh rock to further weathering. The chemical processes involved are usually associated with hydration and the addition of water plays a significant role in the physical breakdown of rock. Such hydration weathering can occur simply through wetting and drying which causes expansion and contraction. The process is most active in some clay minerals, notably those of the smectite group, which are capable of absorbing large quantities of water into their crystal structures. A further hydration weathering mechanism, restricted to environments in which freezing occurs, is hydration shattering. This can occur in very fine-grained materials such as clays which are capable of retaining significant quantities of unfrozen water at temperatures well below 0 °C. At these low temperatures the dipolar molecules of this supercooled water become ordered in such a way that a repulsive force is set up across voids. It is thought that while this force is insignificant in large voids it can be sufficiently strong in the small pores in very fine-grained material to cause rock fracture. Chemical weathering is also probably associated with the mechanism which so far has only been investigated on clay':rich rocks, and it may be restricted to such lithologies.

Volumetric changes within rock voids and fissures:

Under this heading we need to consider two main sets of processes: the effects of crystal growth and expansion, primarily of ice and salt, and the stresses induced by biological activity. The latter are induced by both fauna and flora. Earthworms, for instance, ingest and excrete huge volumes of soil, as so graphically reported in a classic study by Charles Darwin in the last century. The resulting mixing of the soil, or bioturbation, probably averages about a 5 mm depth annually over the more humid parts of the continents. Other soil fauna contribute to this effect, with termites being particularly significant in moving large quantities of soil in the tropics and subtropics. The growth of tree roots and their penetration and enlargement of incipient rock fractures can

contribute significantly to the break-up of rock which may have experienced relatively little weakening through chemical alteration. Large tap roots can penetrate rock to a depth of several metres and the process commonly works in association with other mechanisms, particularly ice crystal growth. It is also likely that chemical weathering is enhanced in the vicinity of roots where carbon dioxide concentrations are likely to be higher and where the extraction of metal cations is occurring. Although such organic activity can be significant, especially locally, it is the physical processes of crystal formation and expansion that are of more fundamental importance in rock weathering.

Frost weathering

In arctic and alpine environments the surface is often seen to be composed of a layer of angular rock fragments commonly described by the term **felsenmeer** (*(German: "sea of rock"), exposed rock surfaces that have been quickly broken up by frost action so that much rock is buried under a cover of angular shattered boulders.*), and attributed to the operation of **frost weathering**. This process, which is also referred to as **frost shattering** and frost wedging, involves the breakdown of rock or other solid materials as a result of stresses induced by the freezing of water. Early explanations of frost weathering alighted on the simple and seemingly obvious effect of the 9 % volume expansion which accompanies the phase change from water to ice. Under ideal conditions it was estimated that ice formation could exert a maximum pressure of around 200 (**mega pascal**) MPa (at which point the freezing temperature is -22 °C), but it was appreciated that this theoretical maximum pressure could never be attained under natural conditions, not the least because it far exceeded the tensile strength of most rocks (around 25 MPa). Even to build up moderately high pressures sufficient to shatter rock it was realized that a closed system was required in order for the pressure not to be relieved by the expulsion of water into adjacent voids.

Salt weathering

Salt weathering involves three major processes: the ❶ precipitation of salt in voids and ❷ the expansion of salt crystals through hydration or ❸expansion of salt crystals through heating. It is most active in arid environments where rates of evaporation are high relative to precipitation, and consequently surface and soil waters can become saturated with respect to a variety of salts. Although frequently associated with hot desert regions, the effects of salt weathering have also been observed in

high-latitude arid regions such as Antarctica. Moreover, the high salinity of sea water makes it likely that salt is an active weathering agent in coastal environments. In inland areas salt is derived from a number of sources. These include surface and soil waters containing cations released during bedrock weathering which have subsequently been concentrated by evaporation, precipitation with a high salt content (especially in regions near coasts), saline ground waters and saline deposits blown inland from the coast or originating in inland saline basins.

Lithology and weathering forms

Just as variations in the physical and mineralogical properties of bedrock can influence the mineralogical products of weathering so different lithologies give rise to a range of weathering forms. On most lithologies landforms related directly to weathering tend to be minor features, but on rock types such as limestones, in which a large proportion of the products of chemical weathering processes are removed in solution, major landforms can be produced. In this section we will focus on weathering forms in limestone, although it should be pointed out that similar weathering forms can develop on other lithologies.

Karst weathering forms:

Karst is the German form of a Slovene word meaning 'bare stony ground' and is used to describe limestone terrain characterized by a lack of surface drainage, a discontinuous or thin soil cover, abundant enclosed depressions and a well-developed system of underground drainage including caves, all features attributable to the solubility of limestone. Limestone is defined as a rock which contains at least 50 % $CaCo3$, which nearly always occurs as the mineral calcite. Other carbonate minerals include aragonite (a rare form of $CaCo3$) and dolomite ($Ca\ Mg\ (Co3)z$). Rocks in which more than 50 % of the carbonate content occurs as the mineral dolomite are given the rock name dolomite. The term **karstification** is used to refer to the process of karst landscape development. The term **pseudokarst** is applied to landforms in non-carbonate rocks which are morphologically similar to those characteristic of limestone terrains. In most cases pseudokarst develops as a result of processes analogous to those operating on true karst.

Minor forms

The German term karren and the French term lapies refer to the small-scale solutional form developed on limestone. (Note that the terms 'solution' and 'dissolution' are used widely in discussions of karst landforms. Several factors influence the form of karren. Probably of most importance is the ***presence*** or ***absence*** of a cover of soil or vegetation during its formation. A second factor is ***lithology***. Karren are best developed on relatively massive and uniform, mechanically strong and impermeable limestones in which there is a sharp contact between soil and rock.

Table 6.6 Classification of solutional microforms developed on limestone

		FORM	TYPICAL DIMENSIONS	COMMENTS
Forms developed on bare limestone	*Developed through areal wetting*	**Rainpit**	<30 mm across, <20 mm deep	Produced by rain falling on bare rock. Occurs in fields on gentle rather than steep slopes. Can coalesce to give irregular, carious appearance
		Solution ripples	20–30 mm high; may extend horizontally for >100 mm	Wave-like form transverse to downward water movement under gravity. Rhythmic form implies that periodic flows or chemical reactions are important in their development
		Solution flutes (rillenkarren)	20–40 mm across, 10–20 mm deep	Develop due to channelled flow down steep slopes. Cross-sectional form ranges from semi-circular to V-shaped but is constant along flute
		Solution bevels	0.2–1 m long, 30–50 mm high	Flat, smooth elements usually found below flutes. Flow over them occurs as a thin sheet
		Solution runnels (rinnenkarren)	400–500 mm across, 300–400 mm deep, 10–20 m long	Down runnel increase in water flow leads to increase in cross-sectional area. May have meandering form. Ribs between runnels may be covered with solution flutes
	Developed through concentration of runoff	**Grikes** (kluftkarren)	500 mm across, up to several metres deep	Formed through the solutional widening of joints or, if bedding is nearly vertical, of bedding planes
		Clints (flackkarren)	Up to several metres across	Tabular blocks detached through the concentration of solution along near-surface bedding planes in horizontally bedded limestone
		Solution spikes (spitzkarren)	Up to several metres	Sharply pointed projections between grikes
Forms developed on partly covered limestone		**Solution pans**	10–500 mm deep, 0.03–3 m wide	Dish-shaped depressions usually floored by a thin layer of soil, vegetation or algal remains. CO_2 contributed to water from organic decay enhances dissolution.
		Undercut solution runnels (hohlkarren)	400–500 mm across, 300–400 mm deep 10–20 m long	Like runnels but become larger with depth. Recession at depth probably associated with accumulation of humus or soil which keeps sides at base constantly wet
		Solution notches (korrosionkehlen)	1 m high and wide, 10 m long	Produced by active solution where soil abuts against projecting rock giving rise to curved incuts
Forms developed on covered limestone		**Rounded solution runnels** (rundkarren)	400–500 mm across, 300–400 mm deep 10–20 m long	Runnels developed beneath a soil cover which become smoothed by the more active corrosion associated with acid soil waters.
		Solution pipes	1 m cross, 2–5 m deep	Usually become narrower with depth. Found on soft limestones such as chalk as well as mechanically stronger and less permeable varieties

Note: The commonly encountered German terms are given in parentheses
Source: Based largely on discussion in J. N. Jennings (1985), *Karst Geomorphology*. Blackwell, Oxford, pp. 73–82.

Major forms

If one landform can be regarded as typifying karst scenery it is the **doline**. A doline is a closed depression which may range in shape from bowl-shaped to cylindrical and in size up to 100 m deep and 1km across. In fact it is likely that doline formation may be a kind of contagious process whereby the establishment of an initial depression will tend, through its effect on lowering the water table in its immediate vicinity, to promote the Major types of doline: *(A) collapse doline; (B) solution doline; (C) subsidence doline; (D) subjacent karst collapse doline; (E) alluvial stream sink doline.*

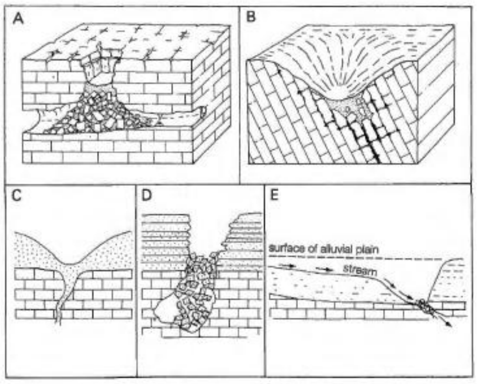


Fig. 6.23 Major types of doline: (A) collapse doline; (B) solution doline; (C) subsidence doline; (D) subjacent karst collapse doline; (E) alluvial stream sink doline. (From J. N. Jennings (1985) Karst Geomorphology. *Blackwell, Oxford, Fig. 37, p. 107).*

Other weathering forms

Exposure of bare rock are found in many environments and result from differential weathering of bedrock and the removal of the weathered debris by slope processes. Rock outcrops which stand out on all sides from surrounding slopes are known as ***tors***. They are particularly ommon on crystalline rocks, but also occur on other resistant lithologies such as quartzites and some sandstones.

Two forms which have attracted particular attention are ***honeycomb weathering*** and ***tafoni***. Honeycomb weathering consists of numerous small pits a few millimetres or centimetres in width and depth and the honeycomb form is produced when they coalesce to create a network structure. Tafoni are larger features ranging up to several metres in size and are cut into steep, bare rock faces.

Case-hardening and core softening may also be important. ***Case-hardening*** appears to result from the localized mobilization and re-precipitation of minerals on the rock surface, thereby strengthening it and rendering it generally less permeable.

Duricrusts

Duricrusts are hard layers formed in the weathering zone at, or near, the landsurface as a consequence of the absolute or relative accumulation of particular components through the replacement or cementation of pre-existing rock, soil, weathering materials or other unconsolidated deposits.

The most important components in duricrust formation are iron and aluminium oxides and hydroxides, silica, calcium carbonate and gypsum. They are of interest not only in terms of their origin but also because of their role in landscape development and the evidence they provide of past climates.

Reference:

Summerfield, M. A., 2013, Global Geomorphology. An Introduction to the Study of landforms-Routledge, London and New York, 537 pp.

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Slope Processes and Forms

Lecture THREE

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Properties and behaviour of slope materials

Slopes constitute the basic element of the landsurface and so it is not surprising that they have long formed a focal point for landform studies. At any point on the landsurface the form of the landscape is dependent upon the **nature**, **frequency** and **intensity** of the geomorphic processes acting upon it and the **strength**, or **resistance** to deformation, of the surface materials of which it is composed. The properties of these surface materials (Two fundamental types of material can be distinguished rock and soil) are clearly important in understanding the form and mode of development of the slopes which make up the landscape.

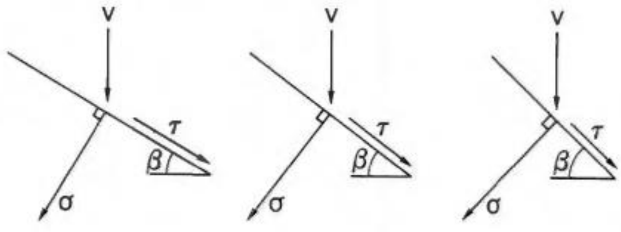
Rock is a hard, coherent material comprising individual particles or crystals. It is discontinuous in the sense that it is broken to a greater or lesser extent by joints and fractures but it is not significantly weakened when saturated with water. Soil, (or regolith), by contrast, is a weak, unconsolidated deposit which forms an essentially continuous mass lacking significant joints or fissures, but which is further weakened when saturated with water. This definition of soil includes unconsolidated weathering and sedimentary deposits as well as materials we would more normally regard as soils. Many slopes are, of course, composed of a mixture of rock and soil, but the distinction between the two is still important when analyzing the behaviour of slope materials.

Unconsolidated material transported across, and deposited on, slopes is termed ***talus*** when composed of relatively large rock fragments, and ***colluvium*** when composed predominantly of finer material. The term ***talluvium*** is sometimes used for material that is a mixture of fine and coarse material. Colluvium may contain fossil soil layers representing periods of relative slope inactivity characterized by low rates of erosion or deposition. Boulders dislodged from cliffs or free faces accumulate at the cliff foot to form a ***talus slope***, (or ***scree slope***). Where the rock fragments are funnelled down a notch or gully in an exposed rock face the material accumulates as a debris cone.

Factors determining the strength of slope materials

The response of slope materials to stress is determined by their strength which we can define as the ability to resist deformation and fracture without significant failure. The shear strength of slope materials as they occur in the field is usually rather variable over time and space because rocks and soils are generally complex mixtures of mineral particles, water and air; nevertheless, the factors determining shear strength are well known.

For soils, the more important factor is the **frictional resistance** between the constituent particles of a material. This is related to the **size of particles**, **their shape** and **arrangement**, **their resistance to crushing** and the **number of contact points per unit volume**. These inherent frictional properties collectively determine the ***angle of internal friction*** of the material. The greatest angles are achieved by materials composed of angular particles of a range of sizes which are capable of close packing.



V is the vertical stress
τ is the shear stress
σ is the normal stress
β is the angle of the shear plane

Fig. 7.1 *Change in shear stress and normal stress with change in angle of the shear plane (After M. J. Crozier (1986)*

Rock properties: Having considered the general factors determining the strength of slope materials we will now look at rock and soil more specifically. Different rock types vary enormously in their **intact strength**, that is, the strength of the rock excluding the effects of fractures and joints, as show in the table below.

Table 7.1 Intact strength of various types of rock

CHARACTERISTICS	SCHMIDT HAMMER REBOUND VALUE	INTACT STRENGTH CLASSIFICATION	EXAMPLES
Requires severe blows from hammer to break intact sample	100–60	Very strong	Quartzite, dolerite, gabbro, basalt
Hand-held sample can be broken with hammer	60–50	Strong	Marble, granite, gneiss
Shallow indentations can be made by firm hammer blow	50–40	Moderately strong	Sandstone, shale, slate
Deep indentations can be made by firm hammer blow	40–35	Weak	Coal, siltstone, schist
Crumbles under sharp hammer blow — can be cut with knife	35–10	Very weak	Chalk, rock salt

Source: Modified from M. J. Selby (1982) *Hillslope Materials and Processes*. Oxford University Press, Oxford, Table 4.3, p. 65.

<u>Mass movement</u>

Mass movement is the downslope movement of slope material under the influence of the gravitational force of the material itself and without the assistance of moving water, ice or air. The distinction between mass movement and the transport of material by other denudational processes is, however, not

always clear-cut in practice since mass movements involving material with a high water content grade into fluvial transport where streams carry very large loads of fine sediment.

Slope stability

The stability of a slope can be expressed in terms of the relationship between those stresses tending to disturb the slope material and cause it to move and those forces tending to resist these driving stresses. Clearly, movement will occur where driving forces exceed resisting forces and this relationship is represented as the safety factor for a slope. This is expressed as the ratio between shear strength and shear stress.

Slopes can exist in one of three states. Where shear strength is significantly larger than shear stress the slope is described as ***stable*** (safety factor > 1.3). Where shear stress exceeds shear strength (safety factor < 1) there will be continuous or intermittent movement and the slope is described as ***actively unstable***. Since shear strength can vary over time, especially in response to changes in the water content of slope materials, the third stability category is the ***conditionally stable slope*** which has a safety factor of 1-1.3 and fails on occasion in response to transient changes in shear strength.

