Chapter 2

Introduction to Sedimentology

Sediment Transport and Sedimentary Environments

Knut Bjørlykke

Sedimentology is the study of sedimentary rocks and their formation. The subject covers processes which produce sediments, such as weathering and erosion, transport and deposition by water or air, and also the changes which take place in sediments after their deposition (diagenesis). Diagenesis includes mechanical sediment compaction and mineral reactions as a function of temperature during burial. Changes in sedimentary rocks at temperatures of over 200–250°C are called *metamorphic processes* and are not dealt with here. In this chapter we shall discuss primarily transport and deposition of clastic sediments and sedimentary environments. These processes determine the distribution and geometry of reservoir rocks in a sedimentary basin and also the changes in rock properties during burial. Accumulation of organic-rich sediments which may become source rocks is also an integral part of sedimentological models. Since both source rocks and reservoir rocks normally are sedimentary rocks, sedimentology is a very important subject for petroleum exploration and production. Sedimentological research is to large extent funded by the petroleum industry.

Like all natural sciences, sedimentology has an important descriptive component. In order to be able to describe sedimentary rocks, or to understand such descriptions, it is necessary to familiarise oneself with quite an extensive nomenclature. There are specialised names for types of sedimentary structures, grain-size distributions and mineralogical composition of sediments. We also have a genetic nomenclature, which names rock types according to the particular way in which we think they have formed. Examples of these are fluvial sediments (which are deposited by rivers) and aeolian (air-borne) sediments. The descriptive nomenclature is used as a basis for an interpretation of how the rock was formed. When we are reasonably confident about their origin we may use the genetic nomenclature.

Sedimentology covers studies of both *recent* (modern) sediments and older sedimentary rocks. By studying how sediments form today we can understand the conditions under which various sedimentological processes take place. From such observations we may be able to recognise older sediments which have been formed in the same way. This is called using the principle of uniformitarianism, which has been of great importance in all geological disciplines since its proposal by James Hutton (1726–1797).

Conditions on the surface of the Earth have fluctuated widely throughout geological history and the principle of uniformitarianism cannot be applied without reservation. One important aspect of sedimentological research is attempting to reconstruct changes in environments on the Earth's surface throughout geological time. This applies particularly to climate, vegetation and the composition of the atmosphere and the oceans.

Palaeontology is important to sedimentology, not only for dating beds, but also because organisms are an important component of many sedimentary rocks (particularly limestones), and organic processes contribute to weathering processes and the precipitation of dissolved ions in seawater. Many organisms make very specific demands of their environment, and

K. Bjørlykke (✉)
Department of Geosciences, University of Oslo, Oslo, Norway
e-mail: knut.bjorlykke@geo.uio.no

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fossils are consequently a great help in reconstructing the environment in which the sediments were deposited. Palaeoecology is the study of ecological conditions as we are able to reconstruct them on the basis of remains or traces of plants and animals in rocks. Traces of animals in sediments have proved to be very useful environmental indicators.

Studies of recent and older sedimentary rocks provide a fruitful two-way exchange of information in sedimentology. From studies of recent environments we can learn about the conditions that particular processes require. In older rocks, however, we can study sedimentary sections which encompass many millions of years of sedimentation, offering us a completely different record of the way sedimentological processes can vary as a function of geologic time. As a result, studies of older rocks also contribute to our understanding of the recent environment and offer non-uniformitarian explanations.

When we study rocks, we should attempt to give *objective descriptions* of the composition, structure etc. of the rocks, and on the basis of these try to *interpret* how they were formed. However, it is impossible to give a completely exhaustive, objective description of a rock. Nevertheless it is often most fruitful to have a theory or hypothesis against which to test our observations. Data collection can then be focused on observations and measurements which can support or disprove the hypothesis.

We know from experience that we have a tendency to observe what we are looking for, or what we anticipate finding.

Early descriptions of sedimentary sequences contain few observations about sedimentary structures which we would consider fairly conspicuous and important today. We observe sedimentary structures because we have learned to recognise them and understand their genetic significance.

Many sedimentologists use a standard checklist for what they should observe in the field, so that their descriptions are as comparable as possible. Nevertheless, it is important that field observations do not become too much of a routine. Facies analysis should be a creative process and the various depositional models should be kept in mind when making the observations. It is also desirable to quantify field observations as far as possible, for example by surveys in the field and a range of laboratory analyses. These might be texture analyses (e.g. grain-size distribution), microscope analyses (perhaps using a scanning electron microscope) or chemical analyses. Pure descriptions of sedimentary rocks are useful because they increase the data base on which we can build our interpretations.

Systematic studies of recent environments of sedimentation are important to find connections between the environment and the sediments which accumulate. The environment governs the sedimentological processes which determine what sort of sediments are formed and deposited. The connection we are trying to understand in modern environments is thus: environment → process → sediment.

Today a large number of modern environments of sedimentation have been studied in great detail. These include aeolian and fluvial environments, deltas, beach zones, tidal flats and carbonate banks. Deep sea environments have naturally not been so easy to study, but modern sampling and remote sensing equipment and underwater TV cameras have made it easier to gather observations from this environment, too. In recent years systematic drilling through sedimentary layers on the ocean floor (Deep Sea Drilling Program – DSDP, and Ocean Drilling Program – ODP) has provided an entirely new picture of the geology of the ocean depths. Specially constructed diving ships (e.g. ALVIN) make it possible for geologists to observe the ocean floor directly at depths of up to about 3,500 m and take samples of surface sediments. In addition geophysical, particularly seismic, surveys provide one of the most important bases for understanding the stratigraphy and geometry of sedimentary basins.

In studying older rocks we base our approach on certain features which we can observe or measure, and attempt to interpret the processes that produced them. Particular variations in grain size and sedimentary structures in profiles can be interpreted as having been formed through particular processes, e.g. aeolian, tidal or deltaic processes. The recognition of such sedimentary processes helps us to reconstruct the environments. The sequence of interpretation in studies of older sedimentary rocks is thus: description of sedimentary rock → processes → environment.

Applied geology has always been important, even for purely scientific research. The interests of economic recovery of raw materials from sedimentary rocks create a demand for sedimentologists and sedimentological research. Exploration for and recovery of raw materials also provide important scientific information. Sedimentary rocks contain raw materials of considerably greater value than those we find in metamorphic and eruptive rocks. The most important are oil, gas and coal deposits, but a very large amount of the world's ore deposits is also found in sedimentary rocks and many types of ore have been formed through sedimentary processes. Limestone, clay, sand and gravel are also important raw materials which require sedimentological expertise.

The petroleum industry employs a very high percentage of the world's professional geologists. This industry has a particular need for research, and also has the financial capacity to invest in it. Because oil and gas are found largely in sedimentary rocks, exploration for and recovery of hydrocarbons is based to a large extent on sedimentology. Much of what we now know about the world's sedimentary basins and their regional geology is derived from seismic profiles which have been shot in connection with oil exploration and drilling for oil and gas. The oil industry has also helped to stimulate pure sedimentological research, and significant contributions to research in this area are published by the research laboratories of the oil companies. Research based on economic interests is also useful from a purely scientific point of view, because it often focuses on particular questions which may be quite fundamental.

Petroleum geology requires close teamwork between reservoir engineers and geologists, to establish in great detail the geometry and distribution of porosity and permeability in reservoir rocks. We also need very much to know more about the physical and chemical properties of reservoir rocks, for reasons which are discussed at the end of the book. Geophysical methods provide most of the information used in petroleum exploration and production and many petroleum geologists rarely examine real rocks in cores and cuttings. It is however important to know something about the textural and mineralogical composition of the sedimentary sequences. The geophysical data rarely provide unique solutions when inverting seismic and log data to rock properties. Facies models provide ideas about how different lithologies may be distributed in a sedimentary basin. With few wells far between it is difficult to make a realistic representation in 3D. The diagenetic processes which change rock properties after deposition are equally important. Sedimentary facies and the primary sediment composition represent however the starting material for burial diagenetic reactions determining the rock properties.

The depositional environment determines to a large extent the shape and extension of potential reservoir rocks such as sandstones and limestones. Sequence stratigraphy of the sequences filling a sedimentary basin often provides the key to predict the distribution of source rocks, migration pathways and traps. The internal properties of reservoir rocks such as porosity, density and velocity depend however very much on the mineralogical and textural composition at the time of deposition.

Sandstones may vary greatly with respect to grain size, feldspar and clay content and compact very differently. The source of clastic minerals (provenance) is an important part of basin analyses which will help to predict rock properties at greater burial depth.

Mudstones and shales should also be analysed with respect to grain size (silt content) and clay mineralogy because different types of clay and mud are important parts of sedimentary sequences, particularly as source rocks and cap rocks.

Limestone rocks are to a very large extent composed of material precipitated biologically and the types of marine organisms forming reefs and other carbonate reservoir rocks reflect to a large extent the biological evolution through time.

Also in the clastic sediments like sandstones and shales biologically precipitated carbonate and also silica are important components controlling rock properties. Quartz-rich sandstones may have very little porosity because of carbonate cement and calcareous mudstones and shales may be very tight and brittle. This also influences the seismic data recorded in sedimentary basins.

Sedimentology and facies analyses represent an important basis for the different petroleum related disciplines discussed in the following chapters in this book.

2.1 Description of Sedimentary Rocks

2.1.1 Sediment Composition. Mineralogy and Other Components

Clastic sediments consist of grains and fine particles which may be composed of single minerals, rock fragments and also some amorphous compounds such as organic matter, biogenic silica and volcanic glass. The sediment composition is a function of the rocks being eroded or weathered (provenance), sediment transport, and depositional environments. The grains-size distribution is very much a function of the mineralogy in the source rock, and shales and volcanic rocks will produce clay-rich sediments.

Rapid erosion of granite and gneisses produces feldspar-rich sand (arkoses) while weathering results in sand corresponding to the quartz crystals in the granite and clay composed of kaolinite from weathering of feldspar. Volcanic rocks produce smectite-rich clays which are very fine grained and have cohesive properties very different from kaolinitic clays. The grain-size distribution and the sand/shale ratio as well as the composition of clay play an important role in the depositional processes.

Descriptions of outcrops and cores should not be limited to studies of lithology and sedimentary structures but should also include sampling and analyses of the textural and mineral composition by microscope (optical microscope and SEM) and by X-ray diffraction (XRD). This will provide a basis for the interpretation of diagenetic reactions occurring during progressive burial in a sedimentary basin and also the distribution of pore space between the minerals.

2.1.2 Textures

The *textures* of clastic sediments include external characteristics of sediment grains, such as size, shape and orientation. These properties can be described relatively objectively and say a great deal about the origin and conditions of sediment transport and deposition.

By grain size we normally mean grain diameter, but the two are only strictly synonymous in the case of completely spherical particles. Most grains are not spherical, however, and it is difficult to identify a representative diameter, particularly in the case of elongated or flat grains. For this reason we have adopted the concept "nominal" diameter (d_n), defined as the diameter of a spherical body which has the same volume as the grain. In practice we are seldom in a position to measure the volume of individual grains, and we therefore use indirect methods to measure the distribution of grain size within a sample.

Sand and gravel can most simply be analysed by means of *mechanical sieving*. A bank of sieves consists of sieves with mesh sizes which decrease downwards. A sample is put in the uppermost sieve and the bank of sieves is shaken (Fig. 2.1a). Grains which are larger than the mesh size will remain, while smaller grains will fall through and perhaps remain lying on the next sieve. By weighing the fraction of the sample which remains on each sieve, we can construct a grain-size distribution curve. The lower practical limit for sieve analyses is 0.04–0.03 mm; finer particles exhibit much more cohesion, which makes it difficult for them to become separated and pass through the finer sieves.

Fine silt and clay fractions can be analysed in a number of ways. Most classic methods are based on measurements of settling velocity in liquids, and are based on Stokes' Law:

$$v = cgR^2\Delta\rho/\mu$$

Here c is a constant (2/9) and μ is the viscosity of the water. R is the radius (cm) of the grain and $\Delta\rho$ is the density difference between the grain and the fluid (water).

When the settling velocity of grains (falling through water, for example) is constant, the resistance to the movement (friction), which acts upwards, must be equal to the force of gravity, which acts downwards (Fig. 2.2).

$$6\pi Rv\mu(\text{friction}) = 4/3\pi gR^3\Delta\rho(\text{gravity})$$

$$v = c\,gR^2\Delta\rho/\mu$$

$$\text{Log}\,v = 2\log R + c(\text{a constant})$$

The settling velocity is sensitive to temperature variations, which affect the viscosity of the water (μ).

We can measure the settling velocities of sediment grains indirectly by measuring the density of the water with suspended sediment sample with a hydrometer, which registers the fluid density. We disperse the sample in a cylinder with a mixer so that at a start time T_o we have an even distribution of all grain sizes, and therefore of density, throughout the cylinder. The individual sediment grains then sink to the bottom at a rate which is a function of their size. The change in the fluid density as progressively fewer grains remain in suspension is therefore a function of the grain-size distribution. This applies for small particles, where the flow of the liquid around the grain is laminar and the concentration of grains is low.

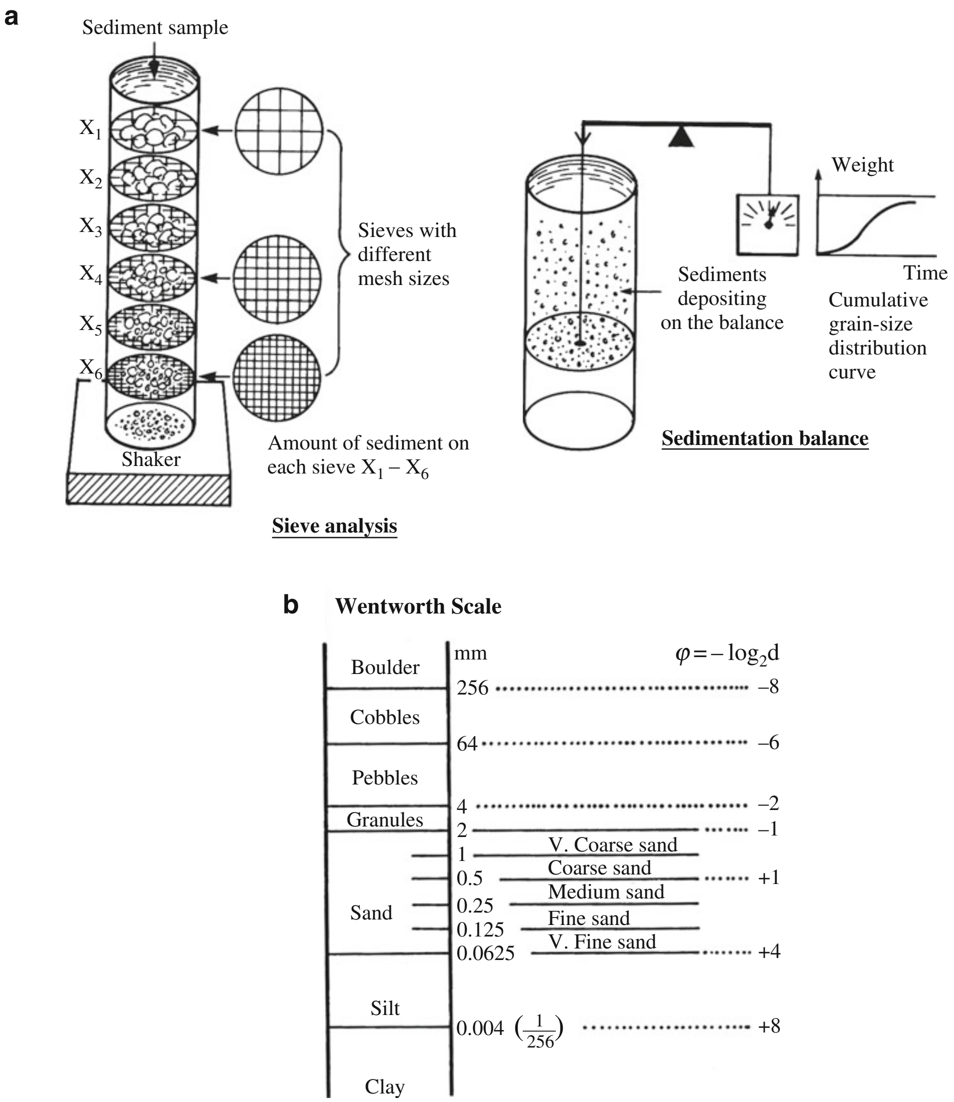


Fig. 2.1 (**a**) Sketch showing the principles involved in sieve analysis and use of a sedimentation balance. Sieve analysis is usually used for grain sizes down to 0.03–0.02 mm, but with wet-sieving even finer sediment grains can be sieved. The sedimentation balance gives us a direct expression of settling velocity, i.e. weight increase as a function of time. This is therefore a cumulative grain-size distribution. (**b**) Grain-size classification of clastic sediments. The grain size (d) is often described in terms of φ values ($\varphi = -\log_2 d$)

When the grains are larger than about 0.1–0.5 mm, the settling velocity increases, and turbulence develops around the grains. The frictional resistance therefore increases, and in the case of larger grains (>1 mm) the settling velocity increases approximately in proportion to the square root of the radius. It is not practicable to measure each grain, but the settling velocity can be measured indirectly. A hydrometer floating in a suspension of sediments and water measures the density of the suspension through time. The rate of density reduction in the suspension is a function of the grains' size. By taking successive readings of the density we may plot a curve which expresses density reduction as a function of time. Since density reduction is a function of settling velocity, this curve can be calibrated to give a grain-size distribution curve.

When we analyse fine-grained sediments with a large clay fraction, or separate out clay fractions, it may be useful to use a centrifuge. We then increase the

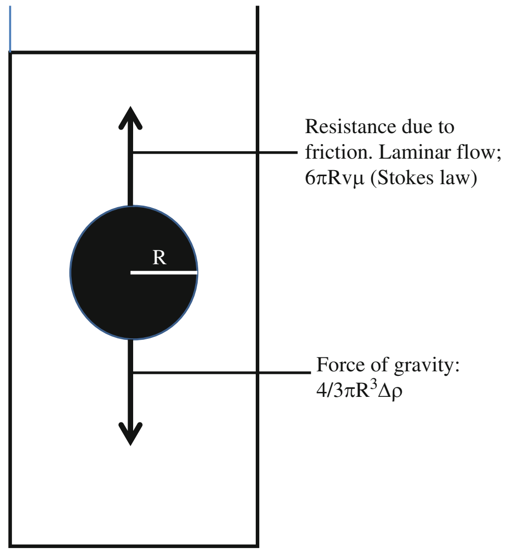


Fig. 2.2 The velocity of a falling grain in water is controlled by the gravity forces directed *downwards* and the resistance to the flow around the grain which is directed *upwards* ($6\pi Rv\mu$). The force of gravity is a function of the volume of the grain ($4/3\pi gR^3$) and the density difference ($\Delta\rho$) between the grain and the fluid ($\rho_g - \rho_f$)

acceleration term **g** in Stokes' Law. There are also "sedimentation balances". Sediments suspended in a cylinder fall through the water column and accumulate on a balance pan at the bottom of the cylinder. This balance records and writes out the increase in weight, which is the precipitation from suspension, as a function of time. This gives a direct cumulative curve.

Other methods are based on the refraction or dispersion of a laser beam passed through suspensions producing a characteristic "scatter" which is calibrated against samples of known grain size. These machines use very small samples and have a high degree of repeatability. Equipment has also been developed which uses X-rays instead of light to produce the characteristic scatter patterns.

It is important to note that *no* method measures the nominal diameter. In methods which measure settling velocity, grain shape is a significant factor. A large, thin mica flake has a settling velocity which corresponds to that of a considerably smaller spherical grain. The diameter of a spherical grain with the same volume and settling velocity is called the *effective diameter* (d_e). With the scatter method, flaky grains are assigned a different, probably greater, diameter than that indicated by the settling velocity method.

2.2 Grain-Size Distribution in Solid Rocks

Lightly-cemented sandstones can be disintegrated by means of ultrasound in the laboratory and then analysed as loose sediment in the normal manner. Carbonate-cemented rocks may be disintegrated using acids. However, we must bear in mind that new clay minerals may have formed through post-depositional alteration (diagenetic processes), and that some of the original minerals may have been broken down mechanically or dissolved chemically. Consequently it is not certain that we are dealing with the original grain distribution. Diagenesis must be taken into account.

Well-cemented rocks must be analysed in thin section by means of a petrographic microscope. It is difficult to analyse the finer fractions (fine silt and clay) in this manner and we must always remember the "section effect", i.e. that in most cases we will not be seeing the greatest diameter of the grains. With spherical grains, the relation between the real diameter, d_r, and the observed diameter, d_o, can be expressed statistically: $d_r = 4d_o/\pi$.

2.3 Presentation of Grain-Size Distribution Data

Grain-size distribution is one of the many types of natural data which must be presented on a logarithmic scale for convenience. Wentworth's scale is based on logarithms to the base 2, and this is now the one most widely found in geological literature (Fig. 2.1b).

For the sake of convenience, these data are commonly plotted against a linear scale. The phi (φ) scale, where $\varphi = -\log_2 d$, allows convenient interpolation of graphic data. The reason this negative logarithm is used is that normally most of the sediment grain diameters (d) are less than 1 mm, so these will have a positive phi value (Fig. 2.3). It is convenient to plot grain-size distribution data as a function of phi values, especially on cumulative curves. In normal descriptions of grain size, however, it is more helpful to state grain size in mm, so the reader does not have to calculate back from phi values.