Numerous factors contribute to the occurrence of mass movements, and these are listed in table below. They can be categorized as either ***preparatory factors*** or ***triggering factors***. Preparatory factors make the slope susceptible to movement without actually initiating failure by transforming it into a conditionally stable state. Triggering factors transform the slope from a conditionally stable to an actively unstable state.

Table 7.4 Factors contributing to the occurrence of mass movement

FACTOR	EXAMPLES
Factors contributing to increased shear stress	
Removal of lateral support through undercutting or slope steepening	Erosion by rivers and glaciers, wave action, faulting, previous rock falls or slides
Removal of underlying support	Undercutting by rivers and waves, subsurface solution, loss of strength by extrusion of underlying sediments
Loading of slope	Weight of water, vegetation, accumulation of debris
Lateral pressure	Water in cracks, freezing in cracks, swelling (especially through hydration of clays), pressure release
Transient stresses	Earthquakes, movement of trees in wind
Factors contributing to reduced shear strength	
Weathering effects	Disintegration of granular rocks, hydration of clay minerals, dissolution of cementing minerals in rock or soil
Changes in pore-water pressure	Saturation, softening of material
Changes of structure	Creation of fissures in shales and clays, remoulding of sand and sensitive clays
Organic effects	Burrowing of animals, decay of tree roots

Mass movement processes

There have been numerous attempts to classify the diverse modes of mass movement, none of them universally satisfactory. Here we identify six fundamental types of movement - creep, flow, slide, heave, fall and subsidence. Each of these can be subdivided into more specific forms of mass movement (Table below).

Table 7.5 Classification and characteristics of the major types of mass movement

PRIMARY MECHANISM			MASS MOVEMENT TYPE	MATERIALS IN MOTION	MOISTURE CONTENT	TYPE OF STRAIN AND NATURE OF MOVEMENT	RATE OF MOVEMENT
LATERAL COMPONENT PREDOMINANT	Creep		Rock creep	Rock (especially readily deformable types such as shales and clays)	Low	Slow plastic deformation of rock, or soil producing a variety of forms including cambering, valley bulging and outcrop bedding curvature	Very slow to extremely slow
			Continuous creep	Soil	Low		
	Flow		Dry flow	Sand or silt	Very low	Funnelled flow down steep slopes of non-cohesive sediments	Rapid to extremely rapid
			Solifluction	Soil	High	Widespread flow of saturated soil over low to moderate angle slopes	Very slow to extremely slow
			Gelifluction	Soil	High	Widespread flow of seasonally saturated soil over permanently frozen subsoil	Very slow to extremely slow
			Mud flow	>80% clay-sized	Extremely high	Confined elongated flow	Slow
			Slow earthflow	>80% sand-sized	Low	Confined elongated flow	Slow
			Rapid earthflow	Soil containing sensitive clays	Very high	Rapid collapse and lateral spreading of soil following disturbance, often by an initial slide	Very rapid
			Debris flow	Mixture of fine and coarse debris (20–80% of particles coarser than sand-sized)	High	Flow usually focused into pre-existing drainage lines	Very rapid
			Debris (rock) avalanche (sturzstrom)	Rock debris, in some cases with ice and snow	Low	Catastrophic low friction movement of up to several kilometres, usually precipitated by a major rock fall and capable of overriding significant topographic features	Extremely rapid
			Snow avalanche	Snow and ice, in some cases with rock debris	Low	Catastrophic low friction movement precipitated by fall or slide	Extremely rapid
			Slush avalanche	Water-saturated snow	Extremely high	Flow along existing drainage lines	Very rapid
	Slide	*Translational*	Rock slide	Unfractured rock mass	Low	Shallow slide approximately parallel to ground surface of coherent rock mass along single fracture	Very slow to extremely rapid
			Rock block slide	Fractured rock	Low	Slide approximately parallel to ground surface of fractured rock	Moderate
			Debris/earth slide	Rock debris or soil	Low to moderate	Shallow slide of deformed masses of soil	Very slow to rapid
			Debris/earth block slide	Rock debris or soil	Low to moderate	Shallow slide of largely undeformed masses of soil	Slow
		Rotational	Rock slump	Rock	Low	Rotational movement along concave failure plane	Extremely slow to moderate
			Debris/earth slump	Rock debris or soil	Moderate	Rotational movement along concave failure plane	Slow
	Heave		Soil creep	Soil	Low	Widespread incremental downslope movement of soil or rock particles	Extremely slow
			Talus creep	Rock debris	Low		
VERTICAL COMPONENT PREDOMINANT	Fall		Rock fall	Detached rock joint blocks	Low	Fall of individual blocks from vertical faces	Extremely rapid
			Debris/earth fall (topple)	Detached cohesive units of soil	Low	Toppling of cohesive units of soil from near-vertical faces such as river banks	Very rapid
	Subsidence		Cavity collapse	Rock or soil	Low	Collapse of rock or soil into underground cavities such as limestone caves or lava tubes	Very rapid
			Settlement	Soil	Low	Lowering of surface due to ground compaction usually resulting from withdrawal of ground water	Slow

Source: Based largely on D. J. Varnes (1978) in: R. L. Schuster and R. J. Krizek (eds) *Landslide Analysis and Control*, Transportation Research Board Special Report 176. National Academy of Sciences, Washington, DC, 11–33.

Gravity tectonics

Gravity tectonics is a useful term which covers a range of processes extending from the very large-scale movements of rock masses involved in the development of thrusts and nappes to smaller scale downdip translocations of material which are transitional to landslides. Many of the massive nappes occurring in intercontinental collision orogens such as the Alps are now considered to be vast gravity slides moving away from their high, axial zones.

Water erosion and solute transport on slopes

In addition to being affected by mass movement processes, slope material can be transported by water. Three mechanisms are involved: **rainsplash** erosion, in which particles on the soil surface are dislodged by the impact of raindrops; **slope wash**, which is the process whereby sediment is entrained and transported by a thin sheet of water flowing over the slope surface; and **solute transport** where soil materials taken into solution during weathering reactions are transported downslope as solutes. These mechanisms are sometimes collectively referred to ***wash processes***. Rainsplash and slope wash are effective only on soil covered slopes, whereas solution can operate on both soil and rock slopes.

Rainsplash erosion

Raindrops possess kinetic energy by virtue of their mass and velocity. Although the impact velocity of raindrops varies depending on *droplet size*, *wind speed* and *turbulence*, raindrops of the maximum size under normal conditions of around 6 mm diameter have an impact velocity of about 9 m/sec. At this speed rain drops can directly move particles more than 10 mm across and coarser material can be dislodged by the removal of downslope support provided by finer sediment. Rainsplash erosion can occur wherever vegetation does not entirely cover the ground, although it is a more potent erosive agent in environments where there is little or no vegetation cover. Both ***slope gradient*** and ***surface characteristics*** influence the effectiveness of rainsplash erosion.

Slope wash

The movement of water across a slope surface, irrespective of how it is generated, is termed sheet flow, although this is a rather misleading description since the water flow is never of uniform depth because of the microtopography of hillslope surfaces. Sheet flow can in fact grade into channelled fluvial flow as the water movement becomes progressively more concentrated into

particular downslope routes, and the distinction between the two is sometimes difficult to make.

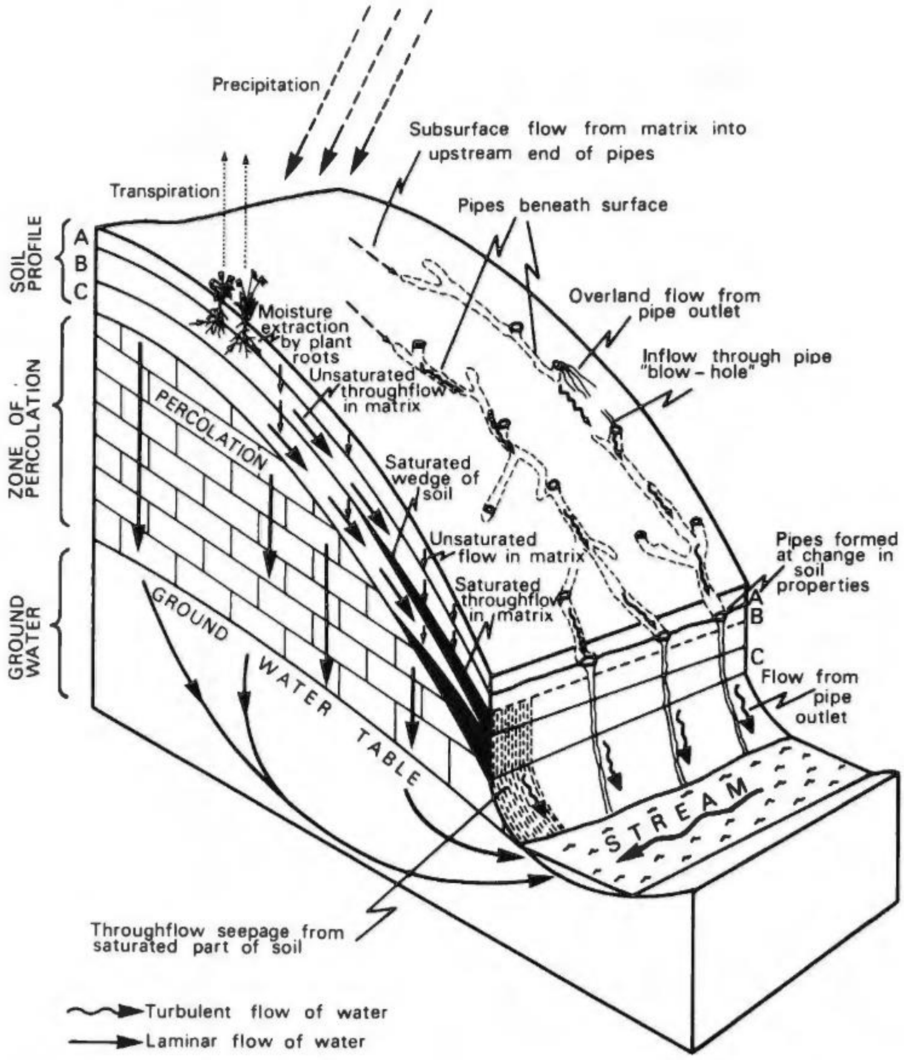


Fig. 7.14 Routes of subsurface flow on hillslopes. (From T. C. Atkinson (1978) in M. J. Kirkby (ed.) Hillslope Hydrology, *Wiley, Chichester, Fig. 3.1, p. 74.)*

Sustained concentrated flow can eventually produce **rills**, micro-channels a few centimetres in depth and width. In humid environments the presence of vegetation means that rills usually develop only on artificially disturbed surfaces, but in arid and semi-arid environments they can occur naturally. Although rills may be destroyed between rainfall events by other slope processes, especially soil creep, those favourably located may eventually be enlarged into **gullies** and form a permanent part of a channel network. Contour curvature has a significant effect on slope wash since where contours are convex in plan sheet flow will be dispersed downslope and erosion will be minimized. Conversely, at valley heads and other locations where contours are concave in plan, the flow will be concentrated downslope and rill erosion will consequently be more effective.

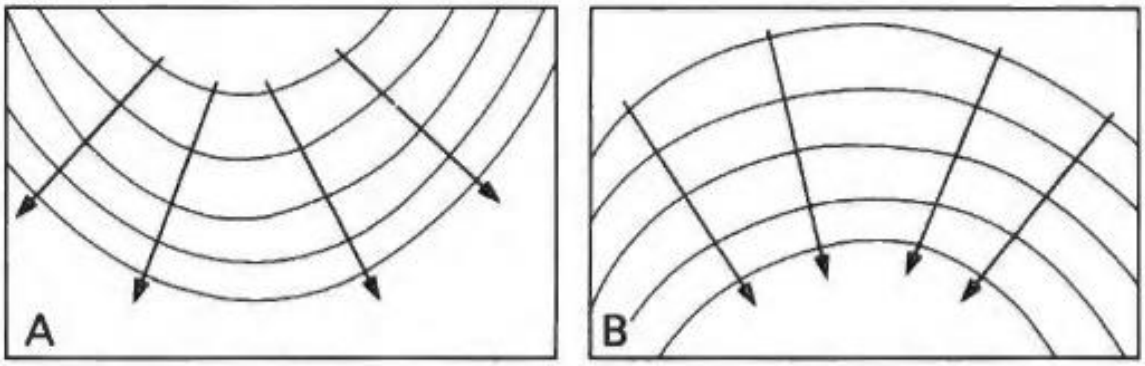


Fig. 7.16 Effect of contour curvature on sheet flow and rill erosion: (A) contours convex in plan with sheet flow dispersed; (B) contours concave in plan with sheet flow concentrated.

Soil erosion

The rate at which material is eroded from a slope is a function of both ***erodibility***, or the resistance of slope materials to entrainment and transport, and ***erosivity***, the potential of slope processes to cause erosion. A plenty of factors influence erodibility and erosivity, and it is extremely difficult to quantify these in order to predict the rate of erosion on a particular slope under a given set of conditions.

Solute transport

The transport of slope materials in solution has been the least studied aspect of slope denudation, and there are comparatively few quantitative data by which direct comparisons can be made with the effectiveness of other slope processes. The loss of weathered material in solution leads to the rearrangement and settling of the remaining particles, although it is difficult to calculate the effect of such loss on the form of the slope itself. The rate of solute transport can certainly be estimated by measuring the discharge of subsurface flow on a slope and relating this to its solute concentration.

The slope system

The ability of processes on hillslopes to fashion slope form is determined by the capacity of these processes to transport the available slope material. There are two kinds of situation for slope development. In one, erosion is limited by the rate at which material is made available through weathering. In the other, there is no effective limit to the availability of weathered material and slope erosion is therefore controlled by the capacity of the transport processes. This

distinction has subsequently developed into the idea that slopes can be viewed either as weathering-limited or as transport-limited.

Slope form

Slope form is most often represented in terms of two dimensional slope profiles. Slope profiles run from drainage lines up the steepest slope to drainage divides. They are usually measured using simple surveying equipment, such as an Abney level or clinometer, which provides slope angle measurements over a known distance on the ground. Slope profiles can be subdivided into individual components or units consisting of convex and concave elements and straight, or rectilinear, slope segments. Slope forms vary enormously, but in many cases they comprise an ***upslope convexity*** leading down to a rectilinear main slope which terminates in a basal concavity. The main slope can consist of either a single segment or a more complex sequence of segments at different angles.

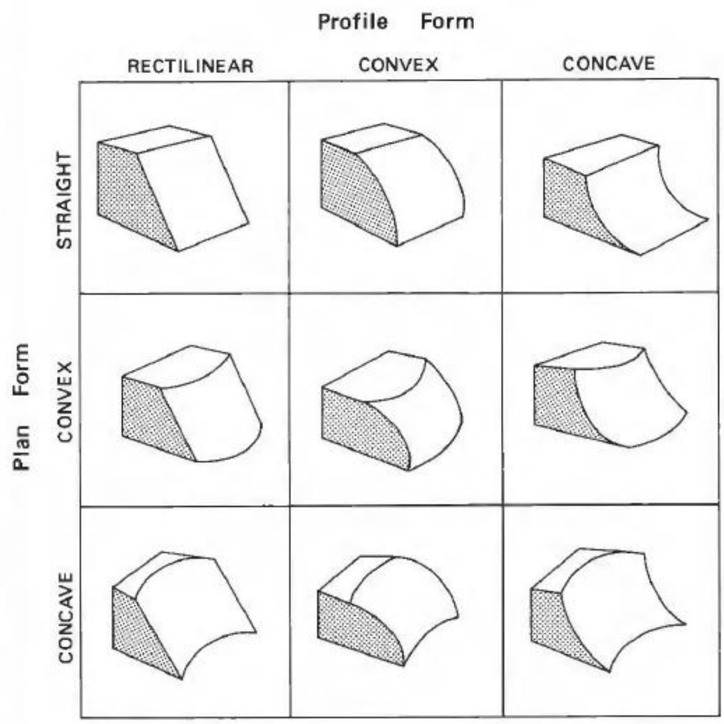


Fig. 7.20 *The nine possible shapes of three-dimensional hillslope forms. (Modified from A. J. Parsons (1988)* Hillslope Form. *Routledge, London, Fig. 2.5, p. 16.)*

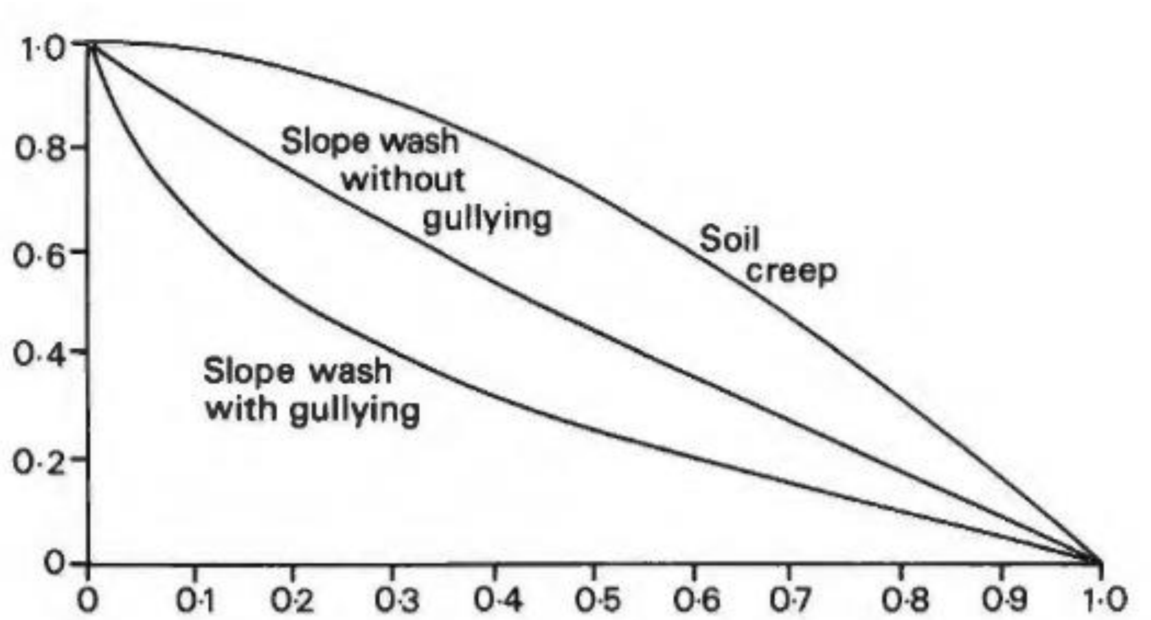


Fig. 7.24 *Characteristic slope forms for different slope processes. (Modified from M. J. Kirkby (1971)* Institute of British Geographers Special Publication **3**, *Fig. 5, p. 26.)*

Slope evolution

The manner in which slope form changes through time provided a central focus for geomorphic research until the 1950s. Various models of slope evolution were proposed, notably by W. M. Davis, W. Penck and L. C. King, which formed a framework within which the development of the landscape as a whole was considered. Davis presented a model of slope decline in which there is a progressive decrease in overall slope angles through time as the rate of basal down cutting by streams decreases and slopes become mantled with weathered material of ever finer calibre which can be transported across ever lower gradient slopes.

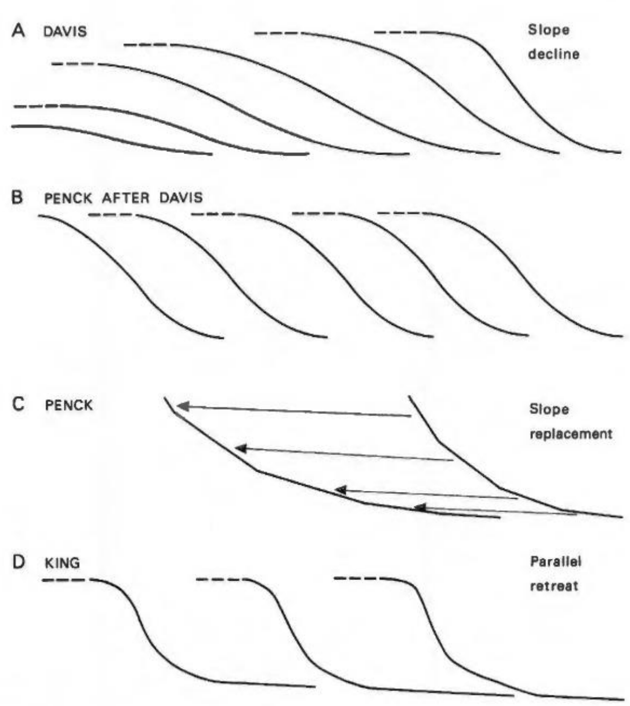


Fig. 7.25 Classic models of slope development through time: (A) Davis's model of slope decline; (B) Davis's misinterpretation of Penck's model indicating parallel retreat; (C) Penck's model of slope replacement; (D) King's model of parallel retreat.

Penck's model was originally presented in German, and it was introduced to the English-speaking world by Davis who seriously misrepresented Penck's ideas. Rather than advocating parallel retreat of the major part of a slope, as implied by Davis (Fig. 7.25(B», Penck in fact considered that where rates of denudation are declining hillslopes evolve through a process of flattening from the base upwards. In effect each part of the slope profile is replaced by a slope of lower gradient as it retreats, and through this process of slope replacement a broadly concave profile is produced. In rejecting Davis's statement that slope gradients decline through time, King pointed to the widespread occurrence of escarpments, notably in southern Africa, which have apparently experienced prolonged parallel retreat. King argued that the free face retreats parallel to itself as material is weathered and then removed by rill erosion, and a low angle slope, or pediment, grows at its base. Such slope retreat eventually creates isolated, residual hills, known as inselbergs. Although they have been extremely important in influencing previous views about landscape development none of these classic models is based on detailed empirical observations. Moreover, they do not yield specific quantitative predictions as to how slope form may be expected to change over time. Studies of slope processes since the 1960s have clearly demonstrated that the evolution of slopes is, in most cases, likely to be far more complex than implied in these classic models, and that the mode of development will depend on structural and lithogical properties as well as the slope processes operating.

A rock slope will experience parallel retreat if the strength of the rock mass remains constant and basal debris is continuously removed, but lithological variations and climatic changes mean that over an extended period of time there will inevitably be some change in slope profile form. Parallel retreat is also likely to predominate in situations where a flat-lying resistant lithology overlies less resistant strata. Such situations are common where weakly resistant regolith or other types of unconsolidated deposits are overlain by duricrusts, or where igneous intrusions, such as dolerite sills, cap less resistant sedimentary strata. Once the resistant cap rock is finally removed by back-wearing, subsequent slope evolution may occur through slope decline. Where rates of basal undercutting decline we might expect a reduction in slope gradient as material can be more intensively weathered and the angle of threshold slopes is reduced. This is suggested by the presence of distinct groupings of threshold slopes, the number and angle of which apparently depends on the weathering sequence of the bedrock. We can envisage slope development in a well-jointed rock in which there is an initial phase of scree formation, a second stage marked by the production of taluvium stable at a lower angle, and finally the development of a cover of colluvium with the lowest angle of threshold stability. Whether the associated changes in slope gradient occur by an overall decline in slope angle or by a replacement of individual slope elements by segments standing at a lower angle is probably dependent on the nature of the predominant slope processes and the way these interact with the slope materials. These factors are summarized in **Figures in the textbook** which illustrate the probable course of profile development on different lithologies and under contrasting climatic conditions. The profiles show the likely change in form for slopes initially subject to active basal downcutting, but which subsequently develop in the absence of active basal erosion.

Reference:

Summerfield, M. A., 2013, Global Geomorphology. An Introduction to the Study of landforms-Routledge, London and New York, 537 pp.

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Fluvial Processes

Lecture FOUR

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Drainage basin hydrology

In Lecture THREE we looked briefly at the movement of water on and within slopes, but it is now necessary to extend our discussion of water movements to the scale of entire drainage basins. All water in rivers ultimately originates as precipitation, although there can be a considerable lag before this water enters the fluvial system. The main components of drainage basin hydrology are illustrated in Figure below.

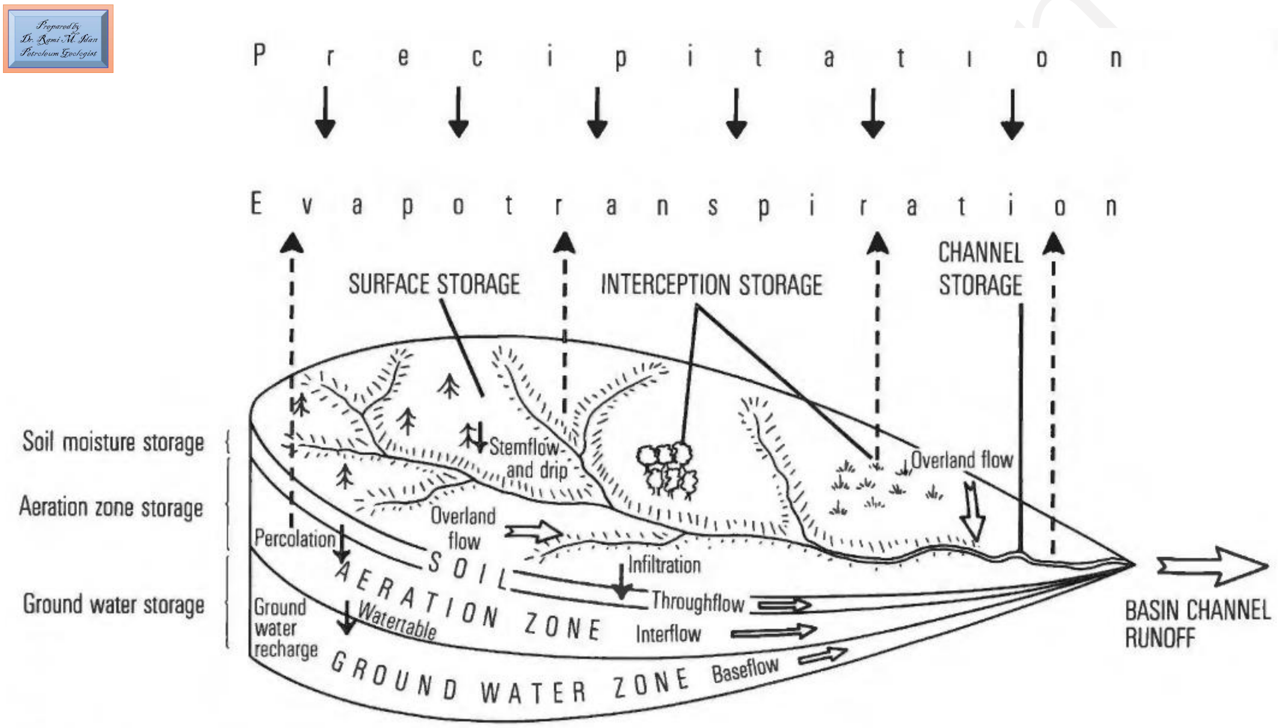


Fig. 8.1 Primary storages and transfers of water within drainage basins.

Runoff, or more strictly **basin channel runoff,** is the quantity of water which enters stream channels in a drainage basin over a specified period of time, and can be determined by a **water-balance equation.** This expresses runoff in terms of precipitation, losses through evapotranspiration and changes in the amount of soil moisture and ground water storage. In environments where much of the precipitation falls as snow the water-balance equation is complicated by having to take into account water released during melting. A further output which has to be considered, in addition to runoff and evapotranspiration, is **deep outflow** from ground water. In most cases this happens so slowly that it can be ignored in calculating the water balance for a basin. In limestone terrains, however, movements of water at depth may have little relation to surface patterns of water flow and a complex hydrological system can result.

Channel initiation

An obvious question to ask about stream channels is how do they originate? Channels may be created on a newly exposed surface or develop through the expansion of an existing channel network, but in order to understand how they are initiated we must look at the conditions under which water flowing on a slope becomes sufficiently concentrated for channel incision to occur. It is also necessary to establish how, once they are established, channels are maintained and enlarged to form 'permanent' features in the landscape. In Lecture THREE we saw how both surface and subsurface flows converge in areas of contour concavity, and such convergence is an important factor in channel development. Moreover, it was also pointed out how infiltration-excess overland flow can lead to the development of rills, although the precise mechanism that brings this about is far from clear. Microtopography on slopes tends to disrupt sheet flow and promote the concentration of water movement and rill formation. But such rill development can be counteracted by the lateral shifting of flow lines, or by rainsplash erosion which tends to even out the surface.

In the model put forward by R. E. Horton, before erosion by overland flow can occur on a hillslope it must reach a critical depth at which the eroding stress of the flow exceeds the shear resistance of the soil surface (Fig. below). Horton therefore thought that a 'belt of no erosion' is present on the upper part of slopes because here the flow depth is insufficient to cause erosion. Subsequent work has shown that some surface wash is possible even on slope crests, although here it does not lead to rill development because the rate of incision is slow and incipient rills are infilled as a result of rainsplash.

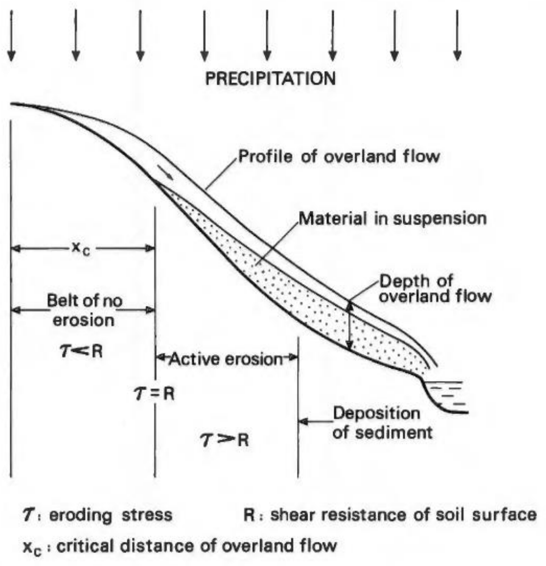
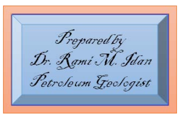


Fig. 8.5 Horton's model of surface erosion by infiltration-excess overland flow. (After R. E. Horton (1945) Bulletin of the Geological Society of America *56 Fig. 14, p. 316.)*



Open channel flow

Two opposing forces act on water flowing in an open channel. The ***driving force*** is gravity which acts in a downslope direction and is determined by gravitational acceleration and channel gradient. The ***resisting force*** arises from friction both within the water body and between the flowing water and the channel surface. The ability of flowing water to entrain and transport material, and hence its capacity to do geomorphic work, is essentially determined by the relationship between these two forces. However, before looking at the processes operating in Natural River channels we need to consider the fundamental characteristics of the behaviour of flowing water.

Resistance to flow

Water is a fluid - that is, its shape is changed continuously by the smallest applied external stress. This change in shape is sustained for as long as the force is applied. Resistance of a fluid to a change in shape is represented by its ***viscosity***. One type of resistance to deformation provided by viscosity arises from internal friction caused by cohesion and collisions between molecules as they move past each other, and consequently this type of viscosity is termed ***molecular viscosity*** or ***dynamic viscosity***. Around 97% of the energy of rivers is expended as frictional heat generated by molecular impacts, leaving only about 3% for the transport of sediment. As with all liquids the molecular viscosity of water increases with a decrease in temperature. Moreover, in the natural environment we are concerned not only with pure water but also with water containing fine sediment or dissolved constituents which has a greater dynamic viscosity.

Laminar and turbulent flow:

A fluid moving over a flat solid surface can act as a series of thin 'layers' sliding over one another, the resistance to movement resulting from molecular viscosity. This form of motion is described as laminar flow and while it is relatively common in highly viscous fluids, such as lava flows, it is extremely rare for water moving in natural channels. In stream channels water movement nearly always occurs as turbulent flow, that is, the velocity of flow fluctuates in all directions within the fluid. Water is constantly interchanged in eddies or swirls between adjacent zones of flow, and local changes in velocity occur which work against the mean velocity gradient and lead to a loss of energy. The resulting additional resistance to shear is termed ***eddy viscosity***. In most channels there is a thin laminar sub-layer within which the velocity of flow

initially increases in a roughly linear fashion. Above this zone the rate of increase in flow velocity is approximately logarithmic (Fig below).