The simplest, and visually most informative, way of presenting grain-size distribution data is by means of histograms (Fig. 2.3). These show the percentage, by weight, of the grains falling within each chosen

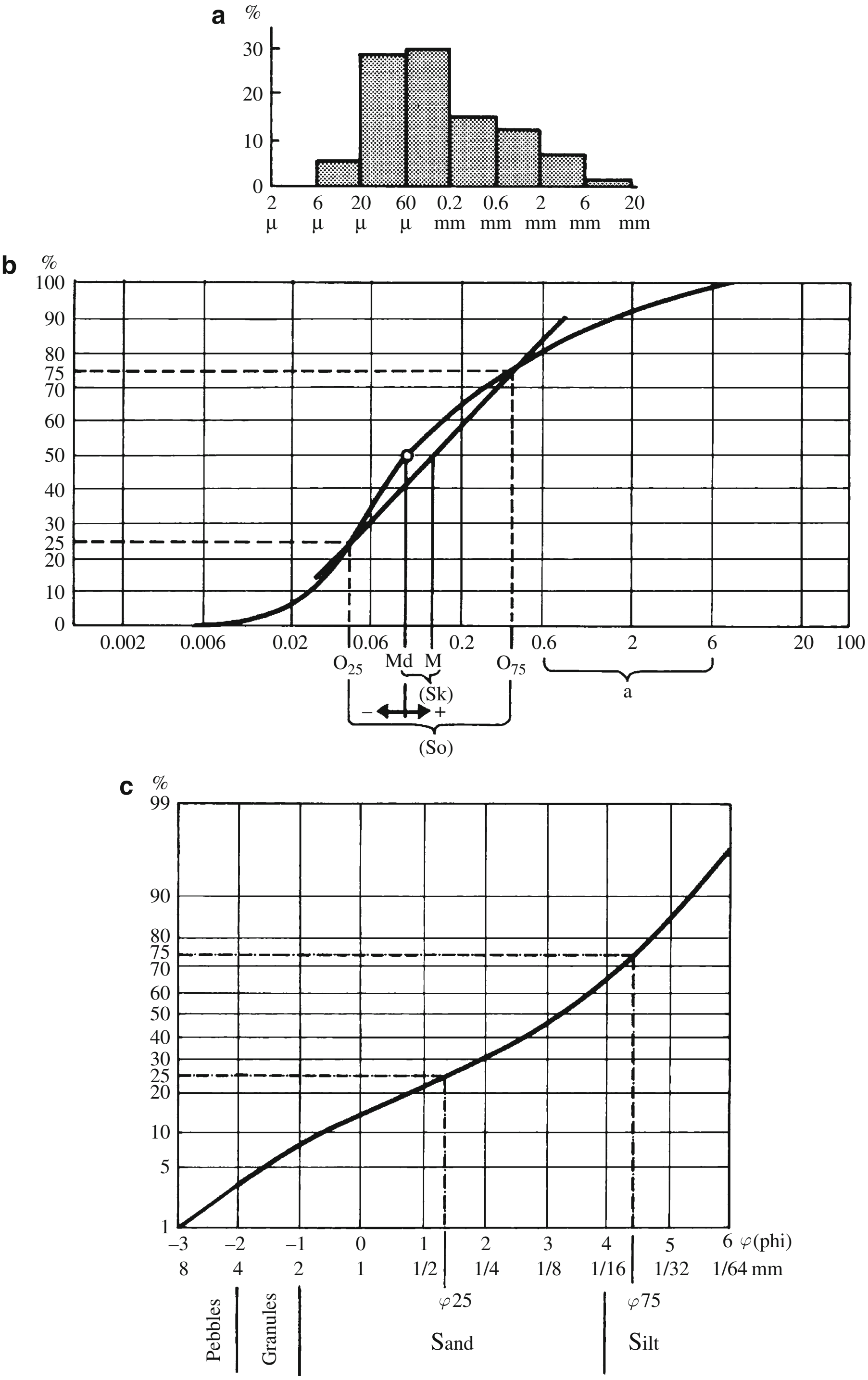


Fig. 2.3 Presentation of the same grain-size distribution data as (**a**) histogram, (**b**) cumulative curve, and (**c**) cumulative curve on probability paper. When plotted on probability paper, a logarithmic normal distribution, like a Gaussian curve, will plot as a *straight line*

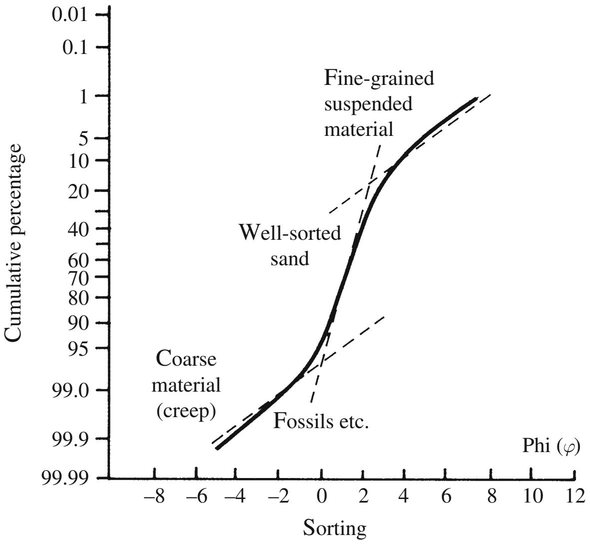


Fig. 2.4 Grain-size distribution curve presented as a function of grain size in φ values against a logarithmic cumulative percentage

subdivision of the size range. It is then easy to see how well sorted the sediments are, and whether the distribution of grain sizes is symmetrical, or perhaps bi- or polymodal, i.e. with two, or more, maxima.

A *cumulative* distribution curve shows what percentage of a sample is larger or smaller than a particular grain size. The steeper the curve, the better the sorting. Note that engineers use the inverse term "grading", whereby well graded = poorly sorted.

If we use probability paper, distributions which are lognormal (following a logarithmic distribution) will plot as straight lines and the slopes of these will reveal the degree of sorting. Even if the whole distribution is not lognormal, it often appears that the curve can be regarded as a composite of 2–3 lognormal grain-size populations. These populations generally overlap, so that some sections of the curve represent a combination of parts of two populations, each of which may be lognormal. Each population may represent a different mode of grain transport, for example saltation, rolling (bedload) or suspension (Fig. 2.4).

It is important that we collect representative samples for grain-size distribution analysis, i.e. each sample only has material from one bed. This ensures that it represents deposition by a single sedimentary process. If we take a sample at the boundary between two beds, we will often get false bimodal distributions which can easily be mistaken for naturally produced bimodal sediments, leading to interpretation errors.

2.4 Grain-Size Distribution Parameters

Phi (φ) $= -\log_2 d$ (after Folk and Ward 1957) where d is the grain diameter in millimetres (as previously defined). The percentage of grains larger than a certain grain size (φ) is called the *percentile*. φ30 means that 30% of the grain population by weight is larger than the grain size. For $\varphi = 4$ the grain size is 0.0625 mm so that 30% of the sample is sand or larger grains.

2.5 Significance of Grain-Size Parameters

The *mean* diameter is an arithmetically calculated average grain size. The *median* diameter is defined by the grain size where 50% by weight of the sample grains are smaller, and 50% are larger. Only in the case of completely symmetrical distribution curves will the mean diameter (M) and the median diameter (M*d*) coincide. The mean will otherwise shift further than the median in the direction of the "tail" of the distribution. If the sample has a wide spread (tail) towards the fine grain sizes (larger phi values) and a relatively sharp delimitation at the large grain-size end, we say that the sample has positive skewness. This will be typical of fluvial sediments. There will be a fairly definite upper limit to the grain sizes that rivers can transport as bedload, while there will be no sorting of the fine fractions which are transported in suspension. Major variations in flow velocity, for instance during floods, will give poorer sorting.

Aeolian (wind) deposits are very well sorted (Fig. 2.5). They also have positive skewness because there is an upper size limit to the grains which can be transported. Although the finest particles may be removed selectively, there will still be a "tail" of fine material. The fine material in dunes may also be protected by a cover of larger particles *(lag)* against further erosion and transport. Beach sand deposits, on the other hand, are clearly negatively skewed, i.e. the distribution curve shows a definite lower limit, while there is often a "tail" of larger particles, i.e. granules and pebbles. The hydrodynamic conditions on a beach are such that each wave brings some sediment in suspension. Whereas sand grains, particularly medium to coarse sand, will rapidly settle from suspension and be deposited on the beach again, fine sand, silt and clay will remain in suspension longer. This finer material will be transported further out and at a depth of

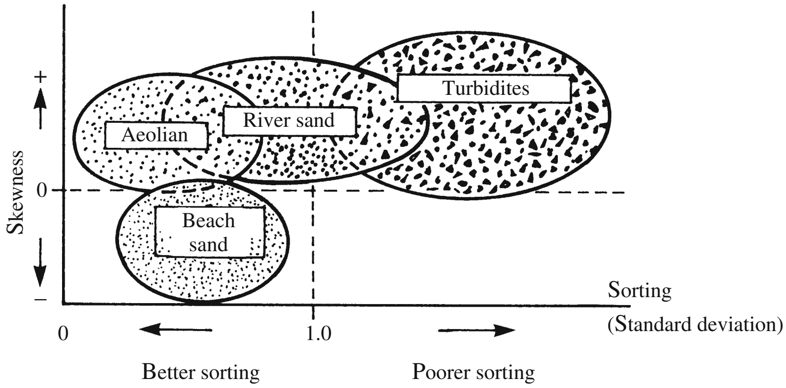


Fig. 2.5 Sorting and skewness are grain-size distribution parameters which are suitable for distinguishing between sediments deposited through various processes and in different environments of deposition. Turbidites and fluvial sediments have a tail of small sediment particles while beach sand may have coarser grains such as pebbles

some metres (1–50 m), depending on how strong the waves are (and consequently the depth of the wave base), we will have poorer sorted deposits because the fine fractions will be deposited and mixed with coarser-grained sediments which are carried out during storms.

Sediments deposited from suspension have poor sorting and positive skewness. This is very typical of turbidites. There are however some turbidites that are composed of relatively well sorted sand. They may represent beach or shelf sand which has been reworked and deposited on a submarine slope. Clay suspensions which are deposited on land as high-density suspensions (mud flows), have negative skewness because they often contain large clasts.

Kurtosis is an expression for the spread of the extreme ends of a grain-size distribution curve in relation to the central part. This distribution parameter is used somewhat less than the others, but relatively small amounts of silt and clay can have a significant effect on the properties of coarse-grained sand. Pebbles and stones in otherwise fine-grained sediment may also be important.

One of the most important points to bear in mind when interpreting grain-size distributions is the availability of grain sizes supplied to the area where the process is taking place. Strong currents or wave energy will only be able to deposit coarse-grained sediments if there is coarse-grained sediment present in the area. A source area where sediments are generated by erosion and weathering will often supply specific grain sizes. Chemical weathering of acid rocks (granites), for example, will lead to the formation of sediments consisting of quartz grains corresponding in size to the quartz crystals in the granite, and clay (kaolinite, smectite, illite) formed through weathering of the feldspars. Weathering of basic rocks (basalts and gabbros) will produce almost exclusively clay minerals, and practically no sand grains.

The grain-size distribution in a depositional area also depends on the transport mechanisms carrying sediment to the area. The grain-size distribution we observe in a sediment therefore reflects both the hydrodynamic conditions and the grain-size population of sediments available from the source area. Furthermore, although a particular grain-size distribution may be characteristic for a type of deposit, it does not point unambiguously to a particular environment of deposition because similar hydrodynamic conditions can exist in different environments. Statistical comparison of high resolution (minimum ½ Phi sampling interval) grain-size distributions, applying pairs of Folk and Ward or Krumbein parameters, can provide some diagnostic criteria for hydrodynamic interpretations.

Sediments may change their grain-size distribution by diagenetic processes at quite shallow burial as well as at greater depth. The formation of the clay minerals kaolinite and smectite from feldspar and rock fragments (volcanics) in a sediment certainly results in very different grain-size distributions from those at the time of deposition.

2.6 Grain Shape

We distinguish between three parameters:

1. *Roundness* is a property of surface shape – whether it is smooth or angular. A visual scale is most commonly used.
2. *Sphericity* is an expression for how much a particle deviates from a spherical form, and is defined as the ratio between the diameter of a circumscribed circle round the grain and the diameter of a sphere of the same volume (the *nominal diameter)*.
3. We also use various expressions for grain shape such as (a) *discoid* or *bladed* for grains which are flat, (b) *Prolate* or *roller* for grains with one dimension considerably greater than the two others, (c) *equant* for grains with three relatively equal dimensions and (d) *oblate* for grains with one large, one medium and one small dimension.
4. *Surface textures* are concerned with the nature of the surface itself, whether it is rough, smooth, pitted, scratched etc. Some textures are diagnostic of specific modes of transport, and superimposed texture features may reveal the transport history of a grain. The surface texture of grains can best be studied under the scanning electron microscope. Aeolian sand grains may develop fine pitting on their surfaces due to the collisions of grains during transport, clearly visible under a binocular microscope.

Large grains become rounded far more rapidly than smaller ones because the impact energy released in collisions with other grains declines in proportion to the cube root of the radius. Blocks may be rounded after only a few hundred metres or several kilometres of transport. Grains less than 0.1 mm in diameter undergo little rounding even when carried very long distances in water, for example by tidal currents.

The grain size and sorting of sand grains at the time of deposition play an important role determining the rate of compaction with increasing overburden stress (See chapters 4 and 6). The grain-size distribution may also change due to grain crushing during compaction and chemical alterations e.g. dissolution of felspar and precipitation of clay minerals like kaolinite.

2.7 Sediment Transport

Sedimentary grains can be transported by water or by air. In order to understand the transportation processes we must know a little about the hydrodynamic (or aerodynamic) principles involved. When a liquid or gas flows in a channel or pipe it exerts a force (shear stress) against the walls or bottom. This force is counteracted by friction from the walls.

Pure water without suspended sediment is a Newtonian fluid which obeys Newton's law:

$$\tau = \mu\, dv/dh$$

A Newtonian fluid has no shear strength, so it will be deformed even by an infinitely small shear stress (dv/dh).

$\tau =$ *shear stress*, which is an expression of force per unit area (N/m^2). μ is the dynamic viscosity expressed in poise (g/cm/s or 0.1 N s/m^2), dv/dh is the change in velocity (dv) or velocity gradient (deformation velocity) as a function of distance from the boundary (dh). The viscosity of pure water decreases with increasing temperature. Suspended material may also affect viscosity, but the concentration of suspended material must be quite high (15–25%) before the viscosity increases significantly. If the water contains a large percentage of swelling clay minerals (smectite), however, the viscosity will increase at lower concentrations. The kinematic viscosity v is the dynamic viscosity (μ) divided by density ρ, i.e. $v = \mu/\rho$ and units are cm^2/s.

We distinguish between laminar flow, where each point in the liquid moves along a straight line parallel to the bed, and turbulent flow, where each point follows an irregular path so that eddies form (Fig. 2.6). Reynold's number (Re) is a dimensionless number which describes flow in channels and pipes. It is defined as:

$$\mathrm{Re} = vh\rho/\mu$$

Here v is the mean velocity, h is the depth of a channel or the diameter of a pipe in which fluid is flowing, ρ is the fluid density and μ its viscosity. If Reynold's number exceeds a certain value, about 2,000, the flow changes from laminar to turbulent. The density of water is 1 g/cm^3 and the viscosity is one centipoise (0.01 Ps = 0.01 g/cm/s). We see that the boundary between laminar and turbulent flow corresponds to 20 cm/s.

This means that for the flow of water to be laminar the product of velocity (cm/s) and depth (cm) must not exceed 20. If the velocity is 1 cm/s, there will be turbulence if the depth (h) is greater than 20 cm. In practice, then, flow in rivers and channels is always

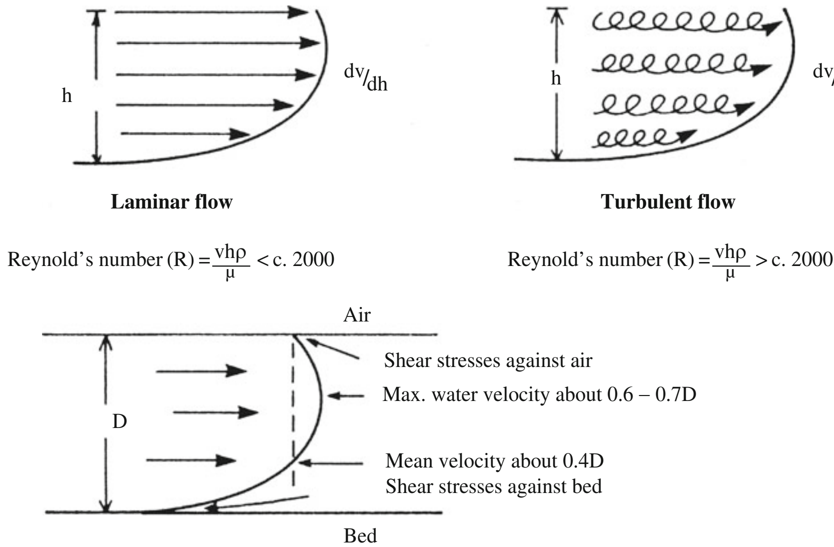


Fig. 2.6 Diagram showing principles of turbulent and laminar flow and the shear stress against the underlying bed

turbulent. For turbulent flow the expression for shear stresses which applies to laminar flow, is no longer adequate. The shear stress in turbulent flow will then increase as a function of velocity because of the eddies which produce an *eddy viscosity* (η) (Fig. 2.6). The total shear stresses will then be: $\tau = (\eta + \mu)/dv/dh$.

2.8 Flow in Rivers and Channels

For all types of water flow the forces acting on the water must be in equilibrium. In most cases it is the force of gravity which balances bed frictional forces. In order to understand geological processes in connection with the erosion, transport and deposition of sediments, it is important for us to be aware of the relationships which govern the flow of water in channels.

If the channel has a cross-section A and we look at a stretch L of the channel, the force of gravity will be:

$$F_1 = \rho \cdot g \cdot L \cdot A \cdot \sin\alpha,$$

where ρ is the density of water, g is the force of gravity (constant) and α is the angle of slope of the channel. The resistance to flow consists of frictional forces against the bed and against the air. If we disregard friction against the air, the frictional forces are:

$$F_2 = \tau \cdot L \cdot P$$

where $\tau =$ shear stress (force per unit area) and $L \cdot P$ is the area of the bed on which the forces are acting. P is the wet perimeter and L is the length of a line along the bed in a section along the channel. If the water flow has a steady velocity, the force of gravity F_1 will just equal the frictional force F_2 (Fig. 2.7). Consequently:

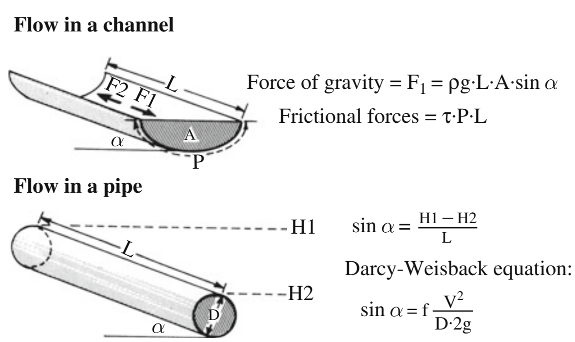


Fig. 2.7 Flow of water in channels is controlled by the ratio between the gravitational forces and the shear stress against the bottom of the channel

$$\tau \cdot L \cdot P = \rho g \cdot L \cdot A \cdot \sin\alpha$$

or

$$\tau = \rho \cdot g \cdot \frac{A}{P} \cdot \sin\alpha$$

A/P is the cross-section of the channel divided by the wet perimeter, and we call this the hydraulic radius, R. For flow in a pipe:

$$R = D/4$$

The shear stresses (τ) $R = D/4$ increase in proportion to the square of the velocity ($\tau = c \cdot v^2$).

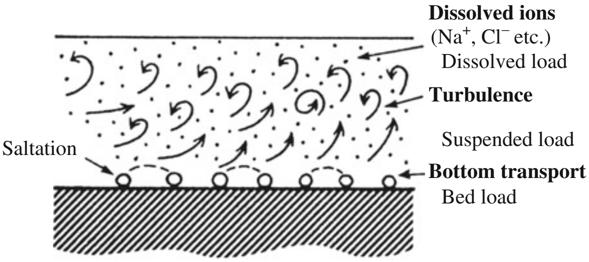


Fig. 2.8 Different forms of transport in water. Sediment grains may be carried in suspension if the vertical component of the turbulence is equal to the falling velocity of the grains. Larger grains are carried along the bottom due to the shear stress

This relation between shear stress and flow velocity can also be used for flow in channels where we have bedload transport (Fig. 2.8). Solving the two equations above with respect to the velocity (v) we obtain:

$$v = C(R \sin\alpha)^{1/2}$$

This is the Chezy equation and C is the Chezy number.

The value of C depends on the roughness of the bed and on the *shape* of the channel, particularly its *sinuosity*.

Often used in engineering for calculating the velocity of water in channels, is Manning's formula:

$$v = R^{2/3} \cdot (\sin\alpha)^{1/2}/\eta$$

where n is the coefficient of roughness of the bed: $n = 0.01$ corresponds to a smooth metal plate and $n = 0.06$ to a shifting bed of gravel. It is of great practical importance to be able to calculate water velocity and thereby the erosion potential of artificial channels.

The Froude number is a parameter which is often used to describe water flow:

$$F = v/(g \cdot h)^{1/2}$$

where v is the average velocity, h is depth of water and g the force of gravity. The Froude number is the ratio between the kinetic energy of the water masses (which is proportional to the square of the velocity) and the force of gravity, which is proportional to the depth, h. For low Froude numbers the water flows out of phase with the bedforms, and current ripples or cross-bedding develop. This is called the lower flow regime. When the velocity, v, becomes high in relation to the depth of water, h, rapid or shooting flow develops where the waves come into phase with the boundary irregularities (Fig. 2.9); this represents the upper flow regime.

The transition between lower and upper flow regimes corresponds to a Froude number of 0.6–0.8.

2.9 Sediment Transport Along the Bed Due to Water Flow

What actually gives flowing water the capacity to carry sediment, and how are sediment particles transported?

We have seen that flowing water exerts shear forces against the stream bed. Frictional forces are converted into turbulence in the overlying water, and have the effect of transporting particles along the stream bed. Under moderate flow conditions the largest particles will be transported along, or just above, the bed as bedload (Fig. 2.8). This takes place partly through rolling or slow creep, partly through saltation, i.e. the grains jump along the bed.

Saltation can be partly explained through Bernoulli's equation:

$$P + g \cdot h + \frac{v^2}{2} = C(\text{constant})$$

Here P = pressure, h = height above the stream bed, and v = velocity. We see that water which flows over a sediment grain on the bed will have a greater velocity than water which flows under the grain. Bernoulli's equation predicts that the pressure above the grain must be less than the pressure adjacent to the grain (P), and when this difference becomes sufficiently great it will be possible to lift the grain from the stream bed. This "airplane wing effect" does not work once the grain is in the water above the stream bed, and the grain will then drop to the bed again.

The condition for sediment grains being transported in suspension is that their settling velocity must be less than the upward vertical turbulence component. This means that the grain must be transported upwards through the water at least as fast as it falls downwards. The magnitude of the vertical turbulence upwards will be a function of the horizontal velocity

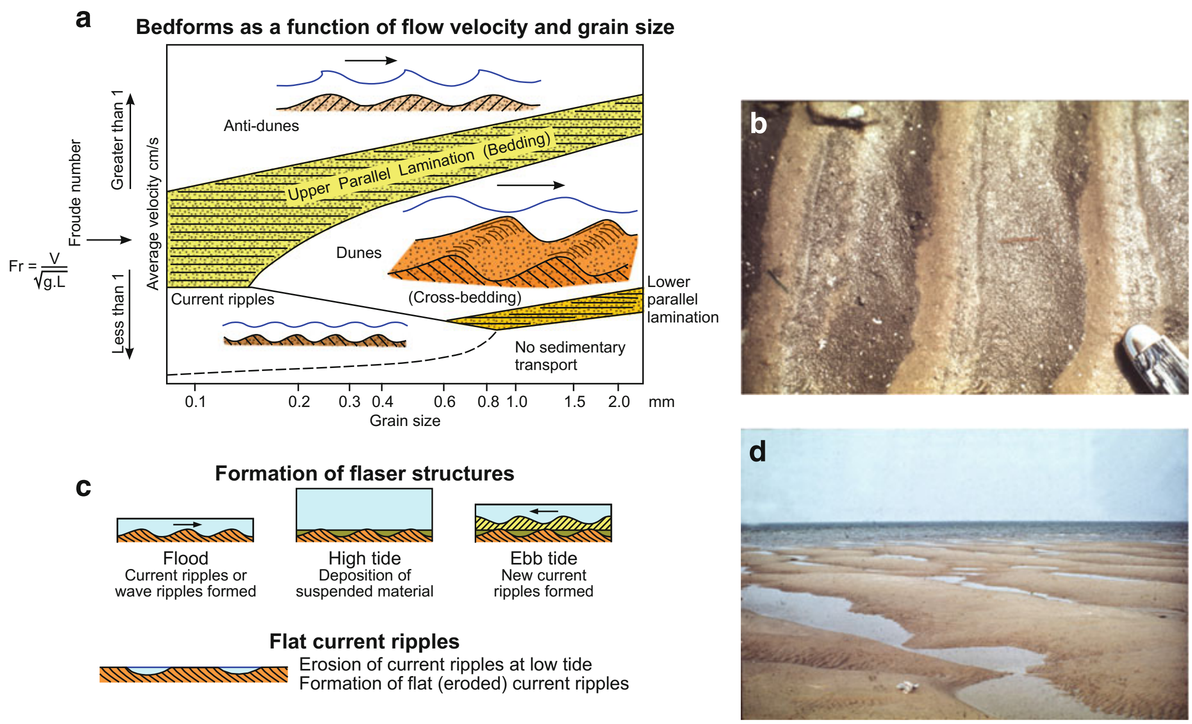


Fig. 2.9 (**a**) Sedimentary structures as a function of flow velocity, grain size and water depth. The Froude number (F) is an expression of the velocity as a function of depth. (**b**) Ripples with clay pellets accumulating between the ripple crests. (**c**) Formation of current ripples and truncated ripples in a tidal environment. (**d**) Dunes formed on a coastline with high wave energy

(about 1:8). Under normal flow conditions (< about 1 m/s) only clay and silt will be transported in suspension. Under high flow energy conditions, e.g. during floods, sand and gravel may also be transported – at least partly – in suspension.