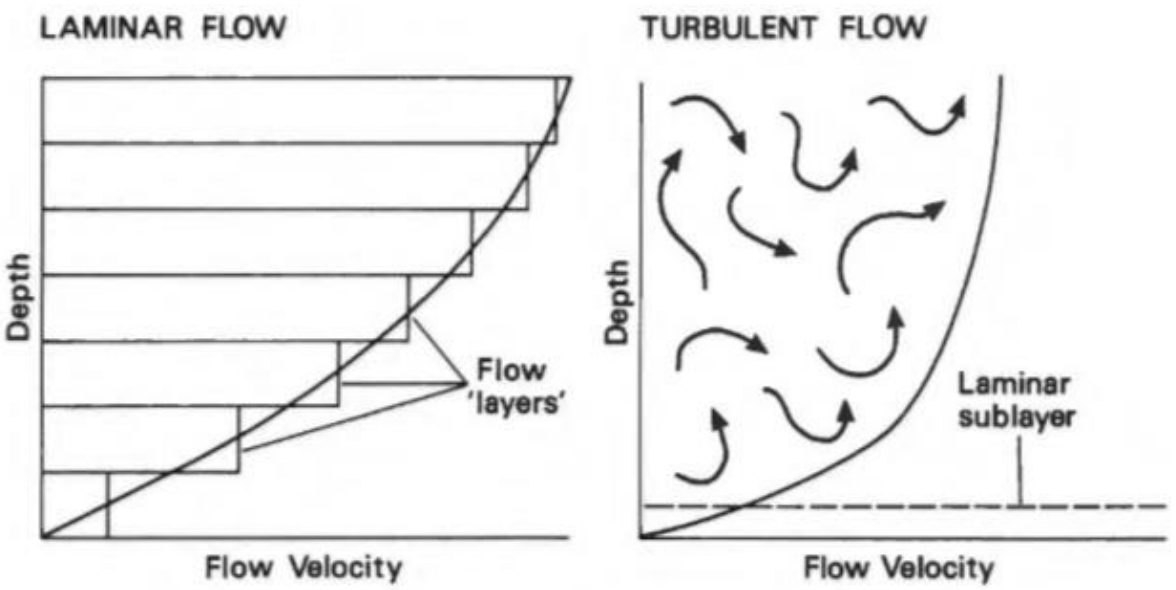


Fig. 8.6 Schematic representation of velocity profiles for laminar and turbulent flow in a river channel.

Fluvial erosion and sediment entrainment

Erosion of bedrock channels

The reduction of the landscape through the action of fluvial processes can involve the incision of stream channels into bedrock as well as the entrainment and downstream transportation of sediment. The erosion of bedrock channels must be of considerable significance in mountainous regions, and although the mechanisms involved are poorly understood three major processes appear to operate.

Corrosion is the chemical weathering of minerals in contact with stream water and the removal of soluble products downstream. The key factors controlling rates of corrosion are bedrock mineralogy, the solute concentration of the stream water, the stream discharge and velocity of flow. Maximum rates of corrosion are achieved where fast-flowing, under-saturated stream waters pass over lithologies with a high proportion of reactive minerals; for instance, corrosion is an important process in bedrock channels in mountainous limestone terrains in humid environments.

Abrasion, or **Corrasion**, and consists of the wearing away or detachment of bedrock by particles moved by the water flow. The particles involved can be of any size that can be transported at prevailing flow velocities, and large boulders several metres across may be in motion in fast-flowing, deep river channels.

The effectiveness of abrasion depends on the concentration, hardness and kinetic energy of the impacting particles and the resistance of the bedrock surface. Since kinetic energy is proportional to the square of velocity, rates of abrasion increase rapidly as flow velocities increase.

Hydraulic action that is the movement of water alone. One way this can occur is through the detachment of loose rock fragments by the force or moving water.

Sediment entrainment

The majority of rivers do not cut directly into bedrock but flow in alluvial channels formed in unconsolidated sediments. These sediments may range in calibre from boulders to clay-sized material. Alluvial channels are 'self-formed' equilibrium or quasi-equilibrium landforms in that their morphology arises from the mobilization, transportation and deposition of sediment and represents an adjustment to prevailing hydrological and sedimentological conditions.

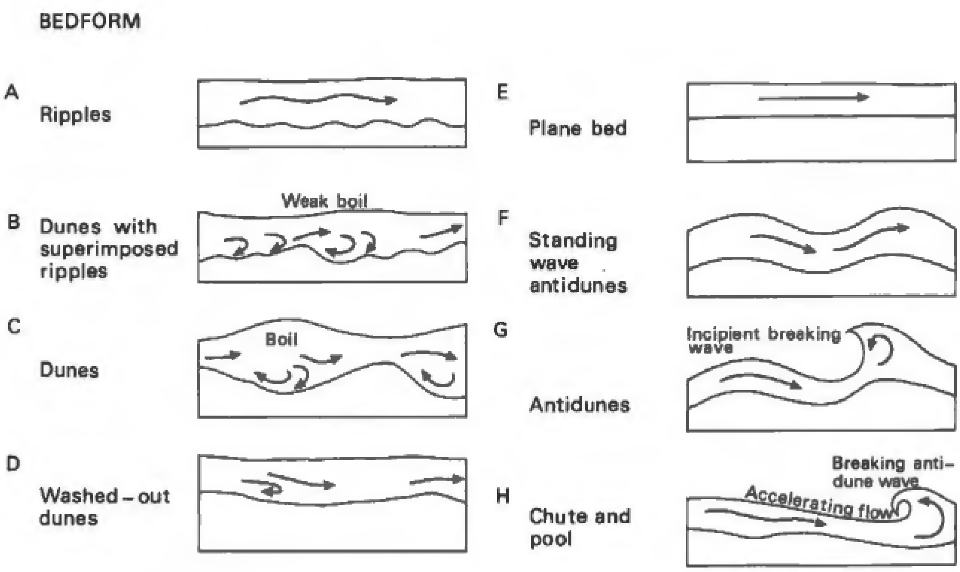


Fig. 8.10 *Bedforms in sandy alluvial channels in relation to flow regimes expressed by Froude numbers. At low flow velocities ripples are formed (A), but as the flow velocity increases ripples are transformed into larger forms called* ***dunes*** *(B and C), both being out of phase with waves on the water surface. With a further increase in velocity bed undulations are planed off, resistance to flow is lowered and sediment transport rates increase (D and E). This is a transitional state between subcritical and supercritical flow. With a further increase in velocity, supercritical flow gives rise to* ***antidunes*** *which because they are in phase with standing waves at the water surface present a low resistance to flow (F and G). Antidunes move upstream since sediment is lost from their downstream side more rapidly than it is deposited. At the highest flow velocities fast – flowing shallow chutes alternate with deeper pools (H). (Based on D. B. Simons and E. V. Richardson (1963)* Transactions of the American Society of Civil Engineers, ***128****, Fig. 2, p. 289.)*

Alluvial channels can generally adjust rapidly to changes in the balance between the stresses generated by the flowing water and the resistance of the channel bed sediments to movement. In this respect they differ significantly from bedrock channels which can usually change only slowly and whose morphology is dominated by structural and lithological controls. Sediment in alluvial channels includes particles previously carried downstream as well as material contributed directly from valley-side slopes and from bank erosion.

The initial setting into motion of a solid particle in a fluid is so called **entrainment** and this occurs when the stresses acting on a particle exceed the resisting forces.

Unless the flow is highly turbulent and energetic larger particles move on a trajectory converging with the channel bed at a low angle (Fig. below). This kind of particle motion is termed ***saltation***. When large numbers of grains are in motion under rapid flow conditions the ideal saltation trajectory is not attained because there are frequent collisions between particles. In this situation there is a concentrated dispersion of particles near the channel bed dominated by interparticle collisions and deflections. Larger particles which cannot be lifted from the channel bed may simply move across it by either ***rolling*** or ***sliding***. Impacting saltating grain returning to the channel bed may help to precipitate this movement.

In addition to rolling, sliding and saltating, solid particles may experience ***suspended motion*** in which their trajectories are more irregular and more prolonged than for saltation. The weight of fine particles in true suspension is entirely supported by the upward pulses of flow generated by eddies. Grain descending during saltation may be temporarily buoyed up by upward movements in turbulent flows and this condition is more appropriately described as ***incipient suspension***.

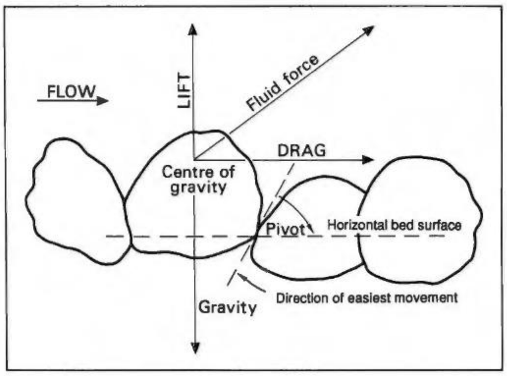


Fig. 8.11 Schematic representation of the forces acting on a grain resting with others of similar size and shape on a channel bed and subject to fluid flow. (After G. V. Middleton and J. B. Southard (1984) Mechanics of Sediment Movement. *Society of Economic Paleontologists and Mineralogists Short Course No. 3 Fig. 6.1, p. 6–3.)*

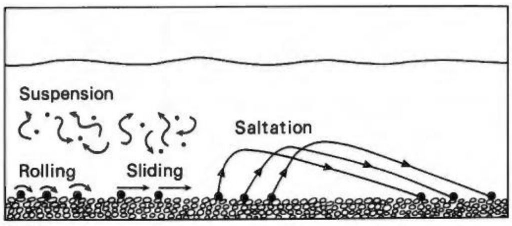


Fig. 8.12 Schematic representation of the modes of sediment transport in flowing water.

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Fluvial transport and deposition

Modes of fluvial transport

Material can be transported by rivers either as solid particles or in solution. ***Solute or dissolved load***, which is derived largely from bedrock weathering, is dispersed throughout the flow. Solid load is of two main types. ***Bed load***, or ***traction load***, encompasses all material rolling, sliding or saltating along the channel bed. Suspended load is invariably of fine calibre and includes all particles prevented from falling to the channel bed by the upward momentum imparted by eddies within turbulent flows. The finest fraction of suspended load, consisting of very small clay-sized particles, is termed ***wash load*** and is able to stay essentially in permanent suspension as long as some flow is maintained.

Fluvial deposition

Just as there are ***threshold flow velocities*** *(This is defined as the state in a specific channel reach when the power available is exactly sufficient to transport the mean available sediment load)* for the entrainment of particles of different sizes so there are thresholds for sediment deposition. The velocity at which a particle settles to a channel bed, known as its ***fall velocity***, is a function of both its density, size and shape and of the viscosity and density of the transporting fluid. Since viscosity and density change with the concentration of sediment in a stream flow, deposition is not related simply to flow velocity. As flow velocity decreases the coarser sediment begins to be deposited while the finer particles remain in motion, and this differential settling of the material in transit gives rise to sediment sorting. Conditions on a channel bed can change rapidly over time and space, so whether sediment is entrained or deposited depends on local rather than average conditions.

Indeed it is possible for deposition and entrainment to be occurring simultaneously in the same channel reach and coarse sediment can be deposited at the same time that finer particles are being entrained. Nevertheless, as discharge fluctuates between high and low stages, episodes of degradation during which erosion predominates will alternate with periods of aggradation during which deposition prevails.

Reference:

Summerfield, M. A., 2013, Global Geomorphology. An Introduction to the Study of landforms-Routledge, London and New York, 537 pp.

Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Fluvial Landforms

(Part ONE)

Lecture FIVE

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

The fluvial system

Rivers and the landforms that they create, can be considered at an enormous range of scales. In this lecture we are primarily concerned with the way fluvial processes create fluvial landforms. These range from the processes operating in a single bend in a river, to the different channel patterns arising from contrasting conditions of water flow, sediment transport and channel gradient, and ultimately to the morphology of entire drainage basins. Rivers and streams can be simply defined as bodies of water flowing in an open channel. They have three important roles in landscape creation: ❶they erode the channels in which they flow, ❷they transport sediments and solutes provided by weathering and slope processes as well as by the other denudational agents of ice and wind, and ❸they produce a wide range of erosional and depositional landforms. Fluvial systems can for convenience be regarded as consisting of three main elements: a zone of sediment production, a zone of sediment transfer and a zone of sediment deposition. This categorization is, of course, oversimplified because some erosion, transport and deposition occurs in all three zones; nevertheless, within each zone one of these three processes is usually dominant. In large basins the upstream zone in which sediment production fluvial landforms predominates is usually a mountainous or upland region. The zone dominated by deposition is generally located along a coast and takes the form of a delta or lowland coastal plain.

The drainage basin

A drainage basin is an area within which water supplied by precipitation is transferred to the ocean, a focus of internal drainage, such as a lake, or to a larger stream. For a number of reasons the drainage basin is the fundamental unit of fluvial geomorphology. Drainage basins are usually well-defined areas, clearly separated from each other by drainage **divides**, within which surface or near surface flows of water and associated movements of sediment and solutes are contained.

Channel network characteristics

River systems are a type of **network**; that is, they consist of a series of links which connect nodes. Networks can be analyzed with respect to two main sets of properties: ❶the topological aspects of stream networks concern the interconnections of the system, whereas the ❷geometrical aspects involve length, area, shape, relief and orientation properties. The basic element of stream networks is the **stream segment**, or **link**. *This is a section of stream channel between two channel junctions or, for 'fingertip' tributaries, between a junction and the upstream termination of a channel.* Stream order expresses the hierarchical relationship between stream segments. (Fig. below).

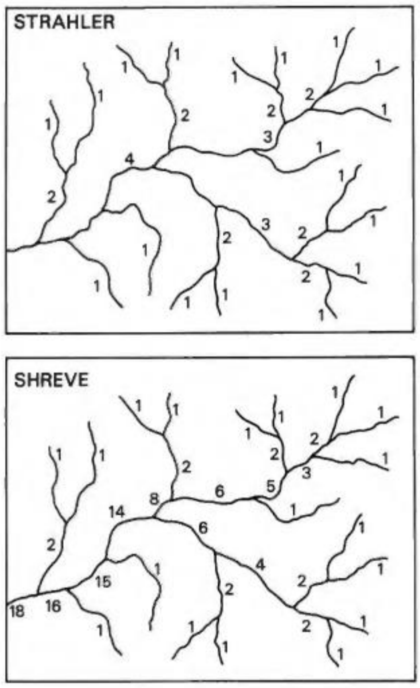


Fig. 9.1 Schemes of stream ordering proposed by A. N. Strahler (stream order) and R. L. Shreve (stream magnitude) (see text for explanation).

In the Strahler system a stream segment with no tributaries is designated a first order segment. A second order segment is formed by the joining of two first order segments, a third order segment by the joining of two second order segments and so on. It is important to note that with the Strahler ordering method there is no increase in order when a segment of one order is joined by another of a lower order. Stream order as defined by Strahler has been applied to numerous river systems and has been shown to be statistically related to various elements of drainage basin morphometry.

Probably the most important of these is **drainage density**. This reflects a balance between erosive forces and the resistance of the ground surface, and, as a consequence, is closely related to climate and lithology.

Areal and relief characteristics

Areal properties express the overall plan form and dimensions of drainage basins, while relief properties express elevation differences.

The relief, or height differences, in a basin can be expressed quite simply using maximum and minimum elevation values.

Local relief is the difference between maximum and minimum elevations within a given area. Relief is related to the slope and stream gradients in a basin, and so indirectly has an influence on the rates of slope processes and sediment transport by rivers.

There is a close relationship between drainage density, mean slope angle and relief (Fig. below). If drainage density is constant and stream channels maintain a constant spacing through time, an increase in local relief due to stream incision must, of necessity cause an increase in mean slope angles in the basin.

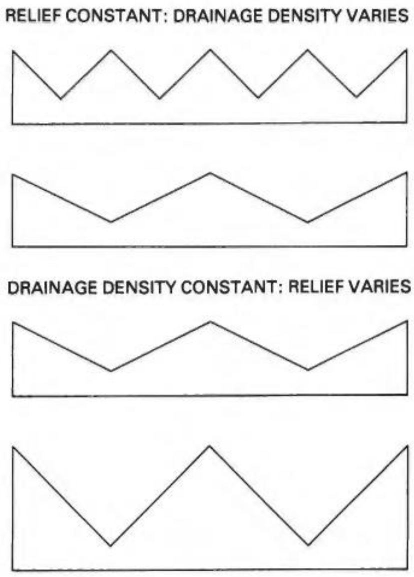


Fig. 9.4 Relationships between relief, drainage density and mean slope.