Erosion and transport are a function of shear stress against the stream bed. This in turn is a function not only of water velocity but also of depth. There is a connection between the flow velocity and the size of the sediment grains which can be transported but the water depth is also important. Hjulstrøm's diagram (Fig. 2.10) applies to channels about 1 m deep. Other factors which complicate these relationships are the viscosity (and hence temperature) of the water and the density and shape of the sediment grains.

With small grain sizes (silt and clay) the flow velocity required to sustain transport is far less than the velocity needed to erode a particular grain. This is because cohesion between sediment particles, particularly clay, is such that once they are deposited, it is difficult to erode them again.

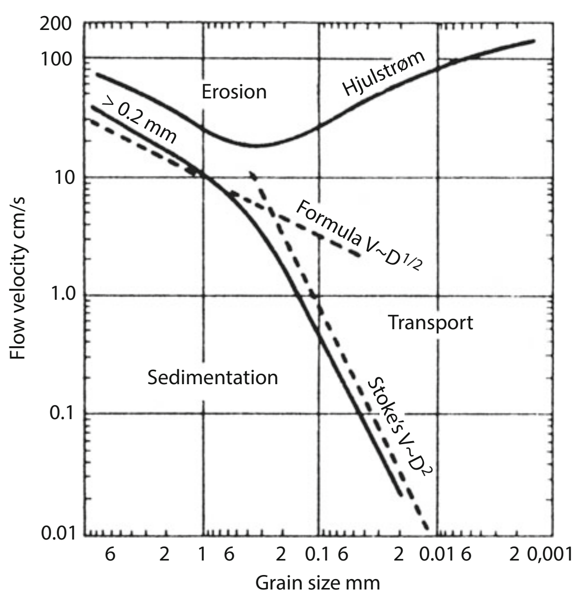


Fig. 2.10 Hjulstrøm's curve showing relations between grain size, flow velocity, erosion and sedimentation. Fine-grained sediments stay in suspension and are transported at low velocities, but require higher velocities to erode than silt due to cohesion

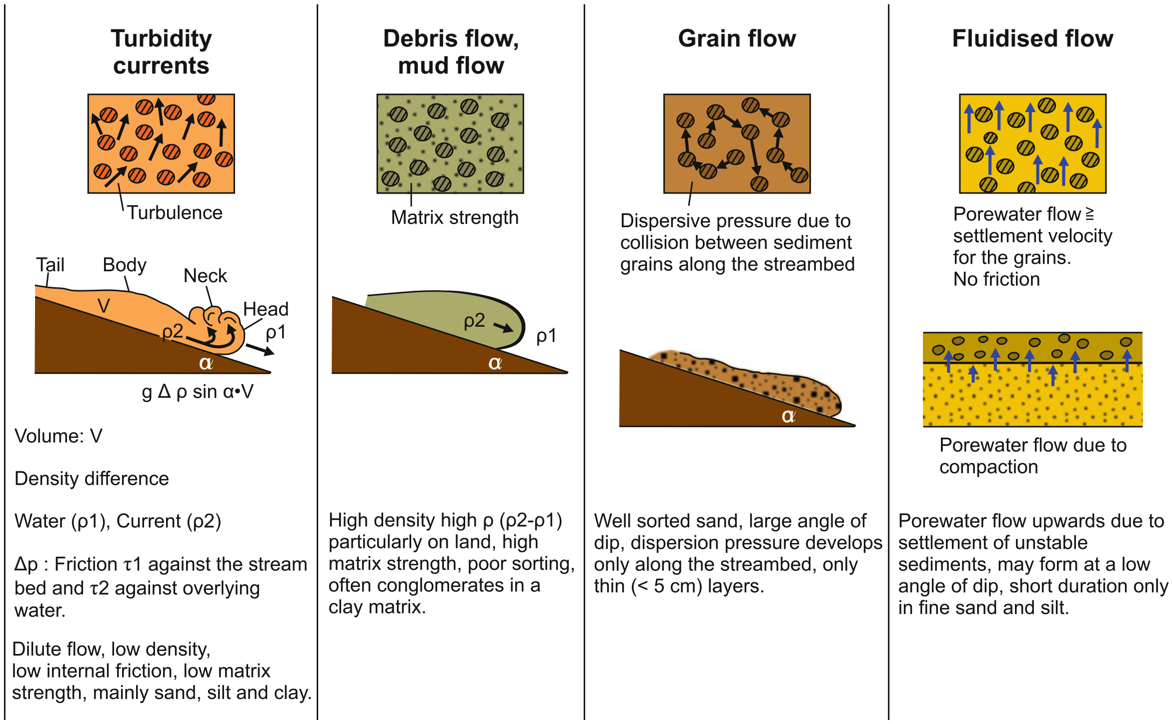


Fig. 2.11 Sketch showing different types of transport on slopes

Note that fine sand (about 0.1 mm) is the easiest sediment to erode. On the other hand, finer-grained particles remain in suspension for a long time at low velocities. Flocculation of small clay particles to form larger ones in seawater increases the settling velocity of clays. Also "pelletisation", through clay being eaten by organisms, is important for the formation of many fine-grained clay sediments.

2.10 Different Types of Sediment Transport

We have shown that water or air flowing over a surface exerts shear forces on the substratum so that sediment can be transported by what we refer to as *traction currents*. When we have a relatively low concentration of sediment in water (or air) there is little increase in the density and viscosity of the fluid phase. The flow will then still have approximately the same characteristics as it had without the sediment.

Another type of sediment transport is due primarily to the density difference between a water mass carrying suspended sediments and the clear water outside the suspension. We call this phenomenom *gravity flow* (Fig. 2.11) and it includes turbidity currents and debris or mass flows. The force of gravity causes movement of sediment/water mixtures because they have a higher density than their surroundings, i.e. they are not in equilibrium with the ambient clear water mass. Gravity flow is thus distinguished from traction currents by the fact that it can take place in otherwise still water. However, there are transitions between these two fundamentally different processes and we often have combined effects.

2.11 Turbidity Currents

Sediment in suspension will be carried down submarine slopes because the suspension is heavier than the surrounding clear water, forming a *turbidity current*.

It may be started by river water containing suspended material entering a sedimentary basin. In marine basins the difference in density between river (fresh) water and salt seawater is so great that even if river water carries sediment in suspension, it will in most cases not be denser than seawater.

In consequence there will not be a positive density contrast, which is the prerequisite for the formation of turbidity currents; instead the flow may become an overflow plume. In lakes, on the other hand, river water is often heavier, due both to suspended material and to its being colder than the lake surface water. It will then be able to follow the bed downslope, and become a turbidite.

Submarine slides may also quickly evolve into turbidity currents. River sediment entering a marine basin will mix with the seawater so that clays flocculate and are deposited on the delta slopes. If the slope angle becomes too steep, we get slides, which may result in turbidity currents. When fine-grained sediments are deposited on slopes from suspension they have a very high water content. Compaction, sometimes caused by earthquakes, will cause upward flow of porewater and may result in liquefaction. This causes the sediments to begin to flow even on gentle slopes because friction is reduced, and they may then turn into turbidity currents.

The forces driving a turbidity current are:

$$F_1 = g \cdot \Delta\rho \cdot V \sin\alpha$$

where g is the gravity constant, $\Delta\rho$ is the difference between the density of the current and that of the surrounding water, V is the volume of water along a certain length of the channel, with the cross-section (A) of a turbidity current with length L, and α the angle of the slope. Acting against the movement are frictional forces (F_2) which, as long as the current is not accelerating, must be equal to the gravitational forces. These are shear forces against the bed, τ_1, and against the overlying water, τ_2, plus internal friction and turbulence within the current which keep the sediments in suspension. In order for the sediment grains to remain in suspension, the turbulence must be sufficiently strong to have an upward component which corresponds at least to the settling velocity of the coarsest grains. Turbulence is greatest near the bottom of the current, where change in velocity as a function of height above the bottom (velocity gradient) is greatest. The largest grains in suspension will thus be concentrated near the bottom of the current. Near the bed, in addition to turbulence, we also have shear stresses which will transport the grains in virtually "pseudo-laminar" flow in a thin layer over the bottom. If the concentration of large sand grains along the bed becomes large, we also get *dispersive energy* because of collisions between the grains (see Sect. 2.13).

We therefore find that both the concentration of sediment in suspension, and maximum grain size in suspension, decrease upwards from the bottom. If we disregard internal friction, we obtain

$$g \cdot \Delta\rho \cdot V \sin\alpha = (\tau_1 + \tau_2) \cdot A$$

where A is the area of contact with the bottom and the overlying water. The ratio between the volume (V) and the contact area A is approximately the thickness of the flow H. The shear stresses are proportional to the square of the velocity ($\tau = cv^2$).

The velocity (v) of a turbidity current is then:

$$v = c \cdot (g \cdot \Delta\rho \cdot H. \sin\alpha)^{1/2}$$

Here the coefficient c includes the coefficient of resistance for friction against the seafloor and against the overlying water. This corresponds to Chezy's number for fluvial flow, so in many ways we can regard a turbidity current as an underwater river.

We see from the above equation that thick turbidity flows will have a higher velocity than thin ones and that thick flows can flow on gentler slopes than thinner ones. This is because the shear stress against the bottom and the overlying water is nearly independent of the thickness of the flow. The flow velocity also increases with increasing density of the sediment/water mixture in the flow, but high density flows will have higher internal friction and require higher velocities to keep the material in suspension.

A turbidity current can be divided into head, neck, body and tail. The sediment particles in the head area move somewhat faster than the front of the current itself. This leads to sediment being swept upwards and then backwards towards the neck, where it mixes with water from the overlying water mass. From there it is carried backwards to the body and tail, where we find a finer-grained, thinner suspension. When the turbidity current loses velocity, the largest particles in the head will settle out of suspension first because of reduced turbulence. Gradually smaller and smaller grains will settle and we get deposition of a bed which is fairly massive, without internal structure, but which becomes finer upwards. In most cases, apart from in proximal turbidites, we also find deposition of some

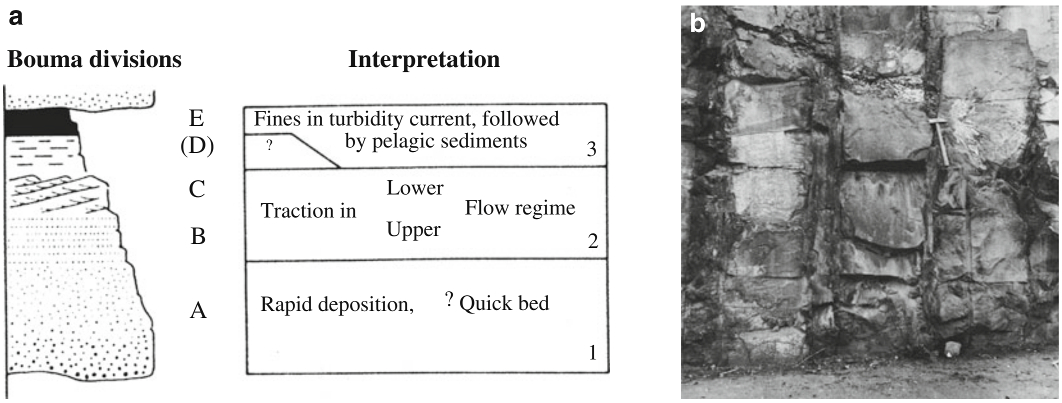


Fig. 2.12 (**a**) Bouma sequence in turbidites. (**b**) Turbidites in a Late Precambrian sequence at Lillehammer, Norway. The sequence is younging to the *right*

fine material in this layer, so there is poor sorting. An example of this is Unit A of the Bouma Sequence (Fig. 2.12).

As settlement from suspension slows down, the water will have time to sort the grains further, and we find lamination and bedding structures. The B Unit exhibits parallel lamination which may be due to flow just above the *upper flow regime* boundary. The C Unit exhibits current ripples and convoluted laminae and represents further velocity reduction, with deposition in the lower flow regime. The D Unit has parallel lamination and was probably deposited from the tail, which consists of very fine-grained sediment. The E Unit consists mainly of pelagic material, fine-grained clay and fossils that accumulated on the seabed during the long periods (often thousands of years) between turbidity flows (Fig. 2.12). The E Unit is therefore not necessarily a part of the turbidite sequence.

The Bouma sequence, first described by Arnold Bouma in 1962, is an ideal sequence in the sense that in most cases we do not find all the units developed. In some sequences, particularly those thought to have been deposited close to the base of submarine slopes where the gradient is still fairly steep, we will find only the coarsest parts of a turbidity current deposited, Unit A or a sequence of A + B. We call these *proximal turbidites*.The finest-grained fractions of a turbidity current tend to be deposited beyond the foot of the slope or out on the ocean abyssal plain. In these areas we often only find alternations between C-D-E, or just D-E. These are called *distal turbidites*. In many cases it may be difficult to distinguish between distal turbidites and alternations between silt and clay formed by traction currents at great depths. Proximal turbidites which are well sorted may also resemble coarser sediments deposited by powerful traction currents in submarine channels.

At the base of turbidity sequences, particularly at the base of the A Unit, we often find well-developed erosion structures, particularly *flute casts* and *groove casts*. Flute casts are formed by the turbulence of turbidity currents when they pass over a substratum which consists of finer-grained sediments. The structures, which point upcurrent, are produced by vortices in the turbulent flow eroding into the sediment surface. Groove casts are formed by larger grains being dragged along the bottom. Even though flute casts and groove casts are typical of turbidites, they cannot be used as proof that we are dealing with turbidites because similar structures can also be formed by various types of traction currents where there is turbulence and transport along the bottom, e.g. in fluvial environments. Sequences resembling Bouma sequences may also be produced by processes other than turbidity flows, for example rapidly accelerating fluvial flows. To assist our interpretation we should therefore look at the entire sequence and also try to obtain palaeo environmental information from fossils or oriented grains. On submarine slopes there may also be very swift traction currents, particularly in submarine valleys due to the focusing of the tidal forces, and relatively strong currents may go up the canyon as well. These traction currents are capable of transporting coarse sand and at times even coarser material.

2.12 High Density Mass Flows – "Debris Flows" and "Mud Flows"

Debris flows occur both on land and under water, and represent a type of mass transport where the sediment/water ratio is very much greater than in turbidity currents, resulting in high viscosity and high internal friction during flow. This also means that the density of the mass is from 1.5 to 2.0 g/cm^3, while most turbidity currents have a density of 1.1–1.2 g/cm^3 or less. The high density of a debris flow means that all clasts have increased buoyancy because of the dense matrix. The high viscosity of the matrix also means that large stones do not sink rapidly towards the bottom of the flow. In flows with high density and viscosity, blocks may remain near the surface of the flow until it solidifies, through loss of water or reduced gradients, and becomes quite rigid as a result of increased density and cohesion. The shear strength of the matrix is often referred to as matrix strength. Debris flows may be rich in stones and other coarse material because of the matrix strength. Mud flows typically have a more clay-rich matrix. But there is no clear distinction between the two.

Because of the high matrix strength, little sorting takes place in debris flows (Fig. 2.11). Large blocks are often concentrated at the front or on the sides of the flows, and there may be less coarse material near the base of the flow because of the shear movements.

On land, the density difference ($\Delta\rho$) between the flow and its surrounding fluid (air) is much higher than under water. Flows with a particularly low water content and high shear strength will only flow slowly down a slope, and we get transitions into what we call *solifluction* (creep). Sediment flows with a higher water content can move faster, however. In debris flows with their high internal friction most of the shear forces will be released along the bottom of the flow, so that the overlying mass moves more or less as a coherent mass with little internal deformation. This helps to reduce the total frictional resistance to movement. Shear strength and viscosity tend to decrease with increasing rate of shear, and this means that when a flow first gets going it will tend to accelerate.

Debris flows and mud flows normally have *thixotropic* properties: the shear forces must reach a critical threshold before deformation (shear) takes place, and the material loses much of its shear strength following deformation. In clay containing smectite (montmorillonite) this property is particularly well developed. Under shear stress, water will be released. The house-of-cards packing of clay minerals is destroyed, the clay particles tending to develop parallel alignment, with the release of water which will reduce the shear strength causing shear weakening. Some of the water in the bottom layer in which deformation is taking place will be lost to the other sediments, however, and friction will mount again. If large clasts enter the basal shear zone, friction will also increase and the flow may stop.

The stability of mud on slopes and the flow properties of mud flows depend on the clay mineral composition of the mud and on the geochemistry of the porewater and the ions adsorbed on the clay minerals. The presence of potassium and sodium tend to stabilise the mud.

Debris flows are often described in continental deposits, but are also found on submarine slopes. In water the density difference between sediment and surroundings is far less than on land, so that the angle of the slope must be greater for flows with the same internal friction. Submarine mud flows, on the other hand, will not dry up and they can easily take up more water as they move.

Debris flows are particularly common in desert deposits. This is because of the often powerful rainstorms which mobilise sediments that in a wetter climate would have been transported by fluvial processes. In addition, there is little vegetation in deserts to bind the sediments, so they are more easily set in motion. Also of great significance is the fact that the clay mineral smectite is formed particularly through weathering in desert environments. Clay containing smectite will expand when it begins to rain, preventing the water from filtering rapidly through the soil profile. Instead, water will be bound to the sediments and the viscosity may be reduced enough for mud flows with thixotropic properties to form. In continental environments with freshwater the content of stabilising salt (K^+, Na^+) is low.

The term "quick clay" is used for extremely thixotropic clays. Undisturbed clays have a relatively high shear strength, but after shaking or some other type of deformation they can flow like liquids, with a very low internal friction. In Scandinavia, Holocene marine clays which have been uplifted by glacio-isostatic rebound have been slowly weathered by percolation of rainwater so that sodium has been leached out, making them more prone to landslides and to form mud flows.

Systematic surveys using modern coring and remote sensing techniques, such as underwater cameras, side scan sonar and 3D seismic time slices, have shown that large-scale debris flows are rather common on continental slopes.

On the eastern slope of the Norwegian Sea a huge slide (the Storegga slide) occurred about 8,000 years ago, involving about 3,500 km^3 of sediment. The slope scar stretches for nearly 300 km and parts of the flow extended up to 800 km across the deep ocean floor. The transport mechanism was chiefly debris flow, where the sediments were riding as a plug on a wedge of water that reduced the friction against the bottom.

2.13 Grain Flow

Grain flow is flow of relatively well-sorted sediment grains which remain in a sort of suspension above the substratum due to collisions between the grains. We see this if we make a little landslide in a dry sandpit or pour sugar out of a bag. Grain flow can develop only when the initial flow is near the angle of repose (about 34°). Bagnold (1956) described how collisions between sediment grains led to *a dispersive stress*. However, this stress is only significant near the base of a flow, where we have rapid variation in flow velocity as a function of height above the base (dv/dh). Here grains with very different velocities will strike one another, and transfer velocity components to one another. Higher up in the flow the dispersive stress due to collisions between grains will be considerably less as the grains have far more similar velocities, despite turbulence. The dispersive pressure developed near the base cannot support a thick layer of overlying sediment and therefore grain flows have an upper thickness limit of about 5 cm.

Sand grains which avalanche down the lee side of sand dunes form small grain flows and are probably one of the few significant examples of natural pure grain flow. Grain flow may also occur on beaches and in shallow marine environments.

2.14 Liquefied Flow

Liquefaction is the name given to a process whereby sediments lose most of their internal friction, and consequently act almost like fluids. This is the case when the pore pressure is equal to the weight of the overburden. When sediments are deposited, they have a high water content and the sediment grains are packed in an unstable manner. As the overburden increases, the stress on the grain contacts increases, and the framework of the sediment grains may collapse suddenly. Earthquakes produce tremors which may cause this structure to collapse, but it can also take place purely as a result of stress (loading). When the packed framework of grains which was formed during deposition is destroyed, the grains can pack more closely together. For this to be able to happen, however, water must flow out of the bed as the porosity decreases. This leads to an upward flow of porewater and fine sediment particles, which may be as great as or greater than the settling velocity of the grains. This process is called liquefaction. The force of gravity, acting on the contact between the grains, is therefore neutralised, and friction between the grains tends towards zero, resulting in liquefaction. If we measure the pressure in the porewater, we find that it increases during settlement (compaction) when the unstable grain framework is destroyed. At one stage the pore pressure will be approximately as great as the weight of the overlying sediments. We can then use Coulomb's Law:

$$\tau = C + (\sigma_v - P)\tan\varphi,$$

where τ = shear strength, C = cohesion, σ_v is the weight of the overlying sediment, P = pore pressure and φ is the angle of friction (about 34°). $(\sigma_v - P)$ is the effective stress. When the pore pressure, P, approaches the weight of the overlying sediments (σ_v), the friction component $(\sigma_v - P)$ tan φ, approaches zero. Fine-grained sediments like clay have considerable cohesion (C), and this will often prevent clay sediments from sliding even if there is little friction. However, once a deformation plane forms, there will often be movement mainly along it

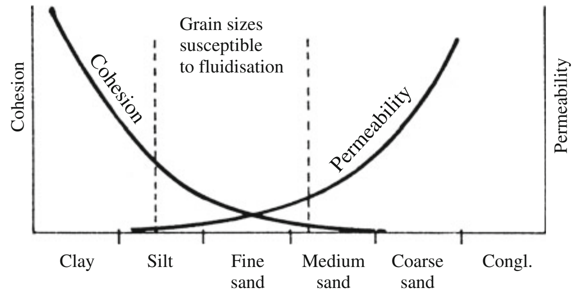


Fig. 2.13 Relation between permeability and cohesion in sediments. In coarse-grained sediments water escapes quickly, preventing build-up of overpressure, and in fine-grained clayey sediments the cohesion prevents mobilisation. Fine-grained sand and silt is therefore most mobile and likely to be liquefied and also injected as sand dykes (injectites)

due to cohesion in the rest of the clay. At relatively shallow depth clays may have a house of cards structure (Fig. 6.2) which will be deformed when subjected to shear stress. Clay minerals will then be more parallel oriented resulting in a shear softening. The excess porewater from the denser packing could produce an overpressure reducing the effective stress so that the frictional forces are reduced. Smectitic clays are much more fine grained and have higher cohesion than e.g. kaolinitic clays. If we have coarse-grained sediments, i.e. coarse sand and gravel, compaction will lead to excess water flowing out so rapidly that the overpressure will drop very quickly, assuming the high permeability has allowed it to build up properly in the first place (Fig. 2.13). It is therefore silt and fine sand, the fractions most susceptible to liquefaction, that are likely to generate high-velocity subsea flows. Liquefaction can, as already mentioned, be triggered by tremors, e.g. earthquakes, and stress. Stresses on sediments (soils) due to buildings, fills, etc. can lead to collapse of the grain frameworks and cause liquefaction. Lowering of the groundwater table on a slope, for example down towards the coast, has a similar effect because of reduced buoyancy in part of the sediment column. Extremely low tides or a combination of a strong ebb and a land wind can trigger a slide in otherwise stable coastal sediments. This is because the effective stress in the sediments increases when the sea level is low.

The stability of slopes can be estimated by calculating the gravitational forces acting on a particular volume of sediment in relation to the frictional forces. During construction work, slides may sometimes be prevented by drilling wells which release the excess pore pressure so that the effective stress and the friction increases.

2.15 Sedimentary Structures, Facies and Sedimentary Environments

It is difficult to observe or take measurements of rocks entirely objectively and consistently. Most types of measurements and observations have a considerable degree of inherent uncertainty, and the validity and usefulness of results often depend on the experience and skill of those carrying out the field work. It has turned out to be very difficult to observe structures which one does not recognise and understand the significance of.

A good description of a stratigraphic profile depends on good theoretical knowledge of sedimentary processes, and of experience from studies of similar rocks.

It is easy to forget to record or measure some of the properties of a rock. In order to obtain a more comprehensive description and avoid forgetting anything, it may be a good idea to have a well-established routine or even a checklist. Photographs from outcrops or cores may help when writing final reports. If our investigation has a definite and limited objective, we measure only the properties we think will be relevant. A list of features which can be observed (measured or registered) in sedimentary rocks:

1. Textures – grain size, sorting, grain shape etc.
2. Grain orientation – fabric.
3. Sedimentary structures and their orientation.
4. Fossils.
   A. Preservation or impressions, casts, or the fossils themselves, and their mode of occurrence.
   B. Trace fossils.
5. Colour.
6. Resistance to weathering and erosion.
7. Composition (a) Mineral (b) Chemical.
8. Thickness and geometry of beds.
9. Variations in texture and composition within a bed, e.g. increase in grain size upwards or downwards in the bed (grading or inverse grading).
10. Type of contact between beds (e.g. erosional contact, conformable contact, gradational contact).

11. Association or any tendency to statistical periodicity in the features of the strata in a profile – bed types, structures.

These observations form the basis for defining facies, which are a synthesis of all the data listed above which can be used to group certain types of rocks. They may be genetic facies, i.e. strata which one assumes have formed in the same manner. All strata which contain criteria which indicate that they were deposited in shallow water can be described (in reality interpreted) as shallow water facies. In the same way we have fluvial facies, deep water facies, evaporite facies, and so on. The facies concept can also be used to distinguish between different rock compositions (lithologies), e.g. carbonate facies, sandstone facies.

2.16 Sedimentary Structures

By sedimentary structures we mean structures in sedimentary rocks which have formed during or just after deposition. We distinguish between *primary* structures which are formed at the time of deposition of the sediments, and *secondary* structures which are formed after deposition.

2.17 Layering and Lamination

Most sedimentary rocks exhibit some lamination or bedding, but we also have massive (unlaminated) rocks. Lamination records variations in the sediment composition as successive depositional layers are draped over the contours of the sedimentary surface. The variations may reflect different grain sizes, sorting, mineral composition or organic matter. *Laminae* are less than 1 cm thick. Units greater than 1 cm are called *beds*. A bed will contain sediments which have been deposited by the same sedimentary processes. Some sedimentary processes, for example deposition of a turbidite bed, may be fairly rapid. Migration of a sand dune to give cross-bedding takes somewhat longer, while deposition of a clay bed from suspended material may take a very long time.

Graded beds have a grain size which tends to decrease upwards within the bed. The opposite is called *inverse grading*. *Normal grading* may be due to deposition from suspension, when the largest particles tend to fall to the bottom first, as with turbidites, or to flow velocities dropping off during deposition in a river.

Inverse grading may be due to increasing flow velocity but if the increase in velocity is too high, the result will be erosion. The supply of coarse material during transport and deposition may also produce inverse grading. In high density sediment currents (debris flows) we may get inverse grading; smaller particles sink more easily to the bottom between the large particles. Massive beds, or at any rate reasonably massive beds without visible lamination or bedding, may be formed during very rapid deposition of sediments from suspension. X-ray photography of apparently massive sediments nevertheless usually reveals the presence of weak lamination even in sand. Massive sand beds occur due to rapid fall out from suspension. In mudstones the primary lamination may be destroyed by intense bioturbation.

On the surface of laminated sediment we may find erosion structures.

Water running over a plane surface, e.g. beach sand, will produce small-scale, branched (dendritic) erosion marks called *rill marks*. The flow of water may be due to runoff from big waves or from groundwater seepage at low tide. These structures are good indicators of inter- or supratidal environments, but they are seldom preserved because they are usually destroyed when the water level rises again. Raindrop imprints are often preserved on bedding surfaces, and are good indicators of subaerial exposure.

2.18 Bedforms

Bedforms are morphological features resulting from the interaction between particular types of flow and the sediment grains on the bottom. A specific type of bedform will only form within a limited flow velocity range and is also dependent on the availability of grain sizes which can be moved by that flow.

Current ripples form in fine-grained sand when the velocity exceeds the lower limit for sediment movement.

Ripples and dunes have a stoss side upstream where erosion takes place and a lee side on which deposition

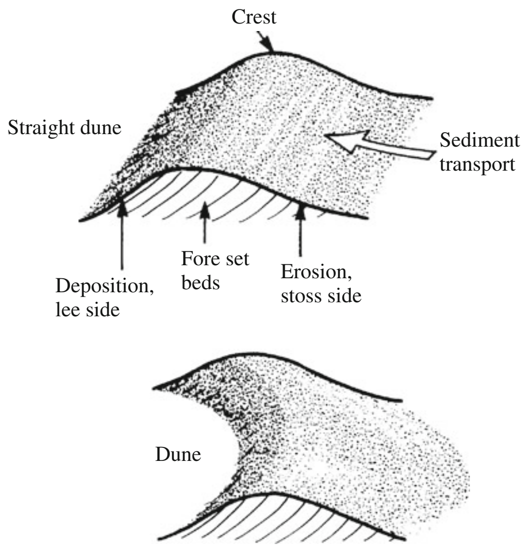


Fig. 2.14 Principle illustration of dune and cross-bedding

takes place (Fig. 2.14). They therefore migrate as a result of the combined effects of erosion and deposition. Current ripples form in fine-grained sand when the velocity exceeds the lower limit for sediment movement. Sections through ripples show inclined foreset laminae, a structure called “small-scale cross-lamination”. Ripples may also form in coarser sand but there is an upper limit of 0.6–0.7 mm for the grain diameter. Ripples are less than 3–5 cm high, and may have a wavelength of up to 40 cm. The ripple index is an expression of the ratio of the wavelength divided by the wave height, and varies between 10 and 40. Waves may generate oscillatory flow that produces symmetrical ripples, which have foreset laminae pointing in both directions.

Dunes are similar to ripples in shape and structure, and form in coarse-, medium- and fine-grained sand, but require significantly higher flow velocities for their formation. Dunes range in height from 5 cm up to several metres, and wavelengths may exceed 10 m (Fig. 2.15).

Cross-bedding (large-scale cross-stratification) is seen in cross-sections through dunes (Fig. 2.16). Each lamination is called a *foreset bed* and represents the lee-side surface of a migrating dune. If the dunes have straight crests the foreset beds on the lee side will form a straight transverse plane (*tabular cross-bedding*). Curved dunes have rounded foreset beds, and

Fig. 2.15 Aeolian dunes several metres high in the Navajo Sandstone (Jurassic)

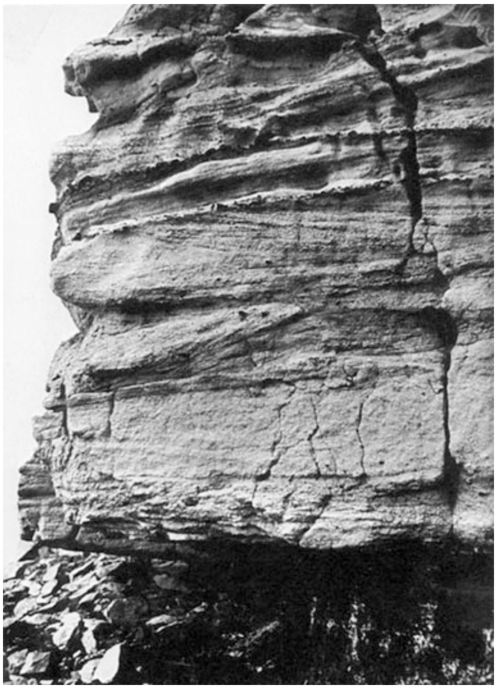


Fig. 2.16 Cross-bedded sandstone which represents the filling of a fluvial channel above a coal bed. The grain size fines upwards somewhat. It is largely trough cross-bedding represented here. From the Tertiary sequence of Spitzbergen. (Photo A. Dalland)

in *trough cross-bedding* the laminae have a rounded surface which is concave in the downstream direction. The foreset lamination may form a relatively sharp angle with the underlying bed, or may have a more tangential contact. The latter is typical of trough-shaped sets.

Aeolian dunes may be many metres high, and their cross-bedding will then be correspondingly large (Fig. 2.15).

Whatever the size of the bedform, ripple movement provides the basic mode of migration. Seabed survey profiling records clearly show smaller ripples climbing the stoss sides of larger ones (so-called "megaripples", i.e. small dunes), which in turn are climbing the stoss sides of sandwaves.

Cross-sections through dunes show "large-scale cross-stratification", which is often referred to as "cross-bedding". This can be observed in real time on seabed video recordings made during periods of strong tidal current flow, where gradual forward movement of "megaripples" is seen, caused by sand cascading down the lee side after reaching the crest.

"Plane beds" (upper stage) may form when the shear stress against the bed exceeds the values which produce dunes. In cross-section we only see planar lamination, which is an internal structure, but on the bedding surface we may see very small ridges, which define a lineation called primary current lineation, parallel to flow.

At even higher velocities in relatively shallow water, standing waves may produce antidunes when the Froude number exceeds 0.8. The antidunes which are produced when standing waves are in phase with the bedforms develop resulting in low-angle cross-lamination which dips up-current.

Both ripples and dunes are formed through sand being transported along the bottom and deposited in sloping strata on the lee side of the structure. In consequence they always have dipping laminations *(foreset beds)* which may lie at an angle (angle of repose) of up to 35° to the surface of the bed, though such high angles are rather rare. Current ripples in plan view may be straight, or form curved patterns *(sinuous crests)*. Ripples with a symmetrical cross-section (symmetrical ripples) are formed by waves as a rule. Asymmetrical ripples are formed by a predominantly unidirectional current and their steeper side faces downstream. Ripples with a high sinuosity are also asymmetrical in most cases.

Tongue-shaped *(linguoid)* ripples have a very high sinuosity and asymmetry and are usually formed in shallower water or under higher velocities than straight-crested types. Wave ripples in particular may split laterally into two ripples. This is called bifurcation. In intertidal zones ripples formed at high tide may be eroded at low tide, and the crests become flattened. When the tidal flat is submerged at high tide it may also be below the normal wave base, resulting at slack water in deposition of clay which tends to collect in the ripple troughs. Current ripples with thin lenses of clay between them constitute *flaser bedding*. Wind may generate waves moving in different directions, particularly at very low water, so that we find two or more sets of ripples at an angle to each other *(interference pattern)*. In most cases each bed with current ripples represents a sort of equilibrium with deposits reflecting current patterns. Isolated sand lenses in clay are called *lenticular bedding*.

Normally current ripples form completely horizontal beds. In some cases, however, we find examples of current ripples appearing to climb downstream in relation to the horizontal plane. They form several sets of cross-laminated beds delimited by erosion boundaries, but with small internal erosion planes. These are called *climbing ripples* and are due to sedimentation taking place so rapidly that, in contrast to normal ripples, there is no equilibrium between erosion and sedimentation. Climbing ripples are therefore typical of environments with rapidly declining flow velocity and consequently a high rate of sedimentation.

Sandwaves are large-scale transverse bedforms, generally 2–15 m high, with a wavelength of 150–500 m. They may be formed at flow velocities of 65–125 cm/s. Cross-stratification may be symmetrical or asymmetrical on both sides of the sandwave, depending on the relative strength of the opposing currents.

At velocities of less than 1 m/s *sand ribbons* may be deposited – longitudinal bedforms developed parallel to the currents. Sand ribbons are typical of subtidal environments (20–200 m), and they may be up to 20 km long, 200 m wide and less than a metre thick.

Relatively low-energy environments (<50 cm/s) are characterised by *sand patches* and mud. The sand forming the patches probably only moves during storms. Lateral structures are typically current ripples and small dunes.

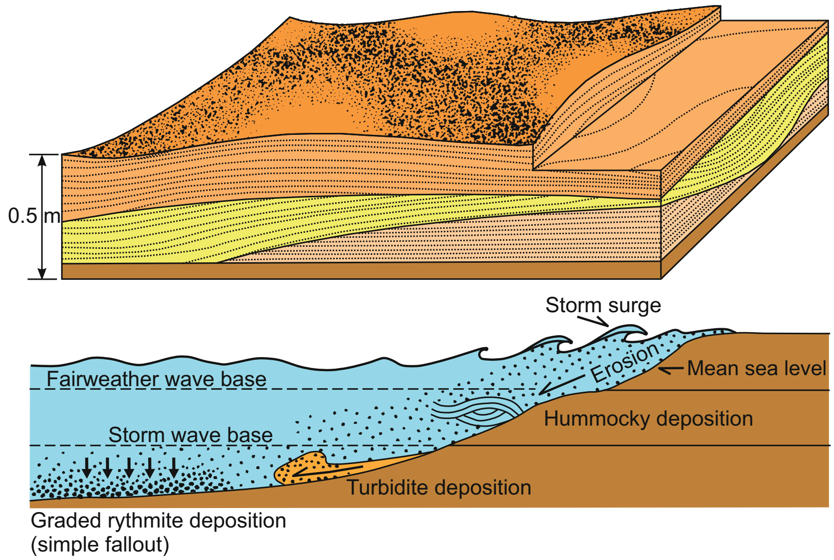


Fig. 2.17 Hummocky stratification (After Harms et al. 1975, Walker 1982)

Sandwaves often have current ripples ("smaller dunes") on their surface. This may also be the case with dunes. Bedforms are a function of both flow velocity and grain size. Current ripples are formed only in silt and fine- to medium-grained sand, while dunes require medium to coarse sand. Upper flow regime plane beds which have an internal structure of planar lamination are formed when the Froude number is about 0.6–0.8. Plane beds typically develop in beach sand. A well-developed lineation on the bedding surface parallel to the direction of sediment transport is typical of plane beds (Fig. 2.9). Antidunes are formed by higher velocities in the upper flow regime; the absolute flow velocities required are lower when the water is shallow.

Below the ordinary, fair weather wave base we find different types of sedimentary structures from those formed through constant wave action. During storms in nearshore areas traction currents may develop and carry fine material (fine sand) from the beach out to greater depths. This material is deposited as *hummocks* consisting of parallel laminae which form dome-shaped structures (Fig. 2.17). Because deposition of beds of this type takes place during short periods during storms, they are often reworked by bioturbation near the top. Hummocky stratification is common in sections through some shallow marine deposits and wave-dominated deltas.

2.19 Erosion Structures on the Underside of Sand Beds *(Sole Structures)*

When sand is deposited rapidly over finer sediments (silt and clay) there will also in very many cases be an initial erosional phase, so that erosion hollows are formed in the substrate. These hollows fill with sand, and we get a sort of *cast*. We see these erosion structures on the underside (sole) of sandstone and conglomeritic strata. We seldom observe these surfaces unless the bed is steeply inclined or inverted. Wherever there are overhanging rocks, it is important to study the underside of the bed.

Flute casts are formed by static vortices in the water eroding into the underlying sediments. Their sharp end points upstream and they broaden in the downstream direction. Flute casts (Fig. 2.18) are good indicators of current direction, which is measured along an axis of symmetry through the structure. They are typical on the underside of turbidite beds, but also occur in fluvial sandstones as a result of fluid turbulence.

Similar structures are formed around objects which project up from the bottom, such as stones and large fossils.

Gutter marks are longitudinal grooves up to 20 cm deep and rather narrow (less than 20–30 cm) with a spacing of 1 m or more. They are the result of

Fig. 2.18 Flute casts on the lower surface (sole) of a coarse-grained sandstone bed in the Ring Formation, Rena, South Norway. Note the small flute casts on the large flute cast structure. Flute casts are casts of the erosion structures formed in the finer-grained underlying bed by vortices

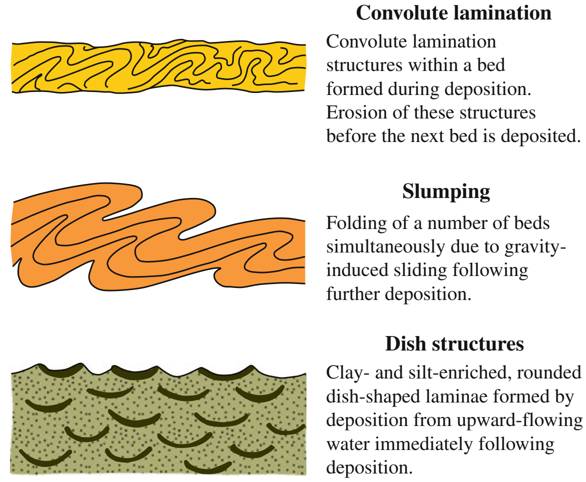


Fig. 2.19 Sedimentary structures due to liquefaction and soft sediment deformation during the deposition process. See text

channelised flow, and flute casts or groove casts may be found along their margin.

Ridges and furrows may also be formed through erosion of the substratum. Some are U-shaped or V-shaped channels. Erosion structures formed by objects which are transported along the bottom are called *tool marks*. The objects may be fossil fragments or larger sediment particles.

Groove casts are long, narrow erosion furrows due to something being dragged along the bottom.

Chevron marks are erosion furrows with V-shaped structures in the clay sediments on either side. The V-structure closes downstream, and is due to a cast forming in cohesive clay.

Prod marks show where an object has dug down into the clay and then been plucked out again by the current. As a result the steep side of prod marks is the downstream side.

Bounce marks are rows of more symmetrical marks due to objects being swept or bounced along the bottom.

2.20 Deformation Structures

Sediments are often unstable immediately after deposition, and later movements will deform the primary structures.

Deformation may be caused by four main factors:

1. Shear stress due to water or sediment movement, e.g. convolute lamination.
2. Expulsion of porewater (liquefaction, dewatering), e.g. dish structures, clastic dykes.
3. Heavier beds above lighter beds (inverse density), e.g. load casts and ball-and-pillow structures.
4. Gravitational deformation. Gravitation-induced sliding, folding and faulting on a slope, e.g. slumping.
5. Shrinkage, e.g. due to dessication, or permafrost (ice wedges).

It is important to distinguish between these types of deformation structures because they have completely different implications for interpreting depositional environments. They are not mutually exclusive, however, and may be found in close association in the same sequence.

Convolute lamination forms in fine sand or silt, and is due to laminae, e.g. current ripples, being deformed almost as they develop. Folded and inverted (overturned) lamination is typical, and the structure is deformed in the downstream direction due to the stress induced by the water movement and the instability (almost a state of liquefaction) of the sediments (Fig. 2.19).

Dish structures (Fig. 2.19) are thin, clay- and silt-enriched dish-shaped laminae in sandy sediments. Their structure is due to the porewater which flows upwards immediately after deposition, transporting clay and silt which become trapped in these thin laminae. *Sand dykes (or clastic dykes)* are intrusions of sand upwards into cracks in a finer-grained sediment, due to porewater overpressure. The overpressure reduces the friction between the grains and injects water with sand into vertical fractures produced by high overpressures or follows bedding planes as sills. Overpressurised porewater with sand may rise right to the surface and form *sand volcanoes*.

Well-sorted sand has at the time of deposition an initial porosity of 40–45%, whereas recently deposited mud contains 50–80% water. When sedimentation is rapid, the mud has little time to lose its excess water, and a sand bed deposited on top may then sink down into the underlying silt and clay and form *load structures*. On the lower surface of a sand layer we often see pillow-shaped depressions surrounded by clay which has oozed up around them. If this process continues, it will form isolated sand pockets in the underlying clay: *ball-and-pillow structures*. Primary structures such as flute casts often sink in the underlying mud and are deformed by *loading* (Fig. 2.20).

This mechanism also operates on a larger scale, for example channel sand will sink down into surrounding clay (see "deltas"). Poorly compacted clay and silt will be lighter than surrounding sediments, and flow upwards to form clay diapirs.

Gravitational deformation occurs in sediments which are deposited on slopes. These forces can be resolved into a vector normal to the bedding, and a shear stress parallel to it. The vector which acts along the bedding is proportional to the sine of the angle of dip, and acts as a compaction force which can slide and fold the beds. In the upper part of the slide, tensional deformation is prominent, producing faulting, while compression occurs near the base of the slide and produces folding. The result is called *slumping* (Fig. 2.21). Slumping occurs most readily where we have rapid sedimentation and therefore relatively thick beds with a high water content and low shear strength. Deformation takes place when the shear stress exceeds the shear strength. The shear stress increases with the thickness of the unconsolidated sediments, but there is not usually a corresponding increase in shear strength with increasing thickness. Slumping may resemble convolute lamination, but normally affects more than one bed. Gravitational deformation also leads to faults on a greater or lesser scale. Sliding of large volumes of sediment down slopes produces slope scars. *Growth faults* and other types of "listric" faults are a result of large-scale gravitational deformation in the upper part of an area which is under tension.

Dessication cracks or mud cracks are examples of contraction or shrinkage of sedimentary beds due to dehydration. Cracks frequently form regular polygons, often hexagons or orthogonal sets. Dessication cracks form only in clay and silt, and the cracks often become filled with sand, resulting in good contrast. The formation of dessication cracks requires that the beds be exposed to the air so that the sediments can dry out.