River channels

Three major types of river channel can be identified. ❶***Bedrock channels*** are cut into rock. In general they experience gradual modification but retain their overall form for long periods of time. Significant lateral shifting of channels may, however, occur where the bedrock is only weakly resistant. In ❷ ***alluvial channels*** the bed and banks are composed of sediment being transported by the river. They can undergo dramatic changes in form as weakly resistant alluvium is eroded, transported and redeposited in response to changes in water discharge and sediment load, among other factors. Semi-controlled channels are of intermediate type, being only locally controlled by bedrock or resistant alluvium. ❸A ***semi-controlled channel*** will be stable where it is cut into bedrock or resistant alluvium, but over time it may migrate laterally into alluvium and be much more responsive to changes in hydrological and sedimentological variables.

Alluvial channels: plan form

Alluvial channels exhibit a great variety of plan form. Numerous channel patterns can be recognized but they all represent variations of just a few basic types. One key property is ***sinuosity*** which represents the irregularity of the channel course and is expressed as the ratio of channel length (measured along the centre of the channel) to valley length (measured along the valley axis). The ratio of valley gradient to channel gradient provides an alternative definition.

Sinuosity ranges from 1.0 for perfectly straight channels to around 3 for highly tortuous river courses. Channels with a sinuosity greater than 1.5 are usually described as ❶***meandering***. A second fundamental form that channels may assume involves ❷***braiding***. This represents the extent to which flow in the channel is divided by islands or bars, that is exposed accumulations of sediment. Islands are vegetated and are relatively long-lived features, whereas bars are

less stable, being composed of unvegetated sands or gravels. A third type of channel pattern is termed ❸***anastomosing***. Anastomosing channels consist of distributaries which branch and re-join. Although many alluvial channels can be described as stable in that they are not experiencing a dramatic change in form, some change is an expected element of the behaviour of all alluvial channels since they are, at least partly, composed of material which is eroded or deposited as the stress exerted on the channel bed and banks by the flowing water changes over time. Such changes can occur in various ways (Fig. below), including the downstream migration of bars, the gradual shifting of meanders and the rapid alteration of course through ***cut-offs*** or channel diversion.

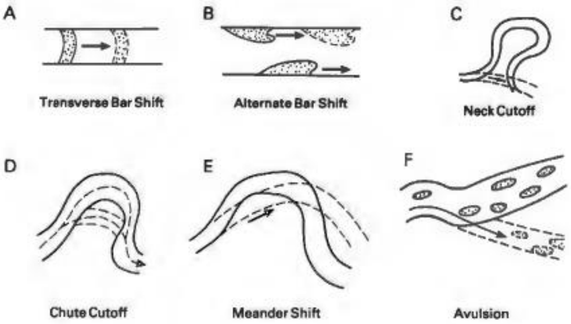


Fig. 9.8 *Types of channel changes. (A) and (B) show the downstream migration of bars or islands within channels; in (B) alternate bars are moving in association with a shifting in position of the deepest part of the channel, or thalweg, with the channel banks alternatively being protected from, and exposed to, erosion as the bars migrate; (C), (D) and (E) illustrate rapid changes in the course of meandering channels; (F) shows the establishment of a new course through avulsion. (After S. A. Schumm, (1985)* Annual Review of Earth and Planetary Sciences ***13***, *Fig. 4, p. 11.)*

Studies of alluvial channels from a wide range of environments have demonstrated that although the size of alluvial channels is controlled largely by the water discharge flowing through them, the channel pattern and shape are related primarily to the quantity and size of sediment being transported and the valley floor gradient.

A genetic classification proposed by S.A.Schumm identifies the three fundamental channel patterns and relates these to the nature of the transported sediment among other factors (Fig. below). Although the proportions of bed load and suspended load invariably change over time it is useful to distinguish between ***suspended-load channels*** transporting less than 3% of total sediment as bed load, ***bed-load channels*** transporting more than 11% as bed load, and ***mixed load channels*** with 3-11% of sediment being transported as bed load (Table below).

Table 9.2 Classification of stable alluvial channels

TYPE OF CHANNEL	BED LOAD AS % OF TOTAL LOAD	CHANNEL CHARACTERISTICS
Suspended load	<3	Width–depth ratio <10; sinuosity usually >2.0; relatively gentle gradient
Mixed load	3–11	Width–depth ratio 10–40; sinuosity usually 1.3–2.0; moderate gradient
Bed load	>11	Width–depth ratio >40; sinuosity usually <1.3; relatively steep gradient

Source: Modified from S. A. Schumm (1977), *The Fluvial System*. Wiley, New York, Table 5–4, p. 156.

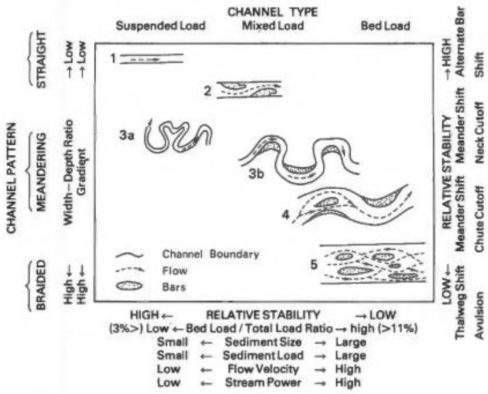


Fig. 9.9 *Classification of channel form based on pattern and sediment load. (After S. A. Schumm (1981)* Society of Economic Paleontologists and Mineralogists Special Publication ***31***, Fig. 4, p. 24.)

Suspended load channels are narrow and deep with a very low width-depth ratio. If the valley gradient is low the channel will be straight (Fig. above, pattern 1), but sinuosity becomes greater with an increase in gradient (Fig above, pattern 3a).
Mixed-load channels have a low width-depth ratio and sinuosity ranges from 1.3 to 2.0 (Fig. above, pattern 3b). Even if the channel is relatively straight the thalweg is nearly always sinuous (Fig. above, pattern 2).
Bedload channels are straight and have a high width-depth ratio. More than one thalweg tends to develop (Fig. above, pattern 4) and where bed-load transport is very high distinct bars form and a braided channel is created (Fig. above, pattern 5). Overall the mean flow velocity, the drag force on the channel bed and stream power increase, while channel stability decreases from patterns 1 to 5 (in Figure above).

Meandering channels

Meandering channels vary in form, but a number of morphological characteristics can be defined which are relatively consistent for a large proportion of rivers (Fig. below). These include the observation that meander wavelength is commonly about ten times channel width and about five times the mean radius of curvature. Natural meanders rarely have a perfectly symmetric and regular form apparently because of variations in channel bed material.

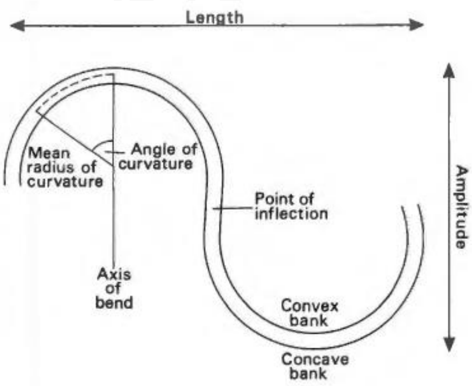


Fig. 9.10 Principal components of meander geometry.

In rivers with coarse bed material meander forms are often highly distorted. The fact that straight channels are rare and meandering channels are common raises the question of what causes meandering. The meandering behaviour of rivers may initially seem anomalous since sinuous channels take a longer, lower gradient course when a steeper, shorter course is available. But we simply have to note that airflows such as jet streams and ocean currents meander to recognize that meandering may be the normal behaviour of fluids in motion. There is an extensive literature on the origin of meanders suggesting a range of possible mechanisms for initiating and sustaining meandering behaviour. Research by hydraulic engineers has now shown that friction with the channel

bed and banks causes shear and turbulence in the water flow and the development of instabilities which promote the formation of alternating bars along the channel. A helical flow is established, with the water surface being elevated on the outer (concave) bank of each curve and return currents at depth directing the flow towards the opposite bank downstream (Fig. below A). The outer bank is eroded as a result of the higher flow velocity, whereas deposition takes place along the inner (convex) bank forming a point bar (Figs below B).

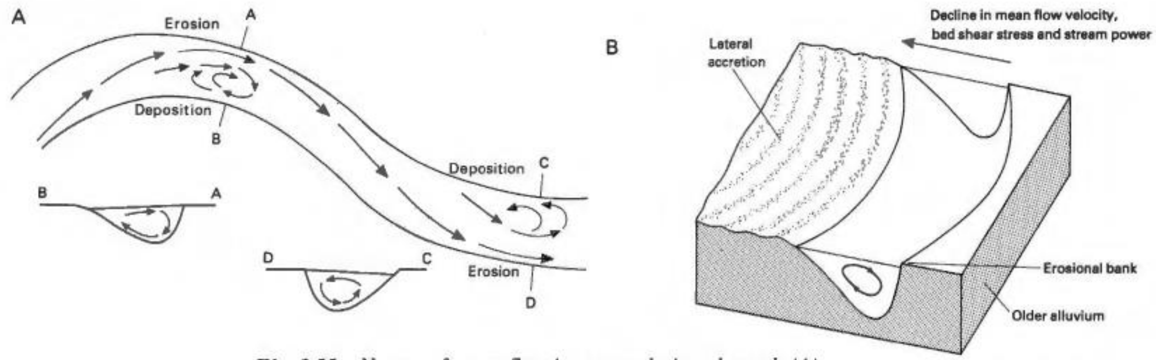


Fig. 9.11 Nature of water flow in a meandering channel: (A) downstream changes in cross-sectional channel form and sites of erosion and deposition; (B) water movement in a channel bend and associated stresses exerted on channel bed material. ((B) based on J. R. L. Allen (1970) Physical Processes of Sedimentation. *Allen and Unwin, London, Fig. 4.5, p. 133(B).)*

Braided channels

The development of braided channels is favoured by several factors (Fig. below).

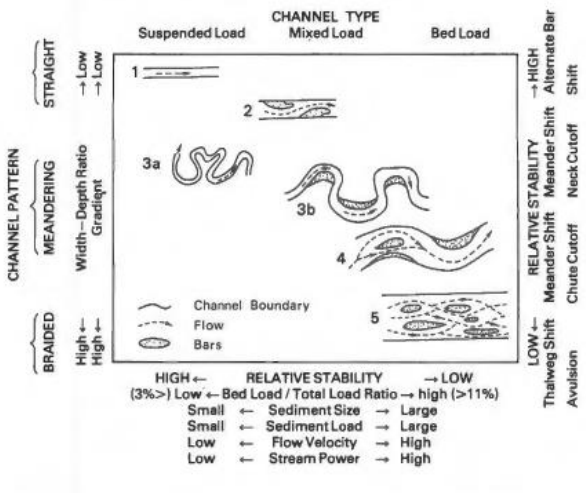


Fig. 9.9 Classification of channel form based on pattern and sediment load. (After S. A. Schumm (1981) Society of Economic Paleontologists and Mineralogists Special Publication *31*, Fig. 4, p. 24.)

In addition to a steep channel gradient the most important appear to be a large proportion of coarse material being transported as bed load, and readily erodible bank material which enables channel shifts to occur with relative ease. Once formed bars in braided channels can become rapidly vegetated and thereby stabilized as islands. This points to the role of the highly variable discharge which is typical of many braided rivers. By promoting alternating channel

degradation and aggradation, large fluctuations in discharge help to suppress the establishment of vegetation on braided-channel bars. As discharge declines after a flood peak the coarse bed load is the first to be deposited in the channel. This material forms the nucleus of bars which grow downstream as the flow velocity is reduced and finer sediment accumulates. With further decrease in discharge the water level progressively falls and the bars are gradually exposed. During subsequent floods some, or all, of the bars in a braided channel may be submerged depending on the discharge attained. During large floods braided channels can experience major diversions of flow.

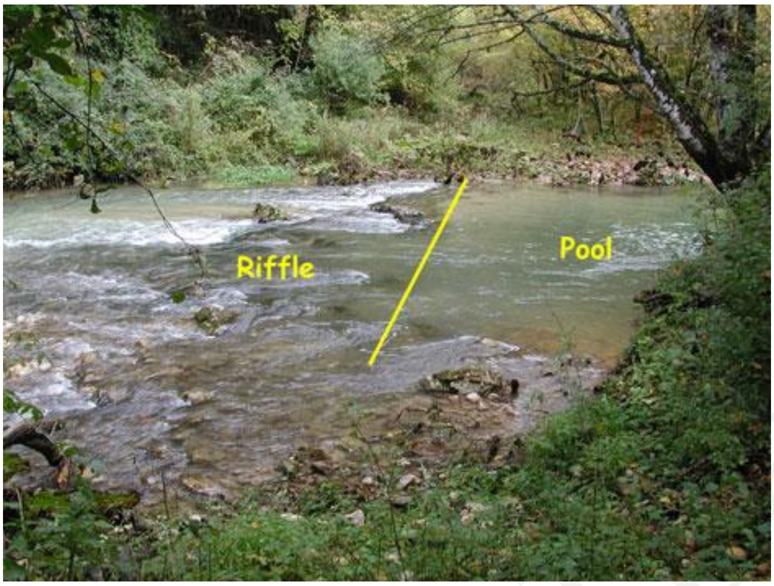
Alluvial channels: hydraulic geometry

In addition to changes in plan form, alluvial channels may also adjust their cross-sectional form in response to changes in discharge. These cross-sectional components of channel form were termed **hydraulic geometry**.

Alluvial channels: longitudinal form

A further component of downstream adjustments in alluvial channels is that of changes in channel gradient, and these can be considered at a range of scales. Looking at long sections of a river course, channel gradient is generally seen to decrease downstream giving a concave-up longitudinal profile (long profile). This downstream decrease in gradient is accompanied by an increase in discharge, and it can be explained by the ability of a river to transport the same quantity and calibre of sediment load in a lower gradient channel as discharge increases. Average sediment size does, of course, tend to decrease downstream as a result of sorting, attrition and weathering, and at a given flow velocity finer sediment can be transported in lower gradient channels. None the less, the relationship between channel gradient and sediment size is not a simple causal one since a progressive decrease in sediment calibre downstream could arise because coarser sediment, once deposited, cannot be transported over lower gradients. Neither channel gradient nor sediment size are independent variables and there is a relationship operates between them. A progressive decline in channel gradient in alluvial channels can be disturbed by either changes in base level downstream or by tectonic activity causing vertical movements of the channel bed itself. A fall in base level can give rise to a Knickpoint (a knickpoint or nickpoint is part of a **river** or channel where there is a sharp change in channel slope, such as a waterfall or lake. Knickpoints reflect different conditions and processes on the **river**, often caused by previous erosion due to glaciation or variance in lithology), a discontinuity in the longitudinal profile of a river, and the downstream channel gradient is thereby increased. Alluvial channels will tend to adjust rapidly to such a change through an increase in flow velocity which leads to increased erosion in the steepened reach. Particularly in channel beds composed of material of a wide range of sizes, shallow and deep reaches are seen to alternate downstream. The shallows are formed by high points in the channel bed known as ***riffles***. They are composed of coarser material and are characteristically spaced at about five to

seven times the channel width. The deep reaches are called ***pools*** and have a bed of finer calibre than the riffles on either side.



At low stage the water surface is steeper over riffles and the flow is shallower and more rapid, whereas over pools the water surface has a lower slope and the flow is deeper and slower. Theoretical work has shown that even in straight, uniform channels, turbulence along the channel boundary will generate large-scale roller eddies associated with alternating acceleration and deceleration of flow (Fig. below (A)). The dimensions of these eddies are a function of channel size, and their spacing averages about six channel widths.

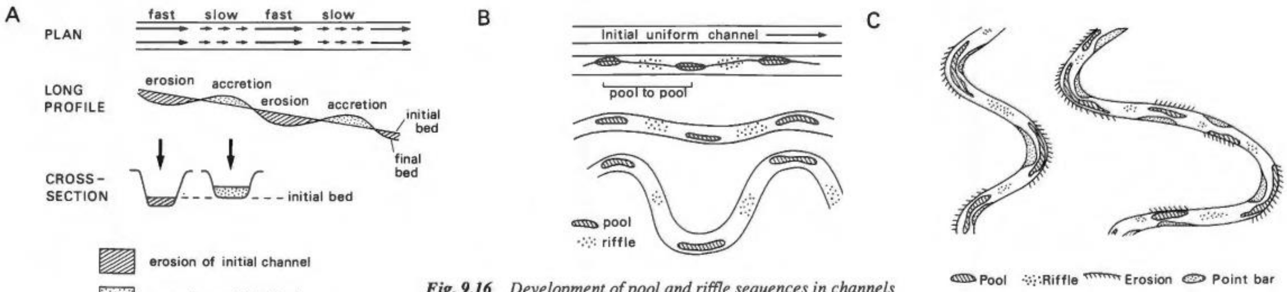


Fig. 9.16 Development of pool and riffle sequences in channels and their relationship to meander morphology: (A) erosion and accretion of a channel bed corresponding to alternate zones of fast and slow flow; (B) transformation of a straight to a meandering channel in relation to pool spacing; (C) the development of additional pools and riffles with the lengthening of a meandering channel. (After K. Richards (1982) Rivers: Form and Process in Alluvial Channels. *(Methuen, London, Fig. 7.2C, p. 184; and G. H. Dury, (1969) in: R. J. Chorley (ed.)* Water, Earth and Man *(Methuen, London) Fig. 9.II.4, p. 180.)*

It seems likely that these downstream alternations of fast and slow flow lead to the development of zones of erosion (pools) and zones of sediment accretion (riffles) which eventually assume an equilibrium form. It is the riffles that are the sites of accretion and this may seem to contradict the faster flow observed over riffles.

As a result of the greater shear stress exerted on the channel boundary in pools during channel-forming high discharges, banks are eroded preferentially in pools. There is consequently a correspondence between the spacing of pools and riffles and the wavelength of the meanders that eventually develop,

meander wavelength being approximately twice the pool spacing (Fig. above (B). This situation may change, however, if the channel between each meander loop is lengthened in which case new pools and riffles may develop (Fig. above (C).

The alluvial channel system

It is a speculative attempt to illustrate the way in which alluvial channels can adjust to changes in the independent variables of valley slope, discharge and bed material size and sorting. Under a relatively consistent hydrological regime alluvial channels experience constant adjustments in response to changes in hydraulic and sedimentological variables which maintain a more or less stable channel form.

Bedrock channels

Bedrock channels represent a dramatic contrast with alluvial channels in that they are capable of only very slow adjustment of form in response to changes in discharge, sediment load, gradient and other factors. Bedrock channels cannot, therefore, be realistically analyzed in terms of the concept of grade because their response time is so long. It is generally held that rivers are most commonly cut into bedrock in the upper part of their courses where channel gradients are usually steeper and the coarseness of the load they carry is greater and thus more effective as an agent of abrasion.

<u>Reference:</u>

Summerfield, M. A., 2013, Global Geomorphology. An Introduction to the Study of landforms-Routledge, London and New York, 537 pp.

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Fluvial Landforms

(Part TWO)

Lecture SIX

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Fluvial depositional landforms

Deposition by rivers occurs predominantly in the bottom of valleys where gradients are low, at locations where there is a significant change in gradient, or where channelled flow diverges, and is consequently reduced in depth and velocity. It must be emphasized, however, that deposition is not exclusive to any part of the fluvial system. Four major types of fluvial deposition can be distinguished: **channel deposits** and **channel margin deposits** which accumulate within and along river channels; **overbank deposits** formed when bankfull discharge leads to the deposition of fine sediments beyond the confines of the channel itself; and **valley margin deposits** which accumulate at the base of valley slopes (Table below).

Table 9.3 Classification of valley sediments

PLACE OF DEPOSITION	NAME	CHARACTERISTICS
Channel	Transitory channel deposits	Primarily bed load temporarily at rest; part may be preserved in more durable channel fills or lateral accretions
	Lag deposits	Segregations of larger or heavier particles, more persistent than transitory channel deposits
	Channel fills	Accumulations in abandoned or aggrading channel segments; ranging from relatively coarse bed load to fine-grained oxbow lake deposits
Channel margin	Lateral accretion deposits	Point and marginal bars that may be preserved by channel shifting and added to overbank floodplain
Overbank floodplain	Vertical accretion deposits	Fine-grained sediment deposited from suspended load of overbank flood water; including natural levee and backswamp deposits
	Splays	Local accumulations of bed-load materials spread from channels on to adjacent floodplains
Valley margin	Colluvium	Deposits derived chiefly from unconcentrated slope wash and soil creep on adjacent valley sides
	Mass movement deposits	Earthflow, debris avalanche, and landslide deposits commonly intermixed with marginal colluvium; mudflows usually follow channels but also spill overbank

Source: After P. C. Benedict *et al.* 1971, *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers* 97, Table 2 – Q.1, p. 44.

1. Floodplains

Except in mountainous terrain most river channels are flanked by an area of subdued relief termed a **floodplain** formed by deposits laid down when the river floods. Low magnitude - high frequency floods cover only a part of the floodplain, and it is only during rare major floods that the entire floodplain is inundated. Depths of flood water range from a few centimetres up to several metres for large floods in major rivers. As with the dimensions of river channels, the width of floodplains in most rivers is roughly proportional to discharge. The active floodplain of the lower Mississippi is about 15 km across, but, as with most flood- plains, this lies within a much more extensive area of older floodplain deposits. Where closely spaced river channels flow across a region of subdued topography adjacent floodplains may merge to form extensive alluvial plains. Two modes of sediment deposition can be identified; **lateral accretion deposits** are laid down as the river channel migrates across

its floodplain, while **vertical accretion deposits** accumulate beyond the confines of the channel during floods. Lateral accretion deposits are formed by both meandering and braided channels, but in the latter case a large part of the floodplain effectively operates as part of the active river channel at anyone time. Rapid changes in channel courses in braided rivers leave abandoned channels floored by coarse bed-load material which, if they survive reworking during subsequent channel shifts, are eventually filled by organic material and fine sediments laid down during floods. Lateral accretion by meandering channels occurs through, the more gradual growth of point-bar deposits on the convex side of meander bends (Fig below).

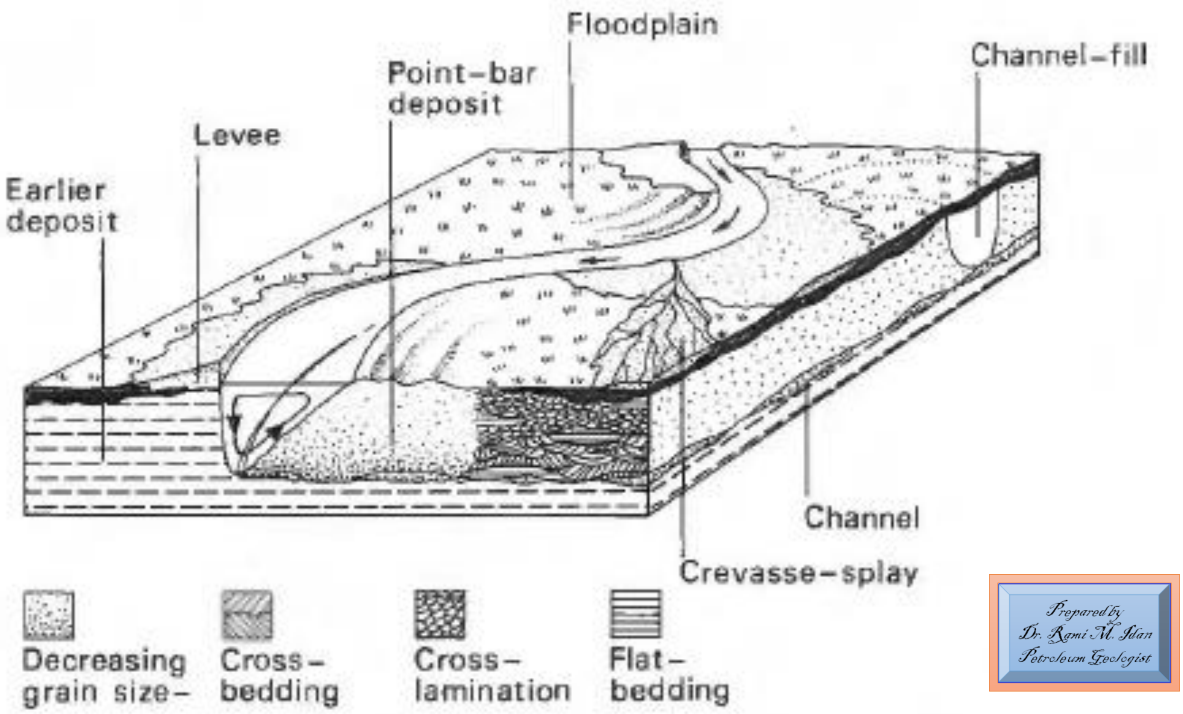


***Fig. 9.20** Depositional forms developed along a meandering channel. Note how levees give rise to steep high banks where they are undercut on the concave side of meander bends. After J. R. L. Allen (1964)* Sedimentology *3, Fig. 4, p. 168 and 1970,* Geological Journal *7, Fig. 1, p. 131.)*

The overrunning of one meander loop by another produces a meander cut-off, or chute, which shortens the channel course. The abandoned meander loop is gradually isolated by the deposition of bed material at each end by the main channel and becomes an oxbow lake. Eventually the oxbow lake accumulates channel-fill deposits laid down during floods. Whereas lateral accretion deposits are constantly being formed and reworked, vertical accretion occurs only during floods when bankfull discharge is exceeded. Morphologically the

most prominent depositional landforms in most floodplains are levees. These are ridges which lie parallel to the river channel and may rise to a height of several metres above the general level of the floodplain. When bankfull discharge is exceeded, water starts to spread rapidly across the floodplain. The decrease in depth in comparison to the river channel rapidly reduces the competence of the flow and the coarser fraction of the suspended load is promptly deposited to form levees. The presence of levees affects the movement of flood waters once bankfull discharge is exceeded. Initially the water is ponded by levees, but as the level rises it breaks through forming **crevasses**. The accelerated flow through crevasses is able to transport a relatively coarse suspended load which is quickly deposited as the flow disperses to form fan-shaped **crevasse-splay deposits** in the **backswamp zone** beyond the levee barrier. Flow diverted through a crevasse may travel many kilometres downstream before it is able to regain access to the main channel. This may not occur until a tributary is reached, but tributaries themselves may be diverted downstream by levee development along the main channel. Occasionally the main channel itself may shift its course into the backswamp zone. Many of the world's floodplains have thick accumulations of vertical accretion deposits, but this is probably a reflection of the particular conditions of the past 10 000 a or so when most river systems have been responding to a rise in base level in response to the post-glacial rise in sea level. Under conditions of stable base levels we would expect lateral accretion deposits to be quantitatively more significant.

2. **Alluvial fans**

An alluvial fan is a body of sediment whose surface form approximates to the segment of a cone which radiates downslope from a point on a mountain front, usually where a stream emerges. Most alluvial fans have a radius of less than 8 km, but under certain conditions fan radii may exceed 100 km. In form they have concave-up long profiles, but convex-up cross-profile. The mean surface slope of fans generally ranges from 1 to 5, but at the fan apex gradients can exceed 10°. Since the slope of streams emerging at the mountain front is usually similar to the gradient of the upper fan surface, it seems that the deposition that causes fan building occurs primarily as a result of the sudden change from a confined to an unconfined condition as the stream leaves the mountain ***gorge*** rather than as a result of reduced channel gradient. The calibre of sediment deposited generally decreases down-fan, but fan deposits are often poorly sorted as they are frequently laid down by torrential floods. In catchments containing abundant, easily erodible debris, sediment concentrations may reach the point where high-viscosity debris flows are generated. Alluvial fans develop where there is a relatively abundant sediment supply, adequate relief for vertical fan

growth and a suitable location for sediment accumulation. Favourable environments include active faulted mountain fronts where the rate of mountain uplift, or adjacent basin subsidence, exceeds the rate of ***down cutting*** of the trunk stream feeding the fan. Individual fans along a mountain front may grow laterally to the extent that they coalesce to form a continuous piedmont sedimentary apron. In arid environments such landforms are known as **bajadas**. In humid environments, where alluvial fans tend to be larger, the coalescence of fans can produce immense, gently inclined alluvial slopes, such as those that flank the southern margin of the Himalayas. Although some alluvial fans in arid environments are simple undissected forms, most are dissected - that is, the trunk channel is entrenched below the upper part of the fan surface and deposition is focused towards the fan toe (Fig. below).

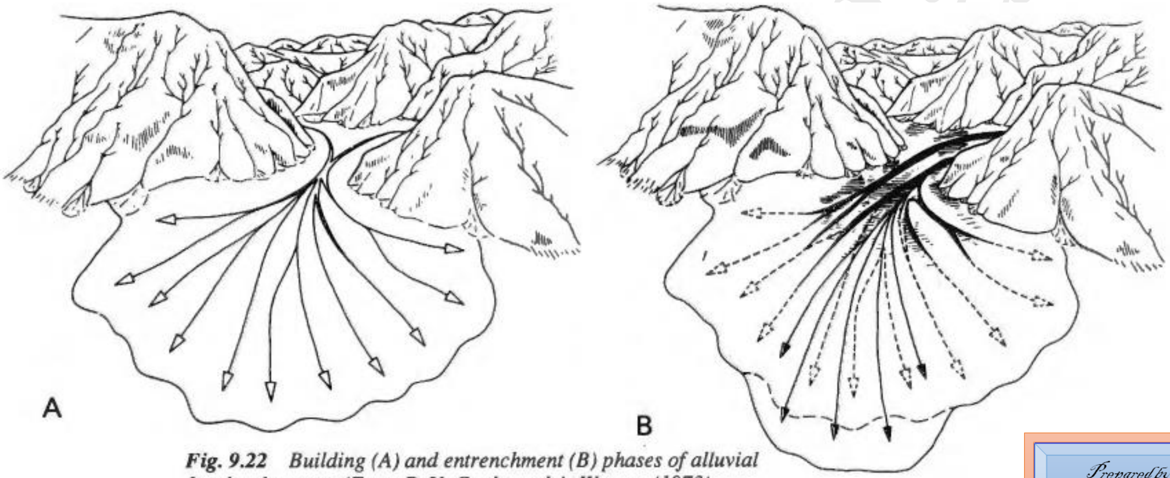


Fig. 9.22 Building (A) and entrenchment (B) phases of alluvial fan development. (From R. U. Cooke and A. Warren (1973) Geomorphology in Deserts. *Batsford, London, Fig. 3.8, pp. 186 and 187, after L. K. Lustig (1965)* United States Geological Survey Professional Paper ***352-F****, Fig. 137, p. 185.)*

Fan-head trenching will occur if sediment yield decreases, or channel gradient and flow velocities increase. Such changes can be brought about by climatic changes, increased tectonic activity or even human-induced land use changes in the upstream basin. But we might also anticipate that fan incision will occur in the long term without any change in external factors simply as a result of the reduction in sediment yield as relief, and therefore rates of erosion in the basin, progressively declines through time as the moutain mass is gradually lowered.
Not surprisingly, fan area has been found to be positively correlated with source basin area, although there may be an order of magnitude difference in fan area for a given contributing area. Some of this variation may be explained by lithological factors. Basins in sandstone, for instance, have been found to have smaller associated alluvial fans than those underlain by shales or mudstone. This is presumably because sandstone is less easily eroded and gives rise to lower sediment yields. Tectonic factors can also influence fan area. In Death Valley progressive eastward tilting has confined the east-side fans but enabled

the west-side fans to extend several kilometres beyond the mountain front. Alluvial fans in humid environments are distinct in a number of respects from those in arid environments. Apart from being generally smaller and steeper, fans in dry regions are fed by ephemeral stream channels and active deposition tends to move unpredictably from one part of the fan to another. In large arid fans only a very small proportion (usually < 5 %) is active during anyone flood event. In contrast humid fans are fed by perennial, and often braided, channels which tend to migrate progressively across the fan surface.