Fig. 2.20 Load cast structures at the base of a sandstone bed in the Late Precambrian Ring Formation at Rena, Southern Norway. The picture covers an area of 3 × 4 m. (Bjørlykke et al. 1976)

Fig. 2.21 Sediment beds which have been folded immediately after deposition (through slumping) due to sliding on submarine slopes. Note that the overlying beds are undeformed showing that this is not tectonic folding. From the Ridge Basin (Miocene-Pliocene), California. Scale, John Crowell

In certain cases shrinkage structures may also form underwater through dehydration of clay minerals (smectite) as a result of variation in the salinity of the porewater *(syneresis)*. Shrinkage structures are less regular than desiccation cracks and are not usually interconnected. In the smectite-rich Eocene sediments from the North Sea basin and the Norwegian Sea, seismic data show large (several hundred metres wide) polygons which have been interpreted as shrinkage cracks.

When the porewater in sediments freezes to ice and remelts, we also find expansion and contraction which results in polygonal surface marks and associated vertical ice wedges.

2.21 Concretions

Concretions are round, flat or elongated structures which consist of cement which has been chemically precipitated in the pores of the sediment. The most common types of concretion are carbonate (calcite and siderite) and silica (chert). Sulphides, particularly pyrite, also form concretions.

A characteristic feature of concretions is that any laminations in the sediment pass through the concretions. This shows that the concretion has been formed through passive filling of its pores. As the overburden increases, the sediments around the concretion will be subject to compaction, while the concretion cannot be compressed because the pores are full of cement. A concretion therefore has a cement content which corresponds to its porosity at the time of formation. Carbonate concretions in clay may have a carbonate content reflecting 50–70% porosity if the matrix does not contain carbonate. Concretions in calcareous rocks, i.e. marls, contain clastic or biogenic carbonate in addition to carbonate cement, and therefore have more carbonate than the matrix. In these cases the carbonate content cannot be taken as an indication of the porosity at the time of formation. Concretions often contain fossils, showing no sign of compaction while the same fossils are severely deformed and sometimes also dissolved outside the concretion. In carbonate sediments, particularly chalk, there are silica concretions (chert). These are formed through precipitation of finely divided amorphous silica to form a type of chert called *flint*. The source of the silica is usually amorphous biogenic silica, frequently sponge spicules.

2.22 Trace Fossils

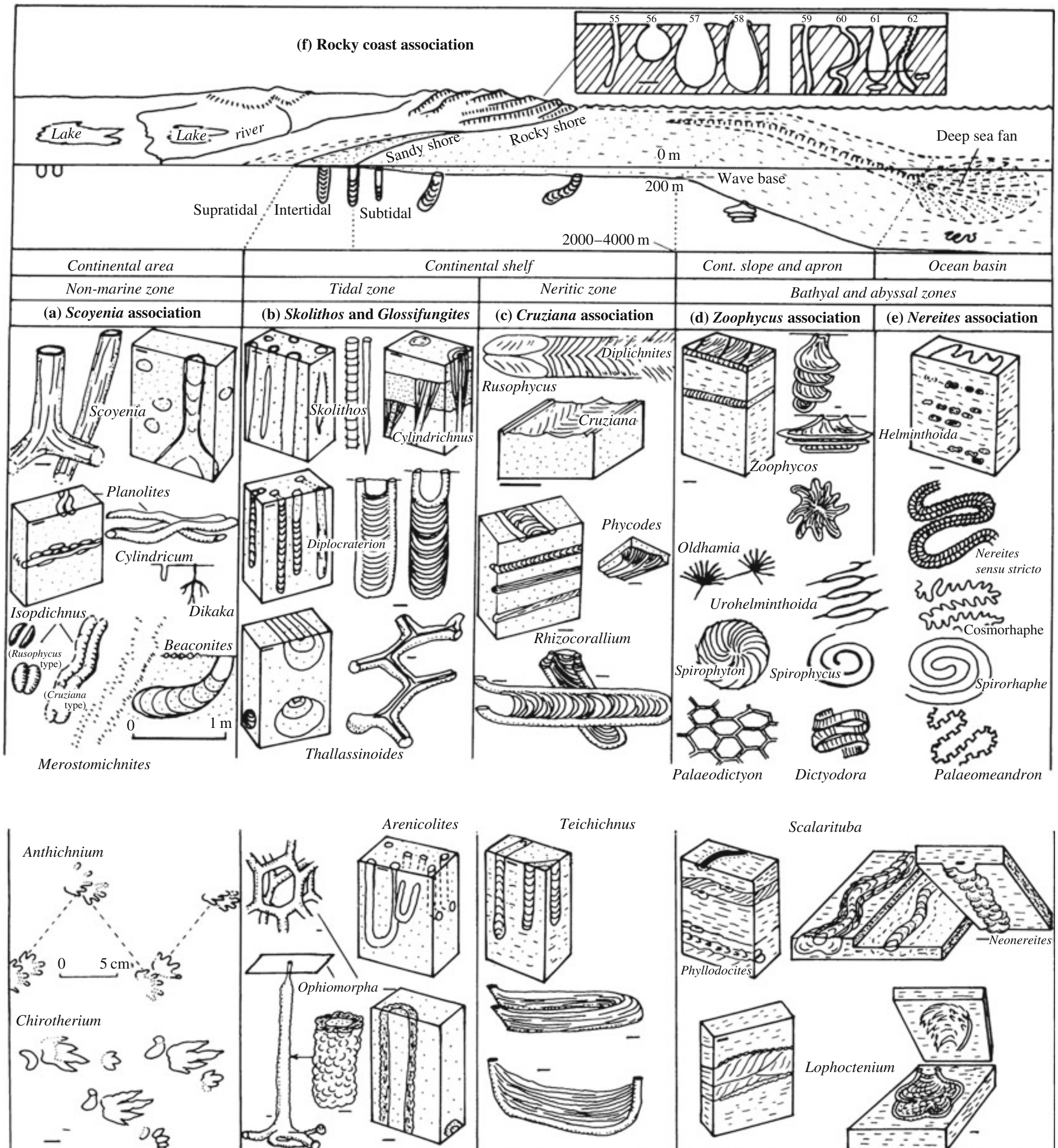
Trace fossils are structures in sedimentary rocks which have been left by organisms that lived on and/or burrowed in the sediment. Such organisms are extremely sensitive to changes in the composition of the nutrient content, the sedimentation rate and bottom currents, and are therefore useful indications of the environment (Fig. 2.22a,b). Trace fossils are therefore good indicators of the depositional environments. They can be classified taxonomically, i.e. according to the animal which left the traces, but it may not be possible to determine which animal was responsible. The same species may form several different types of trace depending on the sediment composition and on its mode of life. Furthermore, different animals may leave traces which are so similar that they are classified as one trace fossil. For these reasons a descriptive, morphological classification of trace fossils is used.

Trace fossils can also be classified according to where they occur in relation to the bed:

1. On top of beds (e.g. a thin sand or carbonate layer Epichnia).
2. Within the bed (Endichnia).
3. On the lower surface of the bed (Hypichnia).
4. Outside the bed (Exichnia).

The impression made by an animal may create a *mould* or a *cast* in the overlying bed. Organisms which burrow into sediments often secrete a cement which ensures that the walls of their burrows do not collapse. These secretions also contribute to preservation.

When worms eat sediment, for example, it passes through their digestive organs and their burrow refills with a sediment which has a somewhat different composition from the surrounding sediments. This is particularly noticeable in burrows at the interface between two strata with different compositions. The amount of bioturbation reflects nutritive conditions and the sedimentation rate. With very rapid sedimentation there will be less time for organisms to burrow through the sediments. Where we have very slow sedimentation or an hiatus, the sediments will often be thoroughly churned up by bioturbation, and thereby homogenised. Bioturbation is most widespread in marine environments, but can also be found to a lesser



	(a)	(b)		(c)	(d)	(e)
Energy (waves, wind)	Variable: waves, river, air currents	Higher energy, constant wave mixing	Lower energy, frequent wave mixing	Lower energy, frequent wave mixing	Lower energy, infrequent wave mixing, some density flows	No wave mixing, some density flows, ocean current flows
Eh	Oxidising, except at lake bottoms	Oxidising	Oxidising	Oxidising	O_2 reduced, high organic content	Little O_2, e.g. from density flows, even anoxic
Salinity	Low, rarely high	Mostly normal	Normal	Normal	Normal	Normal
Temperature	Extreme variety	Daily changes	Daily changes	Seasonal changes	< 10°C, no changes	Down to 2°C, no changes
Light	Daily changes	Daily changes	Daily changes	Daily changes in upper part	None	None
Substrate sediment	Aqueous or subaerial; sand, silt, mud	Sand > mud, less firm, reworked	Mud = sand, firmer, reworked	Mud = sand, stable, rarer reworking and ripples	Mud, stable except in failure and density flow	Pelagic mud, most stable but for density and ocean current flows
Diversity	Lower than in marine realm	Low	Low	High	Low	Low, but higher than in (d), higher through time
Abundance	High in places	Dense	Dense	High	Low	Low, but seems higher due to slow deposition
Dominant organisms, traces	Arthropods (insects) molluscs, vertebrates, small shallow burrows, trails, tracks, herbivores, carnivores	Arthropods, molluscs, echinoderms, corals, 'worms' Unbranched vertical burrows Mostly suspension feeders	Vertical and inclined ear-shaped burrows	Arthropods, molluscs, echinoderms, corals Horizontal crawling – grazing traces > inclined burrows Mostly sediment feeders	Arthropods, polychaete worms, hemichordates (e.g. acorn worms), echinoderms Complex horizontal grazing and shallow feeding traces, spreiten inclined in sheets, ribbons, spirals – bioturbation Sediment churners, feeders	Casts, crawling, grazing traces, spreiten planar and on surface Sediment grazers, farmers

Fig. 2.22 Common trace fossils in different sedimentary facies (from Collinson and Thompson 1982)

extent in freshwater sediments. In poorly oxygenated environments there is, however, little bioturbation.

In high energy environments, e.g. above the fair-weather wave base, we only have vertical trace fossils, e.g. *Skolithos* or *Diplocraterion*. The high current velocity prevents these organisms from crawling around on the bottom; they have to burrow down into the sediment and live by filtering nutrients out of the water. To remain at the same level beneath the sediment surface they must move upwards or downwards in their holes, depending on whether erosion or sedimentation is proceeding in the area. Structures or fillings which reflect such adjustments are called "spreiten".

In modern marine environments we can study a number of organisms which create bioturbation structures. The most common are worms like *Arenicola* which make U-shaped traces in fine-grained sand and silt. Such traces (arenicolites) are also to be found in older rocks. Burrowing bivalves create various types of trace as well. Arthropods like crabs and prawns make burrowing structures in beach sand *(ophiomorpha)*. These vertical trace fossils are grouped together in an ichnofacies called the *Skolithos* facies. *Thallassinoids* are more horizontally aligned networks of arthropod burrows.

Below the wave base and in other protected environments, for example the intertidal zone, we find traces in the horizontal plane from organisms which live off blue-green algae and other organic material on the surface of the sediment. In this *neritic* zone (Fig. 2.22a,b) we find a number of different types of traces from organisms which eat their way through sediments, and which form different patterns. This is called the *Cruziana* facies. *Rusophycus* and *Cruziana* are typical of the neritic zone and represent horizontal traces of arthropods which feed on the sediment surface. *Rhizocorallium* and *Teichichnus* are other trace fossils that occur below the *Cruziana* facies. In deeper water where wave and current energy is even lower, we find *Zoophycus* and *Nereites* facies. It is important to remember that these environments are primarily a function of wave and current energy, and cannot simply be correlated with absolute depth. In shallow enclosed seas with a shallow wave base, e.g. the modern Baltic, we find an effective wave base of only 5–10 m in many areas, while elsewhere we may have stronger currents along parts of the deeper trenches. In the epicontinental Cambro-Silurian marine sedimentary sequence of the Oslo area we find *Nereites* facies in the shales, but the water depth was probably not more than 100–200 m, perhaps even less. In the deep oceans the *Nereites* facies may correspond to a depth of several thousand metres.

Trace fossils are very useful facies indicators and should be noted whenever sedimentary sections are examined for facies interpretations. Certain trace fossils can be linked with animals that have fairly specific environmental requirements.

2.23 Facies and Sedimentary Environments

The word "facies" is used in a number of geological disciplines. A term such as "metamorphic facies" is thoroughly entrenched. Sedimentary facies have also long been identified in sedimentology to distinguish between sedimentary rocks which differ in appearance and have formed in different ways. The term facies can be used both descriptively and genetically. We use terms like "sandy facies", "shaly facies", "carbonate facies" when we are describing properties of the rock that can be observed or analysed objectively. We use terms such as "shallow water facies", "deep water facies", "turbidite facies", "deltaic facies", "intertidal facies", "aeolian facies", "reef facies" etc., depending on which environment we believe the rocks represent. In these examples, the word "facies" represents an interpretation, and is therefore not very suitable for describing sedimentary rocks objectively. For this reason it is important that we define the objective criteria (observations) on which we are basing our interpretations.

What we can do, then, is first describe and take measurements on a series of beds in the field, and on the basis of certain criteria divide the series into facies. The criteria will generally be texture, sedimentary structures, mineral composition, and, if present, also fossils. Interpreting a facies in terms of depositional environment is often very difficult, especially since few criteria are unambiguously diagnostic of one particular environment. In some cases it may be useful to use statistical methods for distinguishing between different facies and for describing facies sequences.

What we observe and measure is a selection of the properties of the rock. When we describe a sedimentary rock, we observe the results of processes which

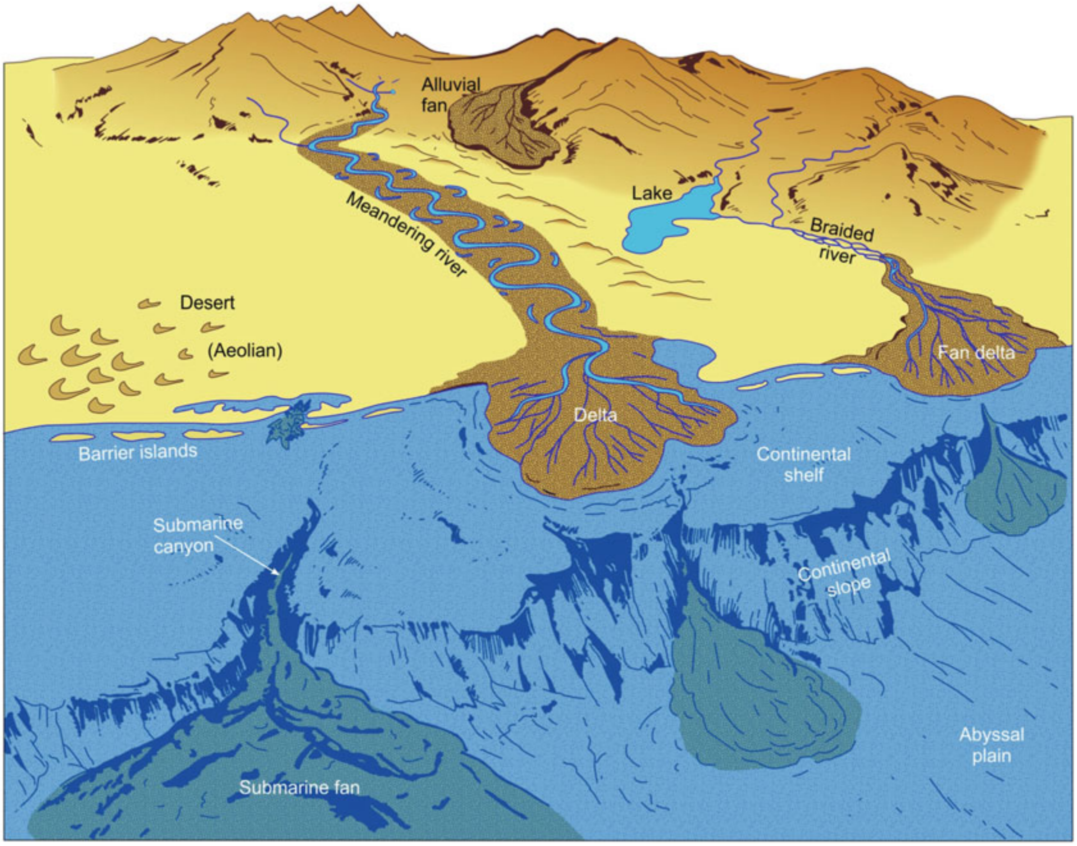


Fig. 2.23 Schematic representation of sedimentary facies on a passive margin

have acted in the environment in which the rock was deposited. The sedimentary structures we observe tell us something of the hydrodynamic conditions during deposition. Organic structures and fossil content help us reconstruct the ecological conditions at the time of deposition in the basin. Evidence of chemical processes such as weathering, diagenesis and precipitation of authigenic minerals also supply important information about the depositional environment, as well as its early post-depositional history. Figure 2.23 is a diagrammatic representation of the major depositional environments on a continent and the various transition stages to deep water.

Following an interpretation of the facies of each bed or sequence of beds, we may try to group these into what we call *facies associations*. These associations of facies can be related to large-scale processes in particular environments. An association of facies coarsening upward from marine shales into delta front sand or into coal beds of the delta plain facies may represent a facies association typical of delta progradation.

Other facies associations may represent transgressions or regressions in a shallow marine environment.

If we have a continuous sequence without breaks (unconformities) the vertical succession represents the lateral succession of environments. This, in essence, is Walther's Law, named after the German geologist Johannes Walther.

2.24 Alluvial Fans

Alluvial fans are accumulations of sediment which has been transported by fluvial processes or different forms of mass transport (e.g. mud flows and debris flows) and deposited in valleys or on slopes fairly near the erosion area. The fans form in areas that have considerable relief and are usually associated with faults which keep uplifting the erosion area in relation to the area of deposition. This tectonic subsidence of the alluvial fans is a necessary condition for their being preserved in the geological series. Alluvial fan deposits in older rocks may therefore supply important information about tectonic movements during deposition. Alluvial fans which have formed in areas with a high relief but little tectonic activity, will merely represent a stage in the transport of the sediments. For example, in glaciated areas we may find alluvial fans on slopes steepened by glacial erosion. They will be eroded again if they are not rapidly covered by a transgression.

Erosion on the uplifted block will form V-shaped valleys (or canyons), and these will drain into the valley. The apex of the alluvial fan is usually near the main fault plane and sediment transport from here will tend to follow the steepest slope downwards so that the sediments will be spread out in a fan (Figs. 2.24 and 2.25). If there is only a short distance between adjacent valleys, and consequently between fan apices, the fans will coalesce. If major drainage systems develop, larger fans will form further apart from one another. In areas with a relatively humid climate, fluvial processes will account for sediment transportation even high up on the fan. In arid climates the water table under the fan will be deep down, and when it rains the water will rapidly filter down into the upper part of the fan. The slope of the fan may then increase due to deposition on the upper part. As a result these sediments may be transported with a high sediment/water ratio as debris flows or mud flows. The upper part of the fan may consist of large blocks or cobbles which form an open network system

Fig. 2.24 Alluvial fan. Death Valley, California

through which finer-grained sediments can pass. In this way the sediment is sieved, and *sieve deposits* are formed.

Downslope on the fan, channels usually split into a number of smaller channels. This reduces the hydraulic radius of the channels and in consequence their velocity, and hence capacity for carrying sediment, is lowered. Sediment will therefore become finer-grained downslope, even if there is no reduction in the gradient.

The water table will be deepest at the top of the fan, and shallowest at the foot. Alluvial fans are good groundwater reservoirs, easy to tap because of their porosity and permeability. The circulation of groundwater through an alluvial fan leads to strong oxidation of at least the upper part of the sediments, giving them a red colour due to iron oxides. In most cases any organic material will be completely oxidised.

In arid climates there will be a great deal of evaporation from the groundwater which emerges from the fan, and the ions in solution will be precipitated as carbonate *(caliche)* and iron oxide. Water flows beyond the arid alluvial fans in *ephemeral rivers*, which only exist after heavy rain, and may then collect in *playa* lakes which dry up each year, forming evaporite deposits (Fig. 2.25).

Humid fans will be dominated on the lower part by fluvial channels. These may drain a major portion of the fan, and thus have a large hydraulic radius and a greater capacity for transporting sediment, even if the slope of the fan is not very great. In the lower part of a humid fan cross-bedding will be fairly pervasive, while the upper part will tend more to consist of massive conglomeratic beds. The foot (distal end) of the fan will merge into the other sediments which cover the valley floor. These may be lacustrine or fluvial deposits.

A characteristic feature of transport and sedimentation across and around arid zone alluvial fans is that sedimentation takes place during short periods in connection with the rains. However, the water rapidly disappears into the ground, increasing the surface sediment/water ratio, and eventually dries up to leave poorly sorted conglomeratic beds. It is important to remember that the sedimentary structures and sorting we observe are only representative of the final deposition phase.

The grain size of the sediments on the fan is a function of weathering and erosion in the source area

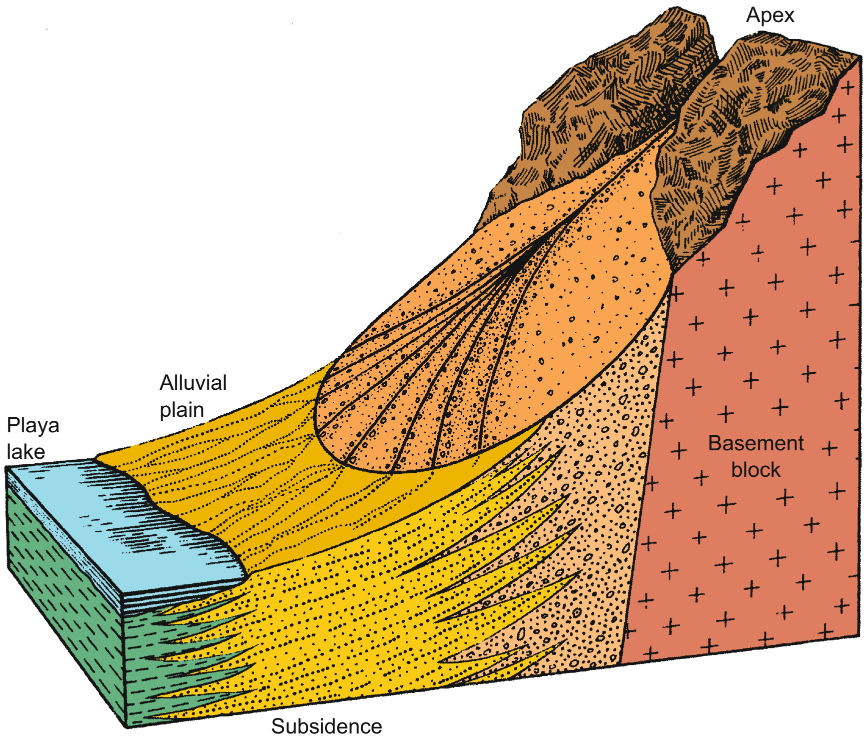


Fig. 2.25 An alluvial fan developed along a basement fault. The degree of progradation of the fan into the basin depends on the relief along the fault, but is also dependent on the rainfall and the catchment area on the upthrown block

and of transport capacity outwards across the fan. With fluvial transport, the velocity *v* and thereby transport capacity is both a function of the slope and the depth (hydraulic radius) of the channels (see Chezy equation, p. 42).

The velocity is therefore much higher during flooding.