3. **River terraces**

Changes in channel gradient, discharge or sediment load can lead to a river channel incising into its floodplain. The original floodplain is thereby abandoned and is left as a relatively flat bench, known as a **river terrace**, which is separated from the new floodplain below by a steep slope. As well as forming in the alluvial fill of river valleys, river terraces can also be cut into bedrock and in this case they are covered by only a thin sediment veneer. The term strath terrace is sometimes applied to these forms. River terraces are inclined downstream but not always at the same inclination as the active floodplain. A valley side may contain a vertical sequence of terraces. The lowest will be the youngest and may retain traces of floodplain morphology, while the highest will be the oldest and will usually be partly degraded. Paired terraces form when vertical incision is rapid in comparison with the lateral migration of the river channel (Fig. below).

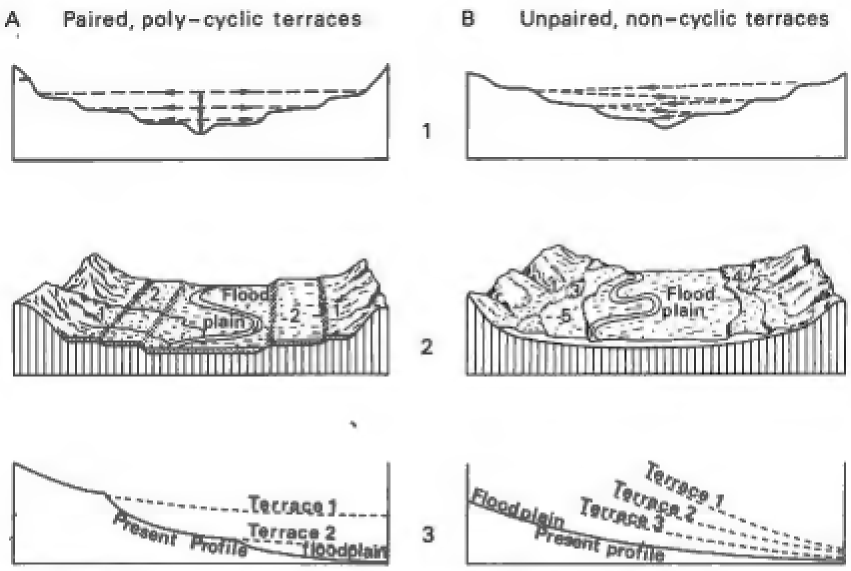


Fig. 9.25 Paired and unpaired terraces. Note how the vertical spacing of terraces is more or less retained in paired terraces but converges downstream in unpaired terraces. (After B. W. Sparks (1972) Geomorphology *(2nd edn). Longman, London, Figs 9.5, 9.7 and 9.8, pp. 296 and 297; and W. D. Thornbury, (1969)* Principles of Geomorphology *(2nd edn). Wiley, New York and London, Fig. 6.9, p. 157 after Longwell* et al.*)*

Morphologically similar features can, however, be produced by resistant beds in flat-lying strata. These give rise to structural benches and structural controls must be eliminated before a river terrace interpretation is accepted. Unpaired terraces form where lateral shifting of the channel is relatively rapid; this results in the river cutting terraces alternately on each side of the valley floor. Since valleys experience phases of aggradation and degradation, once formed terraces can subsequently be covered by sediment. Thus, we would expect complex sequences of buried and partially eroded terraces to lie beneath modern floodplains (Fig. below).

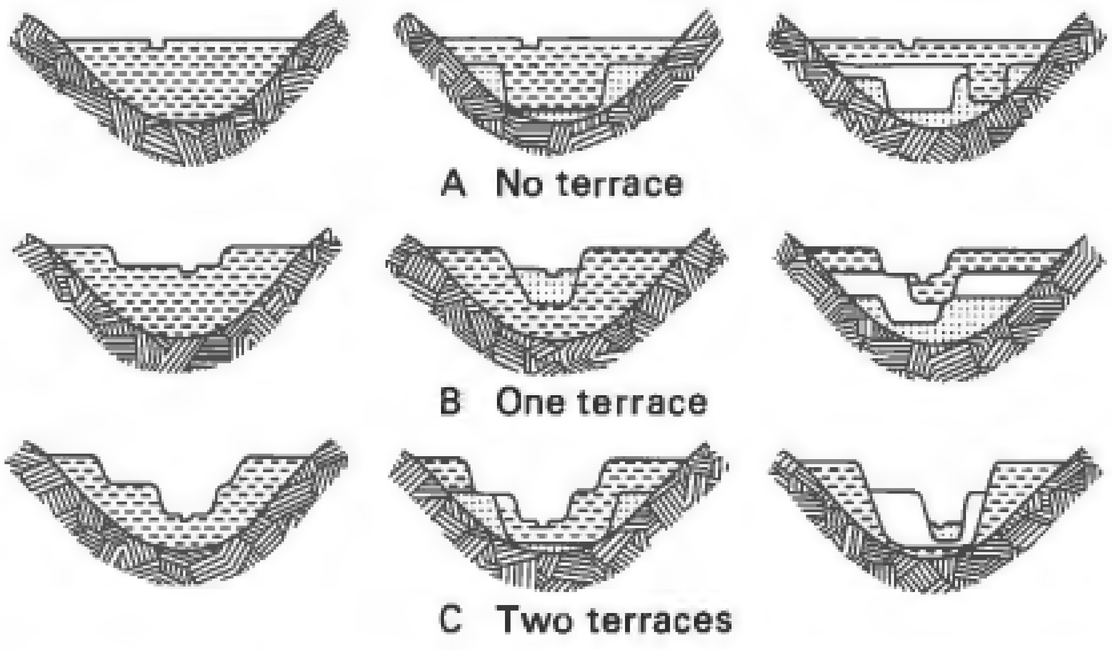


Fig. 9.26 *Valley cross-sections showing some possible terrace and alluvial-fill combinations. (After L. B. Leopold and J. P. Miller (1954)* United States Geological Survey Water-Supply Paper ***1261****, Fig. 2, p. 5).*

Terraces can be formed under a wide range of circumstances. Unpaired terraces do not, in fact, require any change in external conditions since they can form simply through the progressive incision of a river channel migrating from one side of its floodplain to the other. Many episodes of channel entrenchment and terrace formation probably arise from changes in base level or fluctuations in climate, although tectonic deformation along river courses can also lead to local terrace development. A fall in base level can lead to terrace development if it generates a downstream increase in channel gradient. This is then propagated upstream through knickpoint retreat. The past 3 Ma have seen frequent changes in sea level of up to 100 m or so, and this has created conditions highly favourable to terrace formation. Many other river terraces currently present in the landscape are attributable to climatic fluctuations which lead to changes in

channel discharge and sediment load. As with base level changes, climatic fluctuations have' also been frequent and rapid over the past 3 Ma or so. The response of river systems to such environmental changes are complex and can vary dramatically depending on local channel conditions. For instance, during glacials the Mississippi incised into its floodplain in its lower valley producing a series of terraces, but in the upper part of its basin the great increase in sediment supplied from the Laurentide ice sheet led to channel aggradation. The tendency of fluvial systems to aggrade or incise is particularly sensitive to climatic fluctuations in semi-add regions. This is because in such environments modest changes in annual precipitation can produce significant changes in vegetation cover which are reflected in large changes in the rate of sediment supply to stream channels. In the south-west USA many ephemeral stream channels, known locally as arroyos, show evidence of phases of aggradation and entrenchment.

Fluvial systems in limestone terrains

The susceptibility of rocks composed predominantly of calcium carbonate to the weathering process of carbonation gives rise to a distinctive form of fluvial activity on limestone terrains. The distinctive nature of fluvial activity on limestone arises from the high permeability of the bedrock and the diversion underground of a variable, but in many cases, significant proportion of runoff. The permeability of rock is in part related to its **primary porosity**, that is, the proportion of the rock occupied by voids. (Strictly it is the **effective porosity** that is important - that is, the proportion of voids that are connected and thus able to transmit water.) In many types of limestones, however, it is their **secondary porosity**, arising from the solutional enlargement of joints, fissures and bedding planes, that makes by far the greatest contribution to their permeability. Ultimately such enlargement can give rise to caves. Since such secondary porosity increases over time as calcium carbonate is removed in solution, the proportion of subsurface flow depends both on the lithological properties of the limestone and on the length of time that it has been subject to active dissolution.

1. **Surface drainage**

In addition to a topography characterized by numerous closed depressions, karst landscapes may also contain fluvially eroded valleys. In limestone regions, such as those underlain by chalk, where karst has not developed the landscape may be predominantly one of fluvially eroded valleys.

2. **Subsurface drainage**

The size and frequency of joints and fissures in limestone have an important influence on the type of subsurface water flow that occurs. At one extreme

limestones, such as chalk, possess a high density of small fractures and consequently surface infiltration rates are uniformly high. Chalk is mechanically weak so fissures can only be enlarged to a certain point before they collapse. The rock is essentially permeable throughout (although water movement is still concentrated in the larger fissures) and there is a distinct and continuous water table. At the other extreme are the majority of limestones which are densely cemented and through which the greater amount of water flow occurs along major joints and fissures. This creates discrete systems of subsurface flow with no single continuous water table, although in time a somewhat more integrated system of subsurface drainage may develop as conduits are enlarged to the point where the rock is honeycombed with linked cave systems and there is one general water table level. In limestone hydrology the term **vadose zone** is usually applied to the region lying above the water table in which voids may contain either air or water. Below the water table is the **phreatic zone** (equivalent to the ground water zone in which all voids are completely filled with water. At any point below the water table there is a pressure equal to atmospheric pressure plus the pressure head (the product of the depth of water and its unit weight). The hydraulic head at this point is the sum of the pressure head and the elevation above some datum. Points of equal hydraulic head define equipotential surfaces within the phreatic zone. Water moves from high to low potential along flow paths which are orthogonal to these equipotential surfaces (Fig. below).

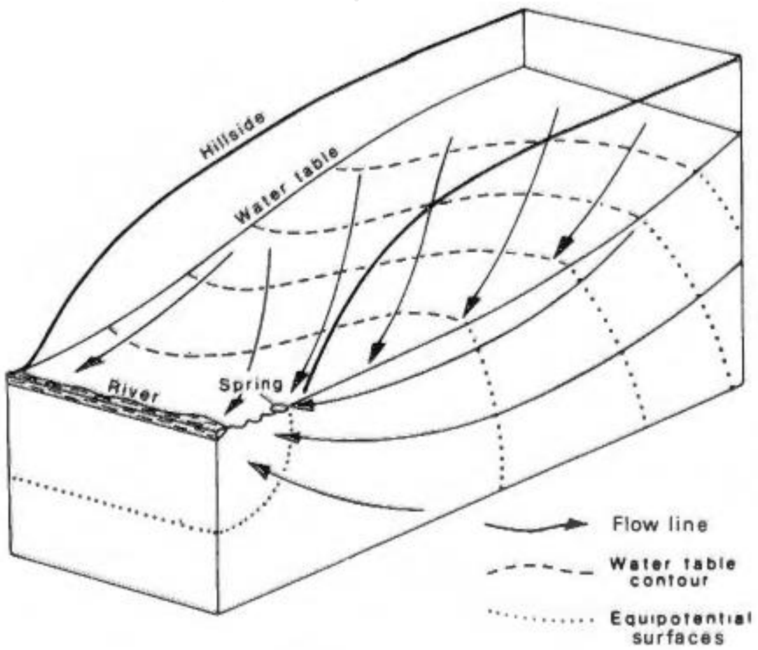


Fig. 9.29 Water table contours, equipotential surfaces and flow lines. (After D. Ford and P. Williams (1989) Karst Geomorphology and Hydrology *Unwin Hyman, London. Fig. 5.7, p. 138.)*

In lithologies with relatively uniform permeability, including limestones such as chalk, water flow in the phreatic zone approximates to Darcy's law.

Box 9.2 Darcy's law

The flow of water flowing through a porous medium is given by

$$v = -K \frac{dh}{dl}$$

where v is the specific discharge (discharge per unit cross-sectional area), K the hydraulic conductivity and dh/dl the hydraulic gradient. The hydraulic gradient is determined by the length of flow (dl) and the difference in the elevation of the inflow and outflow points (dh). The negative sign on the right-hand side of the equation indicates a loss in hydraulic head in the direction of flow. Note that, strictly, hydraulic conductivity is not synonymous with permeability; the former refers to the ability of a medium to transmit fluids in terms of the properties of both the medium *and* the fluid (specifically its density and dynamic viscosity), whereas the latter refers to material properties only (such as void size and connectivity).

Reference:

Summerfield, M. A., 2013, Global Geomorphology. An Introduction to the Study of landforms-Routledge, London and New York, 537 pp.

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Aeolian Processes and Landforms

Lecture SEVEN

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Aeolian activity

Wind is a comparatively weak geomorphic agent over much of the Earth's surface, but in areas free of vegetation it can have significant effects. In the arid regions of the subtropics vast sand seas and extensive grooved bedrock surfaces are a testament to the power of wind action.

The effectiveness of wind action is limited by a number of factors, and on a global basis is a far less powerful erosional agent than fluvial activity. Compared with water, air has a low density and viscosity so only very fine particles can be carried in suspension, except at very high wind speeds. Moreover, vegetation greatly reduces wind speeds near the ground, and together with moisture it tends to bind surface particles together and prevent them from being entrained by the wind. Consequently Aeolian activity is only effective in areas which lack a relatively complete vegetation cover and where the surface material dries out at least occasionally.

1. **Global distribution of Aeolian landforms:**

The most striking aeolian landforms are **dunes**. These are accumulations of wind deposited particles and they can assume a bewildering variety of forms. The great majority of dunes are composed of sand, although silt and clay-sized material may accumulate into dune-like features under certain conditions. Very extensive blankets of silt-sized material, known as **loess**, are found in those mid-latitude areas that were marginal to the Pleistocene ice sheets, and, to a more limited extent, at lower latitudes.

Quartz is by far the predominant component of desert sands, both because of its abundance as a rock-forming mineral and because of its resistance to chemical decomposition and abrasion.

2. **Wind characteristics:**

The largescale properties of desert winds are determined by the general circulation which, at the latitude of the world's hot deserts, is dominated by sub-tropical high-pressure systems. Of particular significance to the movement of particles by the wind is the vertical variation of wind speed above the ground. Close to the surface wind speeds are reduced by friction, and the magnitude of this effect is largely a function of surface roughness. As wind speed and surface roughness increase, the airflow becomes more turbulent and the potential for erosion is consequently increased.

3. **Effect of surface characteristics:**
   The character of desert surfaces merits consideration because surface roughness and the size and cohesion of surface particles significantly affect the ability of the wind to erode.
4. **Sediment entrainment and transport**

Both water and air are fluids and there are close similarities between the mechanism of sediment entrainment in airflows and water flows (see Section 8.3.2). Movement is resisted by the weight of particles, and by friction and interparticle cohesion. Movement is induced by drag and lift forces, and the impact of grains already in motion.

Aeolian erosion

1. Deflation and abrasion

Aeolian erosion involves two processes; **deflation**, which is the removal of moveable particles by the wind, and **aeolian abrasion**, which is caused by the bombardment of rock and other surfaces by particles carried in the airflow. Deflation involving sand-sized particles is rather localized since sand grains cannot be moved great distances except over long periods of time. Silt, and especially clay-sized material, on the other hand, can be lifted by turbulence and carried in suspension in the atmosphere while the very finest material can be carried great distances in dust storms.

Rock- and boulder-covered surfaces can be abraded by particles of a range of sizes. Although it has been widely thought that sand-sized material is the most effective, some research has pointed to the role of fine silt-sized particles and dust, especially in producing fluted or grooved surfaces.

2. Erosional landforms

Although aeolian erosion may be active over alluvial plains and on beaches, wind-eroded landforms are rarely preserved in such environments because of destruction by fluvial processes or wave action. Only in arid areas, where other denudation agents are lacking or weakly active, are aeolian erosional landforms abundant. A significant proportion of many deserts is covered not by sand, but by lag deposits comprising particles of gravel size or coarser (Fig. 10.5). Lag deposits usually form a thin layer lying over predominantly finer material (Fig. 10.6). Where the stone cover is continuous such surfaces are known as desert **pavements**, or **stone pavements**. It was long thought that they form by deflation, with the wind removing finer particles from poorly sorted deposits such as alluvium.