2.24.1 The Water Budget

A part of the total rainfall will sink into the ground and supply the groundwater. Another part will evaporate on the surface or flow on the surface into rivers (Fig. 2.26). In hot, desert areas where the air is very dry, evaporation may correspond to 2–3,000 mm/year, i.e. more than the normal amount of precipitation in most places. Plants also contribute to water loss through transpiration, and in areas with vegetation, evaporation and transpiration may reach up to 2,000 mm/ year. Without vegetation we find a significant amount of evaporation only when there is free water or damp ground right up to the surface. Even if the water table is shallow (less than 1 m), there is little direct evaporation, but some water will be drawn up by capillary forces and evaporate.

Plants are important in connection with evaporation of groundwater as tree roots may penetrate more than 10 m below the surface. Trees can be a major drain on groundwater reserves because they use water which could otherwise have seeped into wells, rivers or lakes. Dense forests may use water corresponding to 200–500 mm of precipitation. In areas with limited water reserves it may consequently be necessary to limit vegetation, particularly of varieties of trees and bushes with a relatively high transpiration rate and no useful function. However, vegetation is important for stabilising the topsoil and thereby preventing erosion. Much of the water evporated from vegetation may return as rain locally. Vegetation may also have an effect on the *albedo*, i.e. the amount of sunlight which is reflected from the Earth's surface, and this in turn may affect the climate.

Runoff is the fraction of precipitation which reaches the rivers. The percentage of precipitation which filters down to and recharges the groundwater is of particular

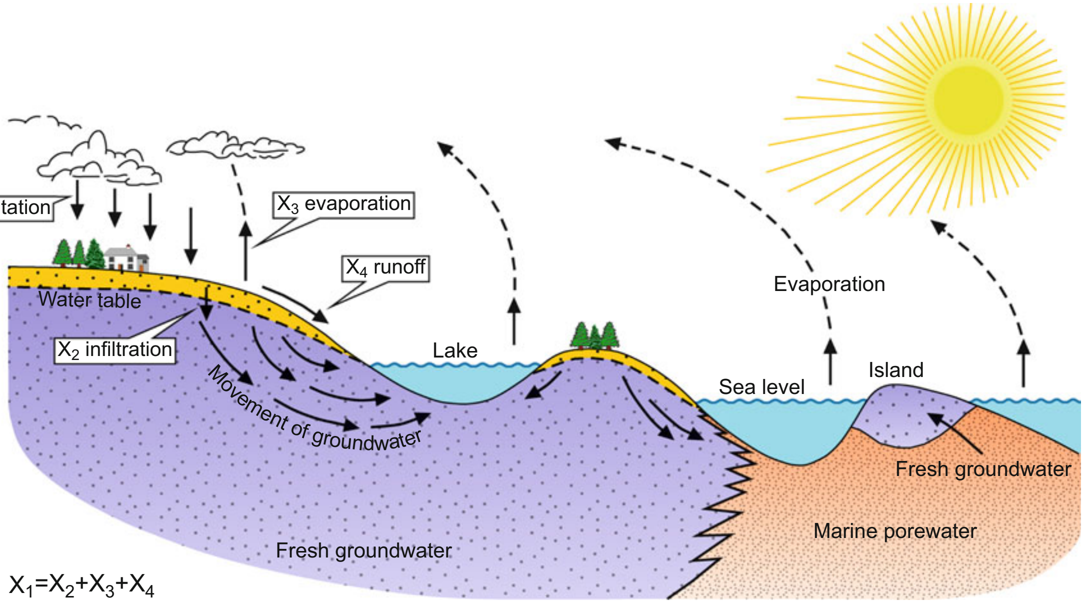


Fig. 2.26 Diagram showing the circulation of groundwater on land and under marine basins

importance. Rivers also have groundwater added to them if their surface is below the water table. If their surface is above the water table they will lose water. In consequence, rivers cannot be considered in isolation from groundwater reserves.

Infiltration of surface water into the ground depends on the permeability of the soil cover and on the vegetation. Clayey sediments allow far slower infiltration of surface water than sand or gravel. Where there are soil types with swelling clay (smectite group), which are very common in desert areas, the first rains will cause these minerals to swell, further reducing their permeability. Consequently, infiltration is reduced, and the runoff which creates floods increases. If the earth is dry above groundwater level, water has to overcome the capillary forces which act against water percolating down through dry, fine-grained sediments. We may find a layer with air in the pores between the groundwater table and the water filtering down from above (trapped air).

Alluvial fans are porous and have a relatively good potential as water reservoirs, and we often find springs at the foot of fans. In dry areas the depth to the water table can be most simply charted through seismic recordings.

Because freshwater is lighter than saltwater, it will flow over saline porewater and form pockets of freshwater under islands. Freshwater may also flow from the continents, following permeable beds beneath the continental shelf. Freshwater of high quality (little pollution) is scarce in many regions, and in recent years in particular a large number of geologists have been involved in mapping groundwater reserves.

2.25 Desert

Deserts are areas with little or no vegetation and we can distinguish between different types of desert:

1. Deserts due to low precipitation and high evaporation in hot climates.
2. Deserts due to low precipitation in cold climates.
3. Deserts due to soil erosion.

The first two types of desert are largely governed by meteorological factors. Areas which lie on the lee-side of major mountains will be dry. At about 30°N and 30°S high pressure areas predominate, resulting in low precipitation. Since the prevailing wind at these latitudes will come from the east, we find most of the deserts on the western side of mountain chains. This is true of Asia Minor, the Sahara, California and Nevada in the northern hemisphere, and Australia, South Africa and Chile in the southern hemisphere. Inland areas surrounded by mountains tend to have a desert climate. During the glacial periods there was less evaporation and therefore less rainfall. North Africa and the Sahara were then very much drier than the present day.

Iceland has good examples of desert regions which are due to a cold climate, and has problems with soil erosion due to lack of vegetation in many areas.

Shortly after the withdrawal of the ice across Scandinavia there was a desert with considerable aeolian deposits, which formed before the vegetation cover developed. Loess deposits are fine-grained sediments which form largely through aeolian erosion and transport from deserts and also from glacial sediments exposed after glacial retreat. Large areas of China, for example west of Beijing, are covered by up to several hundred metres of Quaternary loess.

The lack of vegetation naturally means that aeolian transport and deposition are important in deserts, but only a relatively small part of the desert areas are covered by wind-blown sand.

Large areas consist of bare rocks and mountains with little sediment. In other regions there is only wind erosion (deflation), which leaves the ground covered *(armoured)* by a layer of stones which protect it from further erosion. Even in desert regions like the Sahara many areas are dominated by a fluvial drainage pattern. Although several years may pass between rains in this area, a heavy rainstorm may transport so much sediment that the fluvial drainage pattern survives for many years.

Large expanses of wind-blown sand are called *ergs*. They may be formed by the coalescence of different aeolian bedforms. *Barchans* are crescent-shaped isolated aeolian dunes with a convex erosion side and a concave lee side (Fig. 2.27). Barchans are found mainly on the edges of the erg area, where there is not enough sand to form a continuous thick cover. Transverse dunes are common within ergs. *Seif dunes* are long, straight sand dunes which may occur within an erg but are also found in areas with incomplete sand cover. In the central zones very large dunes form which may be over 100 m high and 1 km long.

Ancient aeolian deposits are recognised by:

1. Their large-scale cross-bedding, up to 20–30 m and often with wind ripple marks on top of slanting cross-bedded surfaces (foreset beds).
2. High degree of oxidation which gives a red colour.
3. Good sorting – largely medium- to fine-grained sand.
4. Lack of fossils or organic material.

The lack of thin silt or clay beds between the cross-bedded layers, and the general association with overlying and underlying sediments, are also important criteria. One of the best-known examples of an aeolian sandstone is the Navajo Sandstone (Jurassic) in the Rocky Mountains. In Northern Europe there are aeolian sandstones in the Lower Permian (Rotliegendes) and the equivalent Yellow Sand in northeastern England (Fig. 2.27).

In pre-Devonian times all continental environments were deserts in the sense that they lacked vegetation. Also in these rocks, however, we can distinguish between dry and wet climates, largely from the nature of the fluvial or aeolian deposits.

Sand dunes are always moving and may end up in the sea, still partly preserving an aeolian sorting and grain-size distribution.

2.26 Lacustrine Deposits

How Are Lakes Formed?

In principle we have three types of lake:

1. Lakes of glacial origin:
   (a) lakes formed through glacial erosion,
   (b) lakes formed through damming by moraines or by the glaciers themselves.
2. Lakes of tectonic origin:
   (a) lakes formed in areas of rapid tectonic subsidence (rifting) or more uniform subsidence,
   (b) lakes formed as a result of damming by horsts which have been elevated through faulting, or damming by lava etc.
3. Lakes formed by sedimentary processes, e.g. *oxbow lakes* in fluvial environments, and delta top lakes.

Lakes of glacial origin will have a relatively short lifetime by geological standards. In 10,000–100,000 years most of the lakes in Scandinavia and North America will have filled up with sediment, if we do not have another glaciation.

Lakes of tectonic origin, however, will continue to subside. If the rate of subsidence keeps pace with the rate of sedimentation, a lake will continue to exist.

Extensive carbonate beds (freshwater carbonates) and diatom deposits are also common in lacustrine basins. Some of the largest lakes formed by rifting are found in East Africa, where a number of lakes occur in

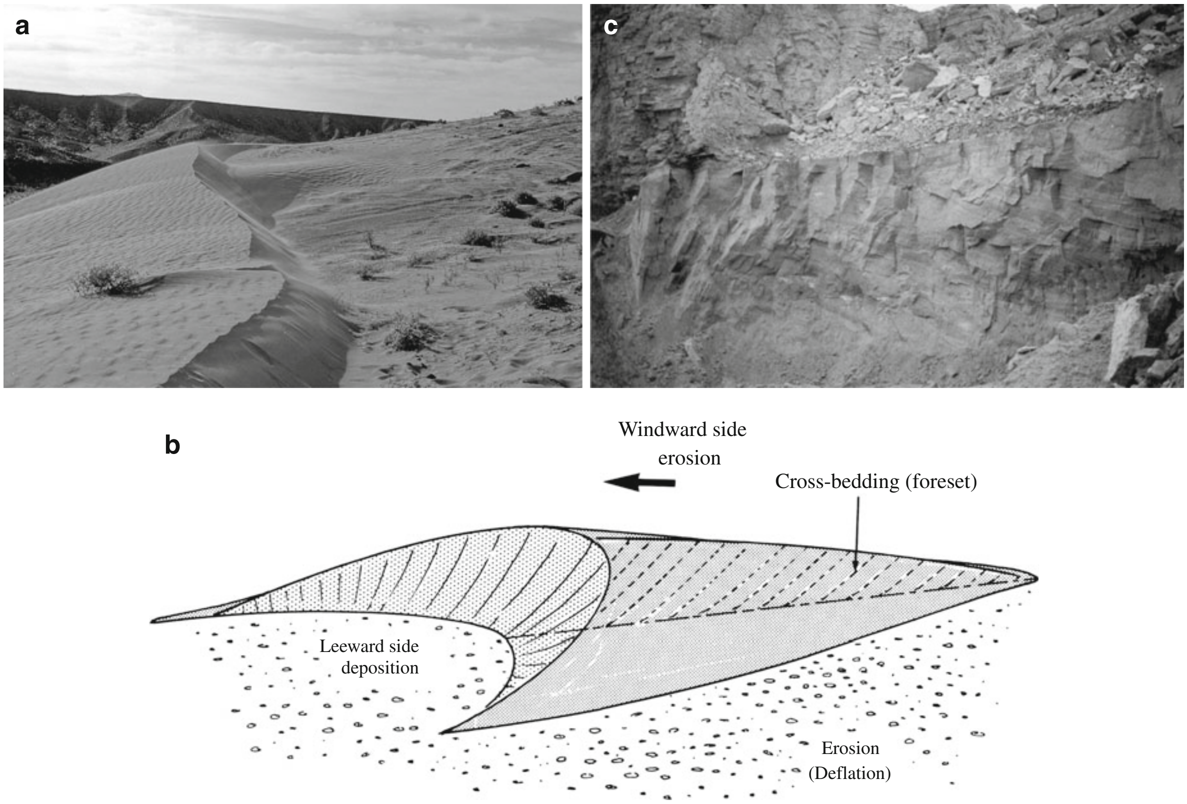


Fig. 2.27 (**a**) Aeolian dune from the desert in Death Valley, California. (**b**) Schematic representation of an aeolian dune (barchan). (**c**) Aeolian cross-bedding 4–5 m high in the Permian "*Yellow Sand*" of northeast England (Old Quarrington Quarry, Durham). This sand is equivalent to the "Rothliegendes" sandstone of the North Sea

the actual rift valley system. Lake Victoria, which lies between two rift valley systems, was formed by tectonic movements about 100,000 years ago.

In humid climates all lakes will have outlets in the form of rivers or via the groundwater, but in arid climates lakes develop into evaporite basins without outlets. Normal, non-saline lakes differ from ocean basins in several ways:

1. Low salinity leads to slower flocculation of clay particles than in marine basins, producing more distinct lamination.
2. Low wave and tidal energy and weaker currents mean less erosion and resedimentation. Seasonal variation in the influx of sediments may produce annual cycles in lacustrine sediments.
3. Limited water circulation makes it easier for the water to develop layering based on temperature (thermocline) and therefore also density (pycnocline) between dense layers at the bottom and less dense water near the surface. This may restrict oxidation of organic material and promote the development of organic-rich sediments (source rocks) in the deeper parts of the lakes. In warm climates the temperatures of the lake waters are always above 4°C which makes the water stratification rather stable. In colder climates the water column will be inverted when the temperatures falls to 4°C which is the highest density. This results in an oxygen supply to the lake bottom.
4. River water will generally have approximately the same density as lake water, and will therefore mix well and deposit sediments rapidly. Cold (glacial) river water or river water with a lot of suspended material will be heavier than lake water, however, and will form turbidity currents along the bottom.
5. Lake sediments have a distinctly different fauna from marine basins. The geochemistry of lake sediments and the composition of carbonates and

evaporites are also different from those of marine sediments.

By comparison with marine deltas, lacustrine deltas are among the most constructive of all (e.g. the Gilbert Delta) because erosion in the basin is so limited. Nevertheless, wind conditions, the size of the lake and the composition of the sediments will be crucial factors. In large lakes we might have relatively high wave energy which could erode sediments from the river mouth and deposit them laterally on beaches.

Wind will also cause water to well up from the bottom, increasing the circulation of oxygen along the bottom.

In dry regions we may get evaporitic lakes and lakes which dry up after each flood *(ephemeral lakes)*.

Lake deposits are characterised by:

1. Lack of marine fossils, but diatoms and algae may be important components.
2. Very fine laminations in clay and silt sediments.
3. Bioturbation structures may be found in lacustrine sediments, but they are less common than in marine sediments.
4. Chemical analyses of freshwater sediments will reveal low concentrations of the trace elements which are enriched in seawater (e.g. Cl, Br and B). Minerals formed in freshwater lakes may have characteristic isotopic compositions. Evaporites in lakes consist mainly of carbonate minerals; sulphates and chlorides are normally less common.
5. The content of sulphur-bearing authigenic minerals must be relatively low since there is little sulphate in freshwater. However, organic material in lakes contains a fair amount of sulphur which can be precipitated as sulphides by sulphate-reducing bacteria. Sulphur may also be derived from volcanic sources.
6. Lack of tidal structures and similar marine structures is also an important indicator.

Amongst the most important ancient lacustrine deposits are the Karoo deposits of South and East Africa, from the Carboniferous to the Jurassic periods. These sediments contain important coal deposits. In eastern China there are large areas of Cretaceous and Tertiary lacustrine sediments, which are also important petroleum-bearing sediments.

The Green River Formation (Eocene) of Colorado, Wyoming and Utah, is a lacustrine sediment of great extent which represents one of the most important petroleum source rocks in the world.

Rifting associated with the opening of the Atlantic Ocean in Mesozoic times resulted in a large number of lacustrine basins, and sediments deposited in such basins now underlie the continental shelves. In dry regions these turned into evaporite basins.

2.27 River Deposits

Transport of sediments in rivers depends on the gradient and cross-section of the river, which determine the flow rate, and on the composition and concentration of sediment. In rivers which flow only during and directly after the rains, but are otherwise dry *(ephemeral streams)*, equilibrium between flow and sediment load is not achieved, so that even fine-grained sediment is deposited in the channel when it dries up, leaving a mud-cracked surface. Intensive oxidation of sediments, including silt and clay beds, is typical of such fluvial deposits. There is little organic production in areas with a dry climate, and clay and silt particles will therefore contain little organic material which could act as a reducing agent. A low water table during dry periods also contributes to oxidation of organic matter.

Most rivers have marked seasonal and annual flow variations. The flow velocity increases appreciably with flow volume because the friction per unit volume of water is inversely proportional to the water depth. When water flow is greater than the capacity of the channel, the water flows out over the banks. The water outside the channel will normally be very shallow and have a low flow velocity and fine-grained sediments, which have been transported in suspension, are then deposited as what we call *overbank sediments*. Overbank sediment may build up into elevated banks called *levees*. The resulting soil along modern rivers is often very rich and ideal for cultivation. In post-Devonian sedimentary rocks we often find traces of plants, commonly roots, in levees. Levee deposits have parallel lamination and in some cases also current ripples, particularly *climbing ripples*, which are typical of rapid sedimentation. The primary sedimentary structures, however, will often be partly or wholly destroyed by traces left by plant roots.

Major floods may cover the areas beyond the levees as well, and clay and silt will be deposited on these *flood plains*. Many rivers which carry a large amount of suspended sediment gradually build up their beds through deposition in their channels and on the levees, so that the surface of the water in the channels may be considerably higher than the surrounding plain. During floods the water may then break through the levee and flow down from the channel to the plain, forming temporary lakes several metres deep. The lower Mississippi River is considerably higher than the surrounding area, including New Orleans. The major Chinese rivers have also built themselves up above the surrounding countryside, so floods can cause very great damage.

However, the stability of the levees will determine how much a channel can build itself up, and clay-rich sediments form stronger levees than sand and silt because of their cohesiveness. When the levees give way, sand will flow out of the channel and deposit sandy sediments in fans or *crevasse splays* on the flood plain. They are characterised by thin sand beds, often fining upwards, with an erosional base in the proximal part (nearest the channel). However, we may also find small sequences which coarsen upwards as a result of progradation of this fan, because grain size decreases towards the edge of the fan.

2.27.1 Channel Shapes

We distinguish between relatively straight fluvial channels and curved channels. Curved channels are shaped rather like a sine curve. To describe the degree of curvature we speak of high and low sinuosity. Sinuosity is defined as the ratio between the length along the channel and the length of a straight line through the meander belt, i.e. the length of the river valley. *Meandering streams* have a single channel and by definition a sinuosity greater than 1.5.

Braided streams have a branched course, but in most cases the river channels are fairly straight. These rivers are branched because the river channel is not very stable, and because sediment is deposited in the middle of the channel, forming small islands or bars. Braided streams typically develop in coarse sediments containing little clay or silt which can form stable levées. A higher stream gradient gives greater energy and greater erosion of the sides of the channel. An abundant supply of sediment leads to sediment being deposited more rapidly than it can be eroded and transported onwards, resulting in the formation of deposits in the channel. Braided streams are therefore also typical of areas where the velocity declines somewhat, e.g. when a river widens onto a plain after passing through a narrow valley. The velocity and flow will in most cases vary considerably with time. When the river is low it will flow round bars of sand and gravel, while at high water and with strong currents sand bars will migrate as they are eroded on the upstream side and accumulate deposits on the downstream side. Gravel bars, on the other hand, have rather complex patterns of erosion and deposition.

Meandering rivers have a wavelength which is a function of the breadth of the river. The wavelength is approximately 11 times the breadth or 5 times the radius of the river. There is also a relatively regular ratio (about 7:1) between the breadth and depth of the river.

Meandering rivers move in loops, with the greatest velocity at the outer bank so that erosion is concentrated there. The velocity along the inner bank is much lower, so sediment is deposited there. The flow velocity and shear forces against the bottom also decrease upwards towards the top of the bank on the inner side of the bend, and sediments are deposited in *a point bar* which reflects the hydrodynamic conditions. At the lowest point there is a lag conglomerate followed by large-scale cross-bedding due to the flow being in the upper part of the lower flow regime. Then come current ripples, and finally fine-grained sand, silt and clay sediments corresponding to the lower part of the lower flow regime at the top of the profile *(overbank sediments)*. This produces a fining-up sequence from sand to silt and clay (Fig. 2.28).

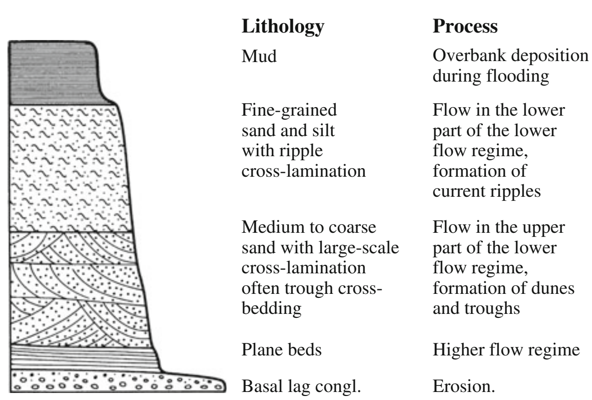


Fig. 2.28 Fining-upwards point bar sequence deposited by a meandering river

The secondary flow, in a vertical section at right angles to the downstream flow direction, moves from the outer bank where erosion takes place along the bottom, and up the inner (point bar) bank where deposition takes place. This is a result of a difference in hydrostatic pressure because the surface of the water slopes inwards towards the inner bank due to centrifugal forces. If we combine this movement with the main flow of water down-river, we find a corkscrew or *helical* movement (Fig. 2.29). At each bend in the river the helical flow reverses direction. The point bar becomes asymmetrical, with coarser material on the upstream side, so that a perfect fining-upward profile is not developed. The upper fine-grained part and the overbank deposit will also be lacking. Variations in the depth of water in the channel will also lead to departures from the ideal fining-upward sequence.

During floods the water may flow over the point bar and form a little channel, or *chute*. This may be widened by further erosion to become the main channel. This process is called *chute cut-off*. The sinuosity may also become so high that erosion cuts a channel through a narrow neck (neck cut-off), making a straight course to a lower bend in the river. The whole meander will then be abandoned by the river, and a small, curved *oxbow lake* is left, which will fill with clay, silt and organic matter.

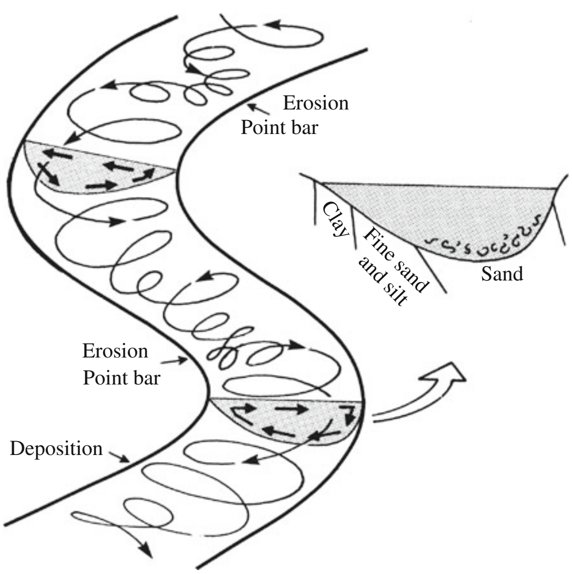


Fig. 2.29 Diagram showing how the flow in a vertical section reverses direction in each bend (helical flow)

Rapid subsidence and low sand/mud ratios will increase the stability of the channels and we may find an anastomising channel distribution where there is little or no lateral acretion of the channel (Fig. 2.30).

When the whole river channel shifts course (avulsion) the abandoned channel will fill up with mud and form a clay plug.

2.28 Summary of Fluvial Sedimentation

Fluvial processes are fairly simple in principle. We know the physical laws which govern the flow of water in channels and transport of sediment in water. The great variation in composition, structure and geometry demonstrated by fluvial deposits is due to all the variables which influence transport and sedimentation. As we have seen, the most important factors are:

1. Climate, particularly precipitation and seasonal distribution of precipitation
   (a) in catchment areas
   (b) along river valleys – vegetation stabilises river banks.
2. The drainage area
   (a) size
   (b) type of rock being supplied
   (c) topography – tectonic uplift.
3. Subsidence of the alluvial plain.

In order for thick fluvial series to be deposited and preserved, the fluvial plain must be located in a tectonically subsiding area.

Catchments with a lot of shale and other fine-grained sedimentary rocks will produce sediments with a high clay and silt content. The products from weathering of eruptive rocks will contain a considerable amount of clay, mainly kaolinite, illite and smectite, which increases the cohesiveness of the sediments and stabilises the fluvial channels. The climate along the river plain may be very different from that in the drainage area but some vegetation can be supported by the groundwater close to the river also in dry climates. In humid climates the vegetation will help to stabilise the river channel and reduce the velocity of the floodwater outside the channel.

The flow of rivers on the river plain will also depend on the groundwater table. The river will contribute water to the groundwater if it is lower than the surface of the river, and groundwater will flow into the river if the reverse is the case.

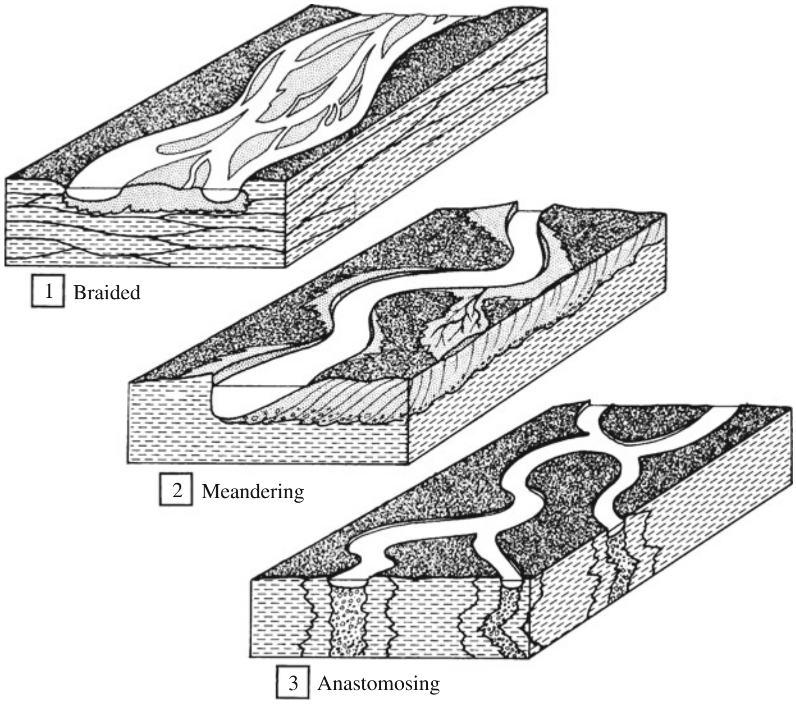


Fig. 2.30 Different types of fluvial channels. The type of fluvial channel determines the distribution and geometry of sand bodies and depends on the sediment composition and slope variations in the water flow

2.29 Delta Sedimentation

Large deltas require drainage areas with sufficient precipitation to produce high runoff. They are formed along passive plate margins where extensive drainage systems can unite into large rivers. The greater part of the drainage from North and South America and Africa flows into the Atlantic Ocean. Only minor rivers enter the Pacific Ocean from the American continent, and relatively little of the drainage from Africa enters the Indian Ocean. The major rivers follow old drainage systems, the main features of which have existed since the Mesozoic, and which often seem to have been governed by Mesozoic rifting along the Atlantic continental margin.

As mentioned in connection with fluvial sediments, the erosion area will determine the composition and grain size of the sediments which are transported by rivers and deposited on the delta. The Mississippi, for example, drains extensive tracts of Palaeozoic and younger sediments which contain a large percentage of shale, and the river therefore transports a high percentage of clay and silt in suspension. Areas of metamorphic and acid eruptive rocks will give sandier sediments because of their quartz and feldspar contents. Deltas prograde outwards into a sedimentary basin and form a surface called a *delta top* (*delta plain*) near the lake or ocean level. The waves break against the *delta front*, beyond which is the *delta slope*.

The formation of a delta can be depicted as a battle between the fluvial development of the delta and its erosion by marine forces. We therefore distinguish first between river-dominated deltas, which prograde far out into the basin and consist largely of fluvial sediments, and deltas which are eroded more rapidly by marine forces (tide and waves) and consist largely of marine sediments (Fig. 2.31).

River-dominated deltas are often referred to as constructive deltas, since they tend to prograde more rapidly into the basin without much marine reworking. Tide- and wave-dominated deltas are often called destructive deltas.

We may therefore distinguish between three main types of deltas: (1) river-dominated, (2) tide-dominated and (3) wave-dominated. Most deltas fall somewhere between these three extremes. The Mississippi, however, comes near the river-dominated

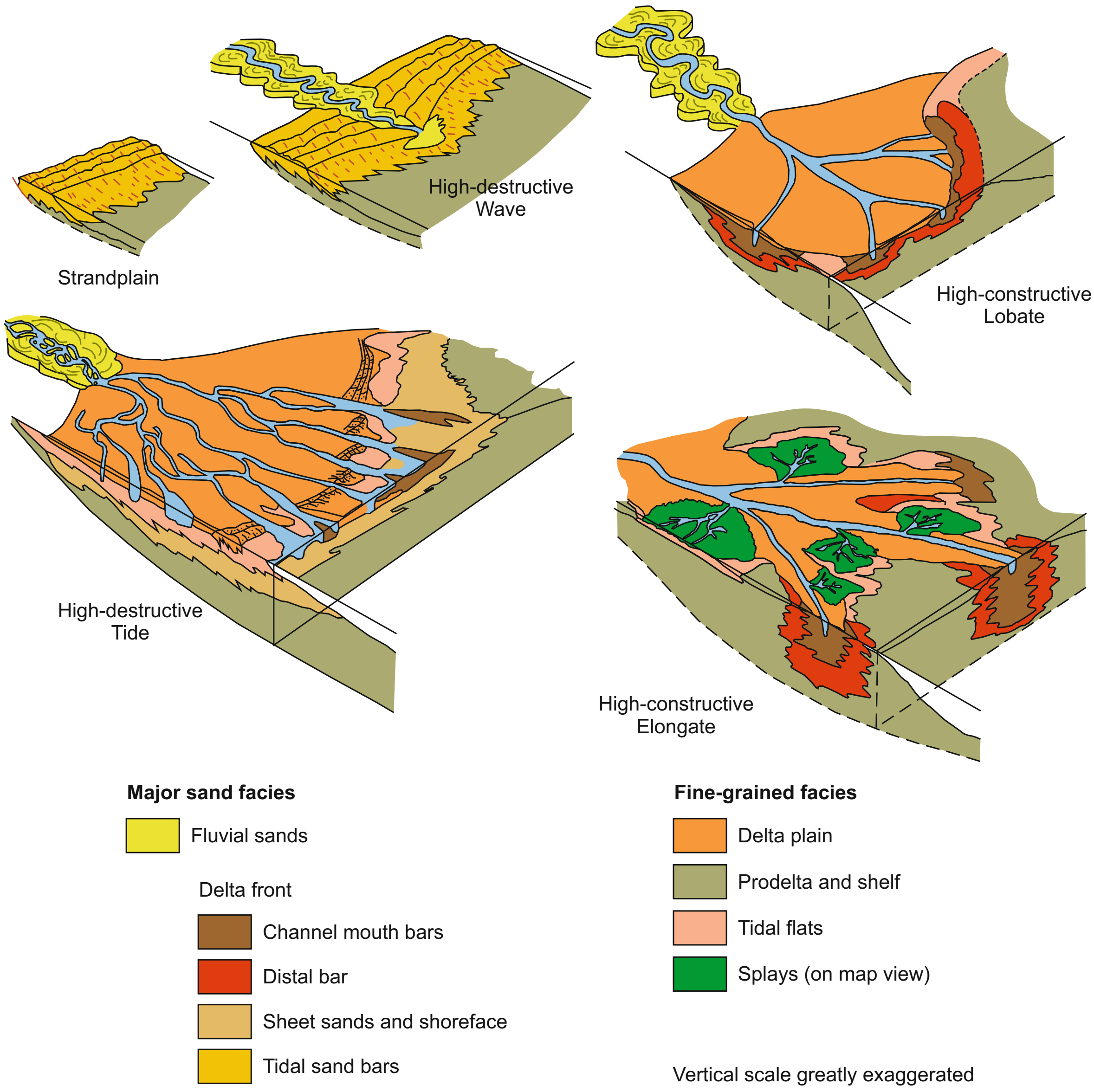


Fig. 2.31 Classification of delta types based on the relative strength of waves and tidal energy in relation to the rate of sediment input. After Fisher et al. 1974

end in a marine environment because the sediments are fine-grained and rather cohesive. The delta front is also protected from big ocean waves by a shallow shelf.

High rates of sediment supply also favour fluvially-dominated deltas. The Rhine is fairly wave-dominated as well, because the sediments are rather coarse-grained and the sedimentation rate is lower. The Ganges and Mekong deltas are typical tidal deltas.

The difference in density between the river water and the water in the marine basin plays a major role. In most cases river water will be lighter than saltwater, even if it contains a good deal of suspended material. In deltas, river water will therefore flow far out over the salt water *(hypopycnal flow)* before salt- and fresh-water mingle. If the river water and the water in the basin have the same density *(homopycnal flow)* there will be more rapid mixing of the water masses in *axial* *flow*, and the sediments will settle out of suspension more rapidly. *Hyperpycnal flow*, where the river water is denser than the water in the basin, takes place only in lakes as a rule and leads to the flow of river water along the bottom of the delta slope, with erosion of the delta front and formation of turbidites on the basin floor.

In studies of modern deltas we must also take into account the fact that most deltas are rather out of balance as regards progradation in relation to sea level because the Holocene transgression after the last glaciation raised sea level by more than 100 m. In the Niger delta we find beyond the present delta front deposits which are 12–25,000 years old, which corresponds to the last advance of the last glaciation.

2.30 River-Dominated Deltas (Mississippi Type)

The Mississippi drains a huge area of the North American continent (about 3.2×10^6 km^2) with an average precipitation of 685 mm/year. It carries vast quantities of sediment (about 5×10^8 tonnes/year) with a high clay and silt content, and the gradient of the lower part of the river is extremely low (about 5 cm/km).

In the 200–300 years during which bathymetric measurements have been taken in the area, it has been possible to record considerable progradation. The high clay and silt content gives the sediments great cohesion, making the fluvial channels relatively stable. Clay and silt sediments which are deposited on the delta plain have a high porosity and water content (60–70%), but they loose much of their porosity by compaction at shallow depth. Sand will be deposited chiefly in the channels, and in mouth bars were the channels enter the sea. Well-sorted sandy sediments, which have only about 40–45% porosity immediately after deposition, will then sink into the underlying clay because of their higher density. The fluvial channels therefore sink into the mud and this contributes to their stabilisation, with the consequence that channels change course *(avulsion)* less frequently. Because of this subsidence while sedimentation is continuing, the channel sand may be thicker than the depth of the channel. The long strings of sand which may then be preserved in the mud-rich environment are called *bar-finger sands*.

During floods large quantities of silt and clay are deposited in overbank areas between channels, and these sediments are stabilised by vegetation. In the Mississippi delta, sedimentation is very rapid so that the channel and the levees build up above their surroundings. This means that the average gradient of the channel decreases. Sooner or later the channel will have to find a new, shorter route (through avulsion) to the ocean, and which therefore has a somewhat greater slope. At any given time most of the sediments are deposited in one delta lobe prograding into the ocean until the slope becomes to low. A major avulsion will then start the formation of a new delta lobe.

We can see that the Mississippi has constantly shifted course, and has consequently been a focus of sedimentary activity in historic as well as modern times (Fig. 2.32). The present course has extended the modern delta far out and should have been abandoned for a shorter course towards the southwest, to the Atchafalaya basin. However the flow in this direction has been artificially limited because of its importance for transport to and from towns like New Orleans which lie on the present channel.

When a delta lobe is abandoned, it slowly subsides because of compaction and tectonic subsidence, while sedimentation takes place elsewhere. The abandoned lobe may sink below sea level before the fluvial supply returns to this part of the delta and another delta lobe is deposited in the same area, allowing intervening deposition of thin marine beds. Thin layers of carbonate or shale represent periods of local transgression *(abandonment facies)* which are time equivalent with regressions in the prograding delta lobes. Between the delta lobes in the interdistributary bay facies wave energy is very low and there may be little or no beach (sand) deposits between the marine mud and mud deposited above sea level.

In vertical profile we observe alternations of fluvial channel sediments, levee deposits, crevasse splays and possibly marine sediments (Figs. 2.33 and 2.34).

Even small variations in sea level have a strong influence on delta sedimentation. In a section where a fluvial facies gives way to a marine bed, it is often difficult to know whether this represents a transgression due to a rise in sea level or whether this part of the delta was abandoned by the fluvial system and is subsiding. Only if we can correlate a transgressive bed over the whole area can we assume that it is due to changes in sea level. Transgressive carbonate or thin sandstone beds are the most useful for regional correlation.

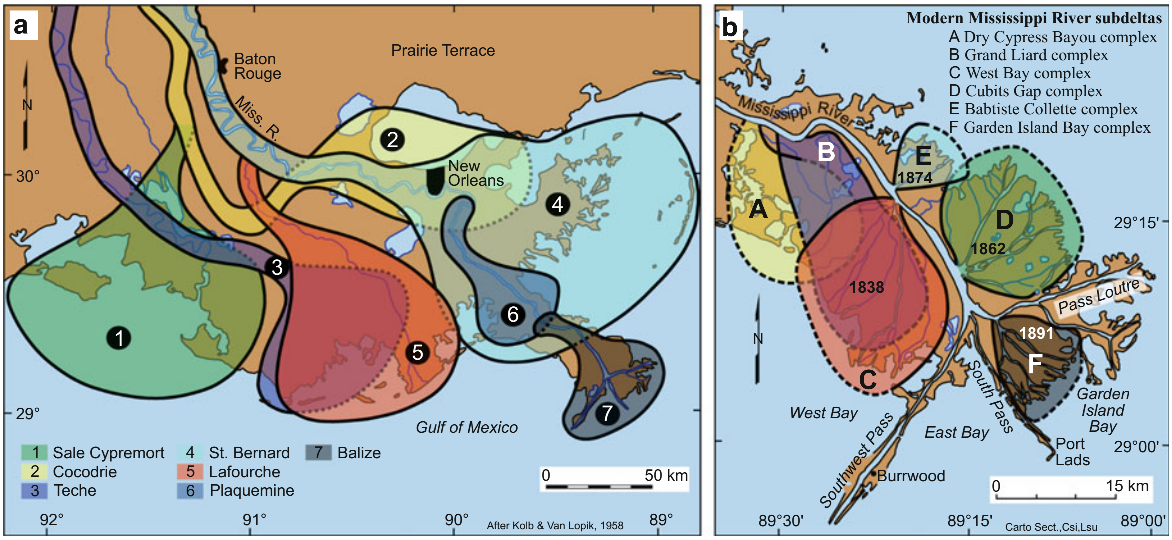


Fig. 2.32 Sedimentation in the modern Mississippi delta. When the river water breaks through the levees, crevasse channels and splays are formed, which help to fill the areas between the channels (Coleman and Prior 1980). (**a**) Delta lobes which show how the sedimentation has changed during the past 7,000 years. Each lobe of the delta appears to be active for 1,000–1,500 years (Coleman and Prior 1980). (**b**) Sedimentation in the last few 100 years

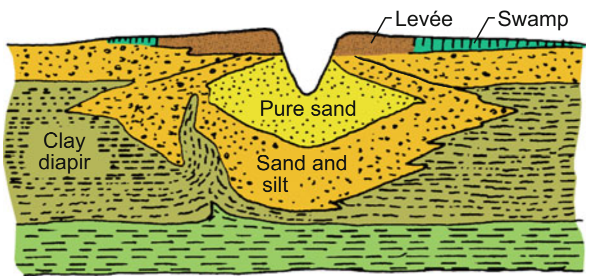


Fig. 2.33 Fluvial channels in a clay-rich delta environment. The channel sand is denser than the clay and will sink into the underlying mud. Clay diapirs may form as a result

2.31 Delta Front Sedimentation

At the channel mouth the fluvial water flow rapidly loses its energy, and all the sand which has been transported along the bottom (the *bed load)* is deposited in the form of a *channel mouth bar* (Fig. 2.34). Most of the suspended material is also deposited fairly rapidly, at a relatively short distance from the mouth. While the river water, which is lightest, flows out of the channel, salt and brackish water flow back in along the bottom, and may encroach a long way up the channel when the rate of fluvial flow is low. Marine fossils can therefore be transported quite a way into the fluvial environment. During floods the saltwater wedge is forced back over the bank at the mouth. The wedge reduces the cross-section of the river, and the velocity increases somewhat in consequence (Fig. 2.35). Spring tides will force the river water up the channel and may cause it to overflow its banks.

The channel mouth bar itself is a very characteristic deposit which grades upwards from delta mud to well-sorted sand on top. Because of brackish water and the high rate of sedimentation, there are few organisms which live right at the mouth of the channel, but bioturbation may occur in the surrounding sediments.

The delta front is the area where fluvial and marine forces meet, and we can virtually quantify the marine influence on the delta. Tidal forces are a function of the tidal range, and from wave measurements we can

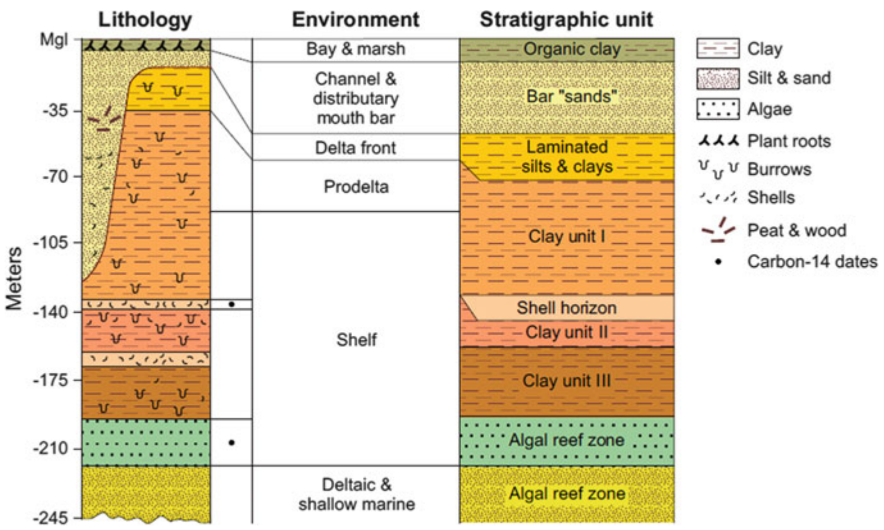


Fig. 2.34 Vertical section through modern Mississippi delta sediments. From Coleman and Prior (1980)

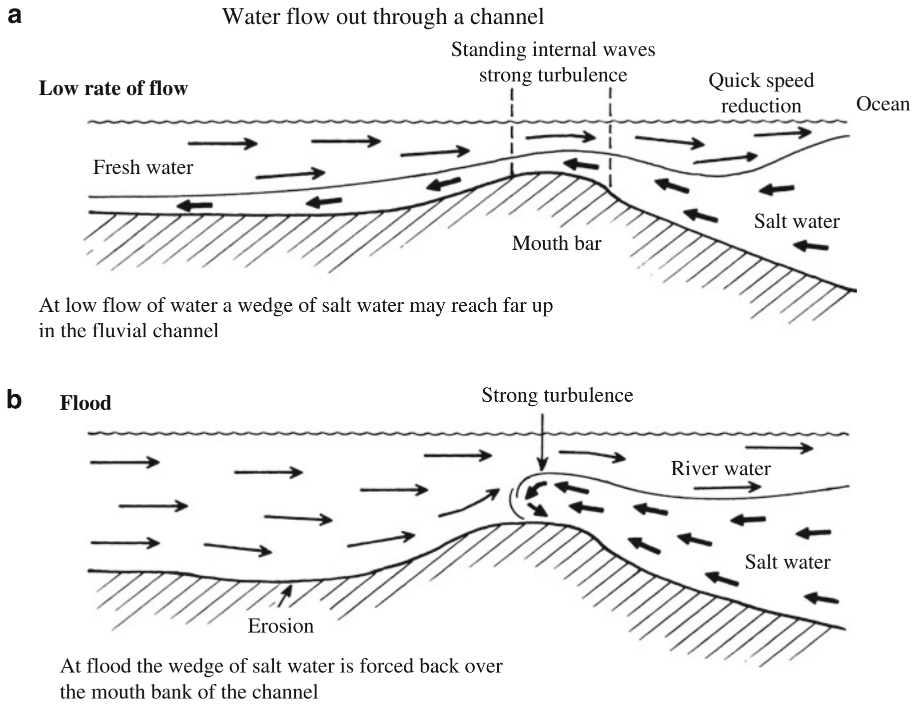


Fig. 2.35 Flow of fresh and salt water at the mouth of a river channel (after Wright and Coleman 1974)

also estimate the *wave power*. While wave power near the beach is estimated to be 0.034×10^7 erg/s on average for the Mississippi delta, it is 10×10^7 erg/s for the Nile, and 20.6×10^7 erg/s for the Magdalene River (Wright 1978). In rivers like the Mississippi, which has a shallow profile beyond the delta front, most of the wave power dissipates before it reaches the delta front. Deltas which develop into deep water right by the continental slope are exposed to the greatest wave power, because waves are not damped before they reach the delta front. The River Magdalene in Venezuela is a very typical example. The thickness and the lateral extent of the delta-front sand is a measure of wave power. Powerful storms may, however, erode vast quantities of sediment, far more than evenly distributed wave power would have achieved.