- *Small-scale forms:* A characteristic feature of stony desert surfaces is the presence of faceted cobbles and pebbles called **ventifacts (Fig. 10-7)**.
- *Intermediate-scale forms:* Much larger than these small-scale bedrock features are a variety of grooved forms and shallow depressions with dimensions of tens to hundreds of metres. The most characteristic grooved form is the yardang, a streamlined parallel ridge usually less than 10 m high and 100 m or more in length aligned with, and typically tapering away from, the direction of the prevailing wind **(Fig. 10.8)**. **Yardangs** are most commonly developed in soft lithologies such as lacustrine sediments and are numerous in some desert lake beds **(Fig. 10.9).**

Fig. 10.7 Ventifact on a gravel lag surface in the Namib Desert in a region of strong coastal winds.

Fig. 10.8 A mud yardang in the Kharga Depression, Egypt. The unidirectional prevailing wind is from the right. Sand-blasting is confined to the blunt, windward face, and the leeward tapering of the yardang indicates the importance of erosion by fine particles carried in suspension in secondary flows across the long leeside tail. (Photo and interpretation courtesy M. I. Whitney, from M. I. Whitney, (1985) Journal of Geological Education **33**, *Fig. 1, p. 94.)*

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Similar in scale to yardangs are the shallow depressions found across many desert regions of low relief. These have been given a variety of local names but are generally referred to as **deflation hollows**.

- *Large-scale forms:* The basins range from landforms a few metres deep and upwards of 100 m across, such as the **pans** which are so abundant in parts of southern Africa, to very large features over 100 m deep and more than 100 km across. The smaller, shallow basins probably represent localized

deflation; some are orientated along drainage lines whereas others are located in troughs between dunes. But in both situations their long axes are aligned with the prevailing wind.

Depositional landforms

1. Basic depositional forms:

Although capable of movement by surface creep and saltation, sand grains spend the vast majority of time in storage in sand accumulations which vary enormously in size and form. The smallest depositional features are called **ripples** and consist of regular, wave-like undulations orientated at right angles to the direction of the prevailing wind (Fig. 10.11).

Dunes are much larger depositional forms having typical heights of 5 to 30 m and wavelengths of 50 to 300 m. Some dunes, however, attain even greater dimensions with heights of up to 400 m and wavelengths up to 4km (Fig. 10.12). The terms **megadune**, are sometimes applied to these very large depositional forms.

Fig. 10.11 *Ripples on the surface of a dune, north of Walvis Bay, Namibia. The scale is indicated by the lens cap in the bottom right corner. Regular ripples such as these form in well-sorted sand, whereas irregular ripples are produced in poorly sorted sediments.*

a. ***Ripples:*** are asymmetric in cross-section with windward slopes around 10° and lee-side gradients near the angle of repose of dry sand of 30-35°.
b. ***Dunes:*** For a dune to form a patch of sand must first begin to accumulate. This occurs where the wind speed is reduced by an increase in surface roughness, or by primary instabilities in the airflow.

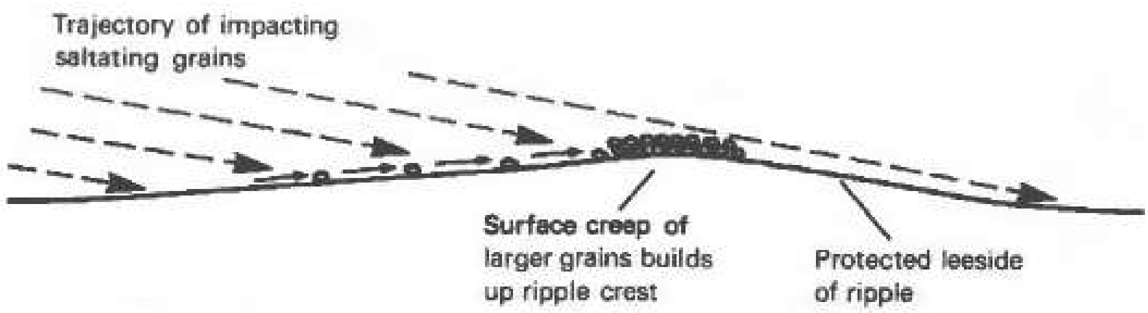


Fig. 10.14 Mechanism of ripple formation according to the model of R. A. Bagnold.

2. Classification of dune morphology

None the less, it is convenient to distinguish between **free dunes**, whose form is primarily a function of wind characteristics, and **impeded dunes** whose morphology is influenced significantly by the effects of vegetation, topographic barriers or highly localized sediment sources.

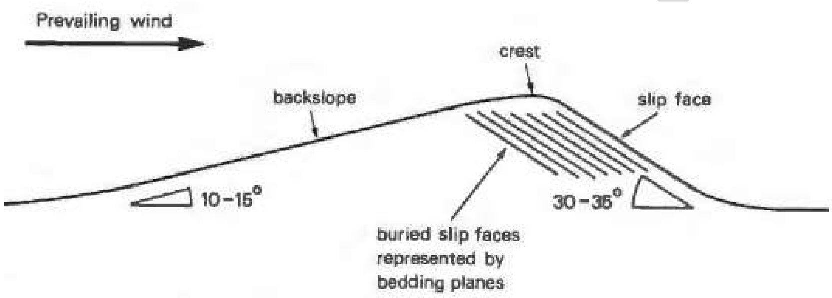


Fig. 10.16 Basic components of the cross-profile of a transverse type dune with a single slip-face orientation. Linear dunes contain two opposing slip-face orientations and star dunes multiple slip-face orientations.

3. Classification of free dunes:

Table 10.1 Classification of basic types of free dune

	DUNE TYPE	MORPHOLOGY
Transverse forms	**Transverse ridge**	Asymmetric ridge
	Barchanoid ridge	Row of contiguous crescentic forms
	Barchan	Crescentic form
	Reversing	Asymmetric ridge
	Linear	Symmetric ridge; straight to sinuous in plan
	Star	Central peak with three or more arms
	Dome	Circular or elliptical mound

The morphology of dunes represents a response to the complex interaction between secondary flows in the wind and the form presented by the sand surface. Cause and effect are so intimately related that it is frequently difficult to establish whether particular types of airflow around dunes are instrumental in their formation or merely a result of their existence.

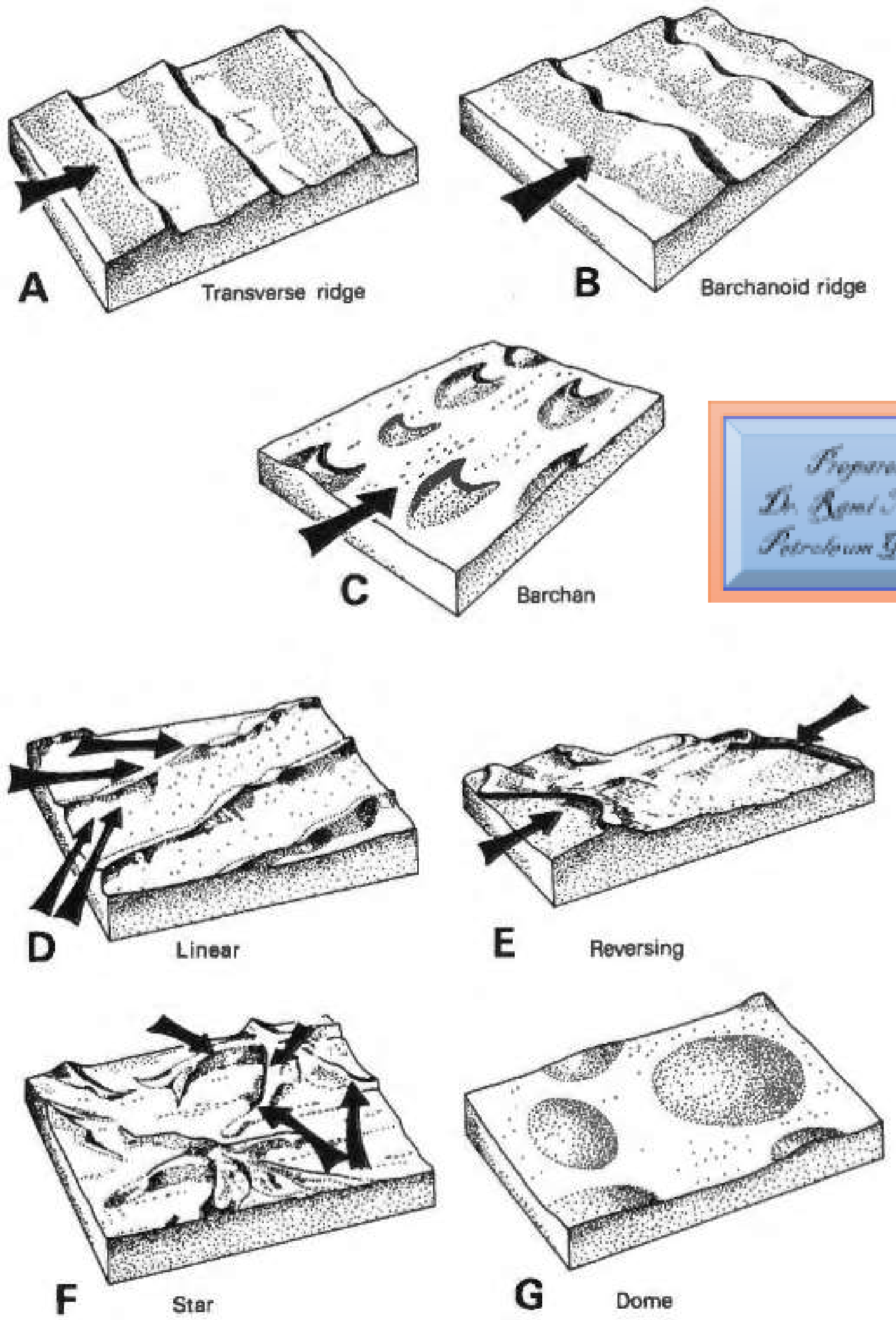


Fig. 10.17 *Morphology of the major types of free dune. Arrows indicate the probable formative prevailing wind direction(s), although note that the origin of linear dunes is controversial (see Section 10.3.5.2). (After E. D. McKee (1979)* United States Geological Survey Professional Paper ***1052****, Figs. 3–5, 7, 10–12, pp. 11–13.)*

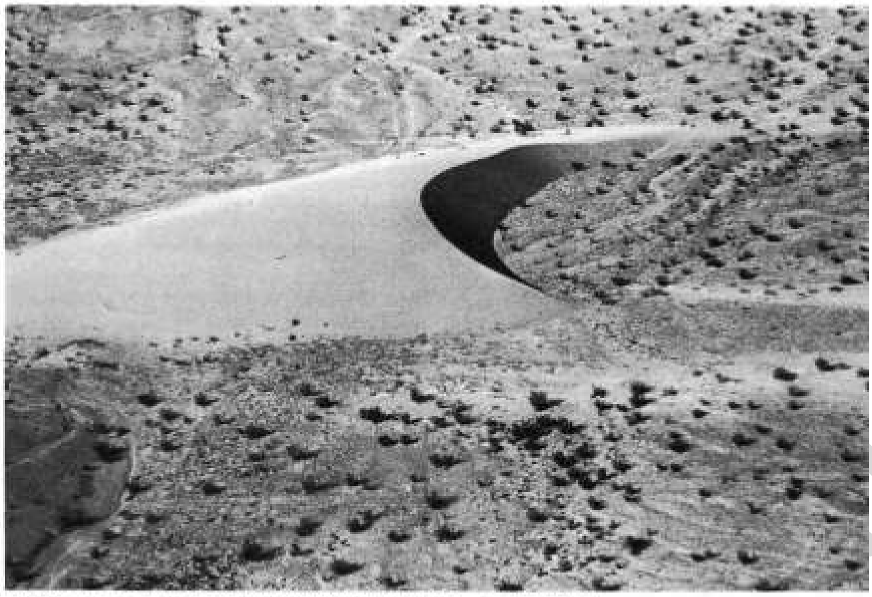


Fig. 10.18 Barchan west of the Salton Sea, California, USA (Photo courtesy K. Mulligan.)

4. Development of impeded dunes

A variety of dune forms are related to vegetation, topographic barriers or localized sources of sediment.

5. Fine-grained deposits

Silts and especially clays can be transported considerable distances by the wind. Loess, consisting of well-sorted, very fine-grained deposits, cover large areas. Unusually thick deposits known as **loess-lips**, are found downwind of major river valleys which provided major sources of silt during glacial phases. A high degree of sorting, angularity of constituent particles, and small particle size associated with relatively strong interparticle bonds and moisture-retention capability makes loess a relatively cohesive deposit which in some cases forms steep cliffs.

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Prepared by
Dr. Rami M. Idan
Petroleum Geologist

Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

<u>Global morphology and tectonics</u>

"Synopsis"

Lecture EIGHT

Prepared by

Dr. Rami M. Idan

Prepared by
Dr. Rami M. Idan
Petroleum Geologist

1- Global morphology

The Geoid

The Earth is only approximately a sphere. As a result of the centrifugal force of rotation it bulges at the equator and its polar radius (6378 km) is 21 km shorter than its equatorial radius (6397 km); thus the Earth is more accurately described as an ***oblate spheroid***. Even this description, however, is not completely accurate as inhomogeneities in the distribution of mass in the Earth's interior produce further small but measurable irregularities on its surface. The shape of the Earth determined in this way is known as the ***geoid*** and is represented by the surface defined by mean sea level over the oceans and the extension of sea level along imaginary canals across the continents. The significance of the geoid for the operation of geomorphic processes is that it defines the ultimate base level for denudation; any change in the geoid would therefore cause consequential changes in base levels. This is an important issue in studies of sea-level change and is examined further in “Stratigraphy Course”.

Major morphological features

Beginning with the more familiar form of the continents we can distinguish between the continental platforms formed by *plateaus* and *lowlands*, and the major linear mountain systems known as ***orogenic mountain belts***, or simply ***orogens.***

*(Orogenesis is the most complex of tectonic processes and interpreting ancient mountain belts. As originally defined, an **orogeny is simply a period of mountain building.** To field geologists the term **orogeny represents a penetrative deformation of the Earth's crust associated with phases of metamorphism and igneous activity along restricted, commonly linear zones and within a limited time interval**. It is eventually a mountain building event, which results from compression of continental crust and addition of igneous material to it.*

Orogen: *A body of rocks affected by mountain building.*

Orogenesis: *The process of mountain building.*

One major mountain system is **Alpine Orogeny,** The position of the continents by 50 million years ago looked quite similar to that of today. This orogeny occurred mainly between 65 and 2.5 million years ago, although it is still active today. It saw the collision of the African and Eurasian plates, and the closure of the Tethys Ocean as oceanic lithosphere was subducted northwards beneath the Eurasian Plate, leaving today what we now know as the Mediterranean Sea.

2- Earth structure

Seismic evidence

❶ Primary, or **P-waves,** are the fastest. They are compressional waves, the energy being transmitted by an initial compression of particles which is then passed on to adjacent particles. This produces a sequence of zones of compression and expansion which travel away from the source.

❷ Secondary, or **S-waves,** are shear waves and they transmit energy by an '*up and down*' motion of particles normal to the direction of propagation, The density and elastic properties of rock are both important in affecting the passage of seismic waves, P-waves can be transmitted through any material, but S-waves can only be transmitted through solids.

Crustal structure

Seismic and gravity data, together with direct evidence from the rocks themselves, allow us to identify the structural and compositional differences between oceanic and continental crust. Oceanic crust has a mean density of about 3000 kg m-3 and is composed of layers of basic rocks, broadly basaltic and gabbroic in composition, with a thin veneer (or layer) of sediments. Over most of the ocean floor this sedimentary cover is only 1-2 km thick and in the vicinity of mid-oceanic ridges it becomes very thin indeed. Most of these regions are also topographically distinct, but some, such as the **continental shield**, or **craton**, and mid-continent types, are primarily differentiated on the basis of their detailed structure and the thickness of their sedimentary cover.

Development of ideas on global tectonics: *Plate Tectonic Theory as shown in Physical Geology Course.*

: Reference
Summerfield, M. A., 2013, Global Geomorphology. An Introduction to the Study of landforms-Routledge, London and New York, 537 pp

http://sp.lyellcollection.org/content/specpubgsl/121/1/1.full.pdf

youtube at

https://www.youtube.com/watch?v=kT6RxtWDeh0
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Prepared by
Dr. Rami M. Idan
Petroleum Geologist