Wave-dominated delta front facies have a typical coarsening-upward sequence: from pro-delta clay to increasingly sandy sediments and finally well-sorted cross-bedded sediments with a low angle of dip, i.e. a *beach facies* profile. We often find traces of plant roots at the top of such sequences, and this shows that the sand bank has had vegetation right down to the shoreline. It was probably protected by shoreline barriers which absorb most of the wave energy. Vegetation can offer protection against both fluvial and tidal erosion, and mangrove swamps such as those in the Niger delta are particularly effective. There will always be some erosion on a delta front. Whether the delta progrades or is broken down by marine forces depends on the supply of sediment. The delta front will be fed with sediment – particularly sand which migrates from the channel mouth bar – along the beach in the wave zone. We often call this "strike feeding" because sediment transport is parallel with the strike of the shoreline, i.e. along a horizontal line. This is in contrast to the transport in channels, which is parallel with the dip of the deposit.

If there is little erosion of the channel mouth bar, the channels will extend far into the sea, and little sediment, least of all sand, will be supplied to the rest of the delta front.

2.32 Stability in a Delta

Sediments which are deposited in a delta possess very high porosity, and clay- and silt-sized grains form a very unstable structure after deposition. Clay and silt have a low permeability, however, and will expel water only very slowly. If sedimentation is rapid the sediment load will increase faster than the water can flow out, and overpressure will develop in the porewater. This means that more of the overburden is carried by the porewater, reducing the effective stresses between the sediment grains and as a result also the compaction. The friction between grains, which is a function of the effective stresses, is greatly diminished, and in consequence so is the shear strength of the sediments. If the pore pressure attains the pressure exerted by the overlying sediments, the effective intergranular stresses will be equal to zero. There is then no friction between the grains, and the sediments can flow like a liquid *(liquefaction)*. The resulting instability may cause diapirs of mud to be squeezed up into the sand bed (Fig. 2.33). The Mississippi delta is characterised by rapid sedimentation, and we find diapirs of overpressurised clays. Clay diapirs rise like salt diapirs because they are less dense than the more compact clay, silt or sand, which have lower water contents.

The stability of sediments also depends on the chemical composition of the porewater. In freshwater, clay mineral particles with negatively charged surfaces will repel one another. In saltwater these surface charges will be neutralised by cations (Na^+, K^+ etc.) so that clay minerals flocculate.

Deltas which prograde out into deep water will develop a slope which may vary greatly. The force of gravity acting on the sediments in the delta front and on the slope produces shear stresses in the sediments. When these stresses exceed the shear strength of the sediments, the sediments will be deformed by some sort of gravity-governed process. This may take place through sliding and slumping which in turn may generate turbidity currents, or steep fault planes may develop, i.e. *growth faults* (Fig. 2.36). The name was introduced during early oil exploration and refers to the fact that beds would thicken (grow) on the downfaulted side. The growth fault plane gradually deflects and flattens out with depth.

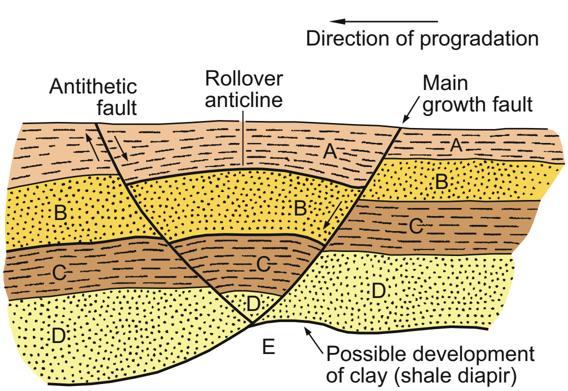


Fig. 2.36 Main features of growth faults. The rollover anticlines may be traps for oil and gas

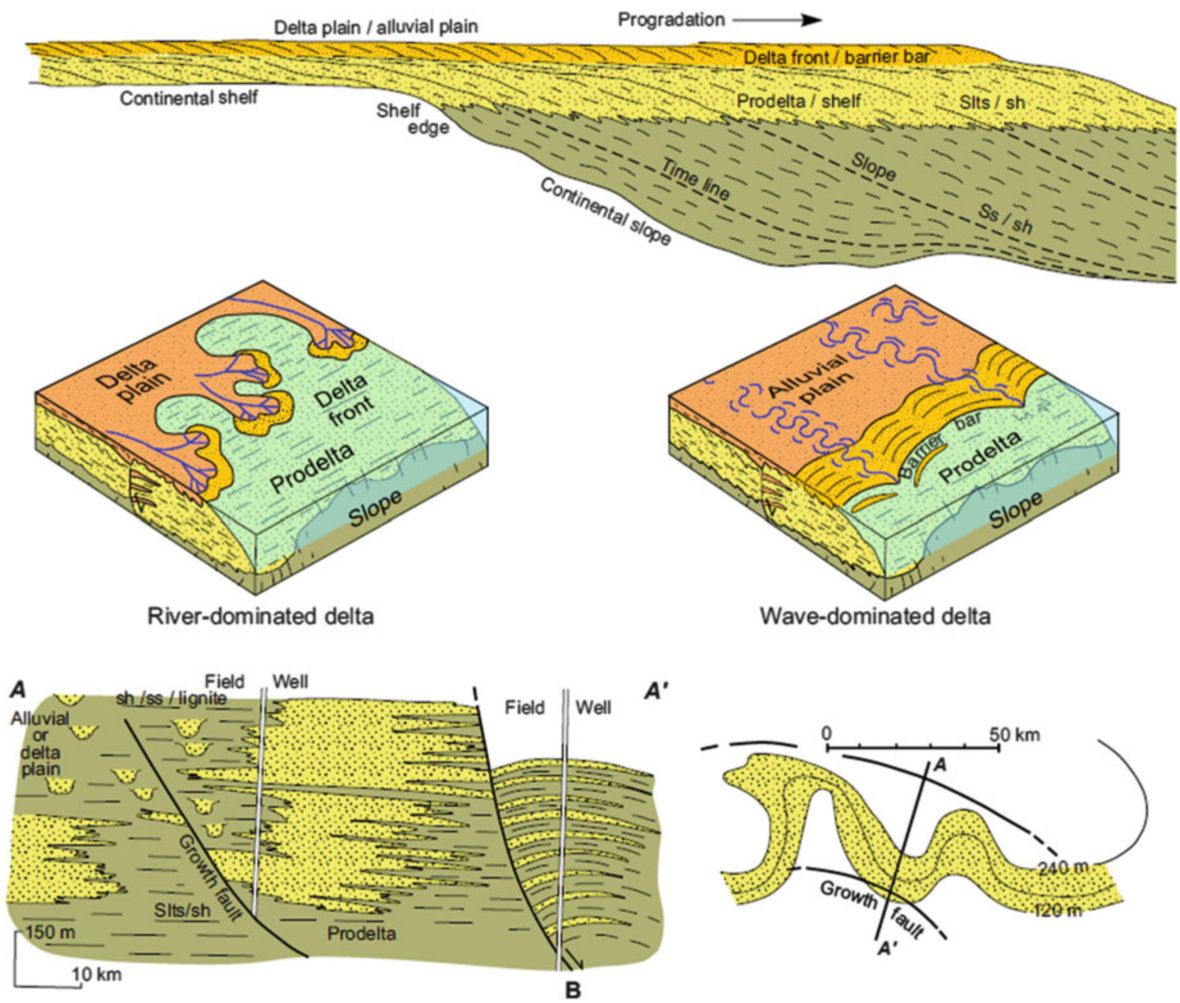


Fig. 2.37 Section through a prograding delta system showing potential traps for oil and gas (Brown and Fisher 1977)

The growth fault may form near the delta front and lead to this part of the delta subsiding rapidly, resulting in very thick deposits with delta front facies stacked on top of one another (Fig. 2.37).

2.33 Tide-Dominated Deltas

A large tidal range affects delta sedimentation very strongly and in a number of ways. The lower part of a major river has a very low gradient, so when the sea level rises and falls by 6–8 m it causes great fluctuations in the gradient, with the tidal range often affecting river flow as much as 50–100 km upstream. During flood tide the river surface gradient is reversed and a great deal of sand (bedload) and biological material will be transported up-river. The difference between ebb and flood decreases gradually up the river, and the periods of rising water (when the tide comes in) will be shorter, right down to 2–3 h. This must be compensated by higher flow velocities. Consequently it is often only the flow up the river channel during high tide which gives rise to shear forces against the bottom strong enough to transport a bedload. Sand deposited in such rivers has structures *(bedforms)* which indicate transport up the river. It is very important to remember this when studying older rocks.

Rivers in areas with high tidal ranges broaden markedly as they approach the sea because of the ever greater volumes of water to be transported out and in. At the outermost point a broad estuary forms, with clay and sand banks cut by channels which transport tidal water. The Thames and the Rhine are examples of modern rivers with well-developed tidal estuaries.

The mixing of fresh and marine water also causes flocculation of mud and deposition in the more protected parts of the estuary between sand bars. Estuarine deposits are therefore characterised by a mixture of sandstone and mudstone, often with marine fossils which may be transported several kilometres upstream by the salt wedges along the bottom. Tidal flats are often found in the inner parts of the estuary. Tidal channels are oriented at a high angle to the coastline, but in the outer part waves may produce elongated bars which are oriented parallel to the coastline. The geometry and orientation of the sandbars are very different from those associated with wave-dominated deltas and coastlines, where the sand bars are parallel to the coast.

Tide-dominated deltas are characterised by the fact that the erosion due to tidal currents is very strong compared to erosion by waves. The fluvial channels have cross-sections which are too small to allow transport of water out and in, and the result is the development of a broad belt of tidal channels, which farthest out are separated only by narrow ridges of sand and clay. Very often separate ebb and flow channels develop, and in each channel one transport direction will dominate. Signs of transportation in two opposite directions have often been used as a criterion for tidal deposits. This will apply to the whole area, but we cannot always expect to find cross-bedding with opposite current directions in the same channel deposits.

2.34 Wave-Dominated Deltas

Wave-dominated deltas have a vertical sequence very like that of a beach. Little of the fluvial part of the delta is preserved, and instead of distinct deposits at each channel mouth (distributary mouth bars), the waves distribute the sediments in a continuous *beach ridge* along the delta. The result is a coarsening-upward profile which is not always easy to distinguish from a beach deposit outside a delta. In a wave-dominated delta, however, beach ridges form right in front of the delta top deposits, which often contain coal beds from vegetation, whereas beach profiles formed by barrier islands have a lagoon behind them. The first unambiguous proof that one is dealing with a delta, however, is if the beach ridges are found to be cut by fluvial channels. In addition to the Rhone and the Niger deltas, the Nile delta is also a good example of a wave-dominated delta Fig. 2.38. Here sediment supply to the delta has been greatly reduced through the building of the Aswan Dam, which traps sediments, and the balance between sediment supply and wave erosion has been disturbed so that the delta is now retreating. This is a modern example of how progradation or destruction (transgression) of a delta depends on a very sensitive balance between sediment supply and erosion.

2.35 Coastal Sedimentation Outside Deltas

Most sediment transport to the sea takes place via rivers which drain into deltas. In fluvially-dominated deltas, the greater part of the sediment is deposited there.

On destructive deltas, a great deal of sediment is eroded and transported out into the sea or along the coast. Coastal areas between deltas are very largely supplied with sediment which has been eroded from deltas and transported along the coast by waves and coastal currents. Since the surface of the coast slopes outwards, the direction parallel with the shoreline can be defined as the *strike* of the deposit, and the direction normal to the coast is called the *dip*.

Sediment supply parallel with the coast is often called “strike feeding”. If the supply of sediment along the coast is greater than the erosion rate, the coast will build out into the sea and we will get regression through coastal progradation. If sediment supply is less than the rate of erosion, the coast will retreat, and we will get transgression.

Because there is a finite amount of sediment being transported along any stretch of the coast, engineering constructions which are sometimes built to accumulate sediment and prevent erosion at a particular spot (groynes) will as a rule lead to increased erosion along other parts of the coast. This has great practical significance in areas with high coastal erosion. Beach sedimentation must therefore be seen, not in isolation, but in the context of the overall sediment budget along the coast.

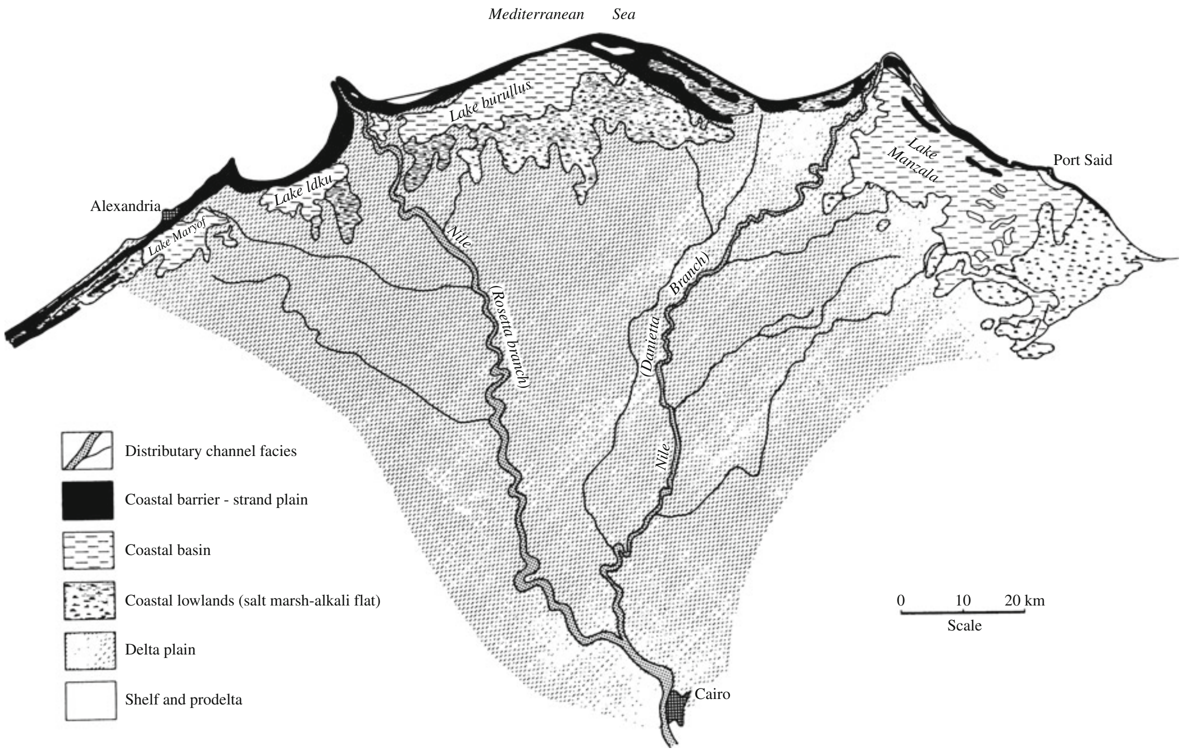


Fig. 2.38 The Nile delta system is an example of a wave-dominated, destructive delta system. (Fisher et al. 1974)

The Mississippi delta is the greatest source of sediment to the Gulf of Mexico, although there are a number of minor rivers in Texas. Here sediment transport takes place along the coast from east to west. Because the active part of the Mississippi delta lies as far east as it does today, less sediment arrives at the Texas coast. The fact that this coastline is subsiding at the same time leads to transgression of the coastline. This example shows that if the supply of sediment due to strike feeding cannot match subsidence, the result will be a transgression.

2.36 The Shore Zone

The shore zone is where land and sea meet, and we can distinguish between different types:

1. *Rocky Beach*. The beach is covered with pebbles and blocks or solid rock representing ancient, resistant bedrock. Much of the coast of Norway is of this type, but only 2% of the coastline of North America. Rocky beaches form in areas which have been uplifted tectonically, and where the coast erodes metamorphic or eruptive rocks.
2. *Sand or Gravel Beach*. This is the most common form of beach zone where we have active sedimentation or erosion of older sediments (makes up 33% of the coastline of North America).
3. *Barrier Island*. This is a beach which is separated from the main coastline by a lagoon. Common in North America (22%), but less common in other areas.
4. *Muddy Coastlines*. Beach zones which consist basically of clay are formed where we have sediments with a very low sand content, and where there is little wave activity to wash out or enrich any sand the sediments might contain. We find this in parts of clay-rich deltas and estuaries and where there is abundant vegetation along the beach zone which protects the clay sediments against erosion (e.g. mangrove forest).
5. *Cheniers* are isolated sandy beach ridges on coastal mud flats. They are abundant along the coastline around major deltas like the Mississippi and the Amazon. They require low tidal ranges, moderate

wave energy and abundant mud, and a limited amount of sand.

The composition of beach sediments varies greatly, depending on the materials available and the grade of mechanical and chemical breakdown of the minerals. Quartz sand is the most widespread because quartz is the most stable of the minerals we find in most beach sediments. But on volcanic islands, which consist largely of basalt, there is no quartz, and we get sand derived from basalt or volcanic glass. Carbonate sand is formed locally from the broken skeletons of carbonate-secreting organisms. This does not only happen in tropical regions, e.g. the Bahamas, but also along coasts with cold climates. In many parts of Norway, for example, the beach sand consists very largely of carbonate sand from molluscs, barnacles, bryozoans and calcareous red algae.

Beach profiles are formed by wave power acting on the coast and depend on a number of different factors:

1. The composition of the available sediments.
2. Wave power
   a. Average wave height
   b. Size and frequency of storms
   c. Angle between the commonest orientation of the waves and the beach.
3. Tidal range.
4. Vertical profile off the beach.
5. Supply of sediment from land or along the coast.

The morphology of the beach zone and nearshore areas is the result of interaction of various features related to wave activity. As waves approach land, they will be affected by friction against the bottom. The depth at which this starts depends on wave height and length. When the depth becomes less than about half the wave length, friction against the bottom will be great, the circular wave motion (orbit) will be distorted and oscillatory sediment transport will affect the seabed, producing ripple marks. The depth at which this occurs is called the *wave base*. Dunes are formed of sediments which are deposited when the waves break, and are at the same time the reason for the waves breaking precisely there. There is thus an interaction between the wave regime and the bottom geometry in beach sediments. In addition to breaking on the *foreshore* itself, we often find that waves break at two or three places offshore, and at each of these places we find a sand bar.

In the French Riviera (e.g. Nice) the beaches are often full of rounded pebbles with less sand. This is because they are near mountains supplying coarse clasts and also close to a shelf edge. Sand is then easily eroded from the beaches during storms while the coarse clasts remain.

2.37 Prograding Beach and Barrier Sequences

The progradation of a beach will produce a characteristic coarsening-up sequence (Fig. 2.39) which will obey Walther's Law of facies succession. The vertical sequence will represent the environments from the shelf to the shoreline. The transition between shelf mud and fine-grained sandstones with ripples may represent the wave base. Isolated sand layers in mud may have been deposited near the *storm wave base*, whereas the transition to continuous fine-grained sand may represent the *fairweather wave base*. The thickness of the sequence from the fairweather wave base to the foreshore is an expression of the wave energy at the coastline when the sequence was deposited.

Bioturbation occurs in the lower part of this sequence. Below the wave base it usually takes the form of horizontal feeding traces, and in the lower and middle shorefaces, where there is relatively high wave energy, as vertical traces (*Skolithos* facies). As erosion and reworking intensify, the preservation potential of bioturbation is reduced and it becomes less frequent. The formation of sand bars and erosion surfaces on the upper shoreface results in cross-bedding, usually trough cross-bedding, representing flow in the upper part of the lower flow regime.

In the breaker zone there is an upper flow regime, producing a planar facies which in vertical section will appear as very low-angle cross-bedding. On the beach we often have a beach bar which is flooded only during storms, and a depression behind it called a runnel, which helps to drain the backshore area. Because the exposed beach is a rich source of sand, the wind will tend to blow sand from the beach and redeposit it as aeolian dunes, usually where it is trapped by vegetation. Aeolian sediments therefore often cap ancient beach profiles, but the aeolian dunes may also be eroded and not be preserved in the geological record.

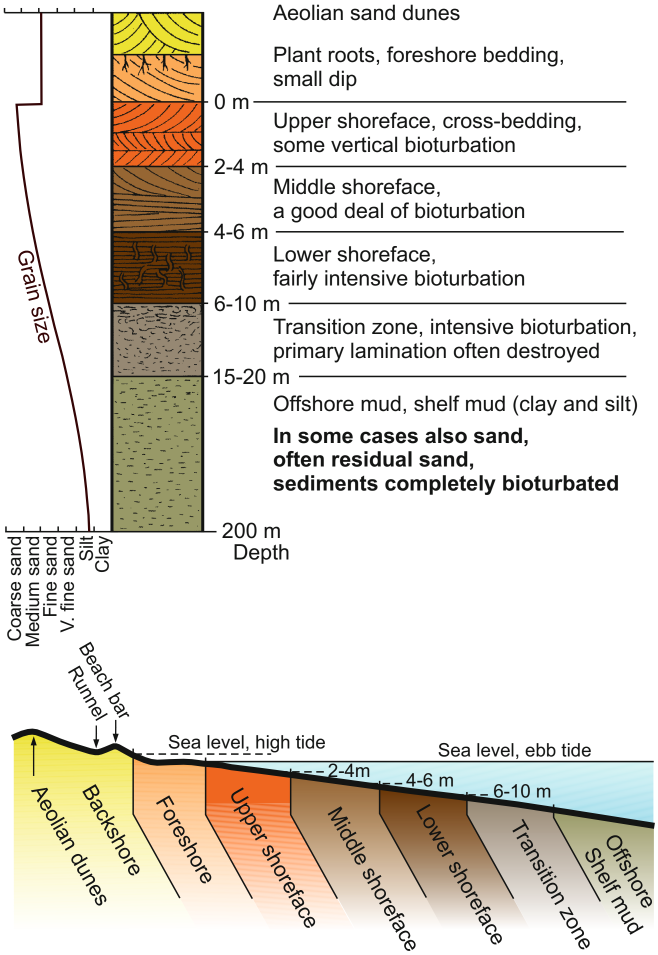


Fig. 2.39 Diagrammatic section through a sand deposit with some typical sedimentary structures. Just under normal fair weather wave base, 5–20 m, "hummocky" stratification occurs

2.38 Barrier Islands

Barrier islands are beach deposits which are separated from the mainland by a lagoon. They form long, thin sand ridges which are often only a few hundred metres to a couple of kilometres broad, and which rise up to 5–10 m above sea level.

A vertical section through the part of a barrier island facing the sea resembles an ordinary beach deposit (Fig. 2.40). We find a coarsening-upwards sequence from marine clay to beach sand, often with vegetation on top. On the lee side, facing the lagoon, there is little wave power and clay, mud and often oyster reefs are deposited in the lagoon. Barrier islands are very well developed along long stretches of the coast of North America, particularly off Texas, Georgia and North Carolina. There are several indications that barrier islands are due to transgressive conditions such as those of Holocene times. This is certainly the case along the eastern coast of the southern North Sea. When the ocean rises in relation to the land, beach deposits can continue to grow through gradual deposition of sand in the beach zone, but the areas behind them sink below sea level and form a protected lagoon with clay sedimentation. More localised transgressions may be caused by compaction and subsidence of sediments along a coastline where the sediment supply is insufficient to keep pace with the subsidence.

One prerequisite for forming barrier islands is an adequate supply of sand, so that the island can grow and keep pace with the transgression. This sand cannot

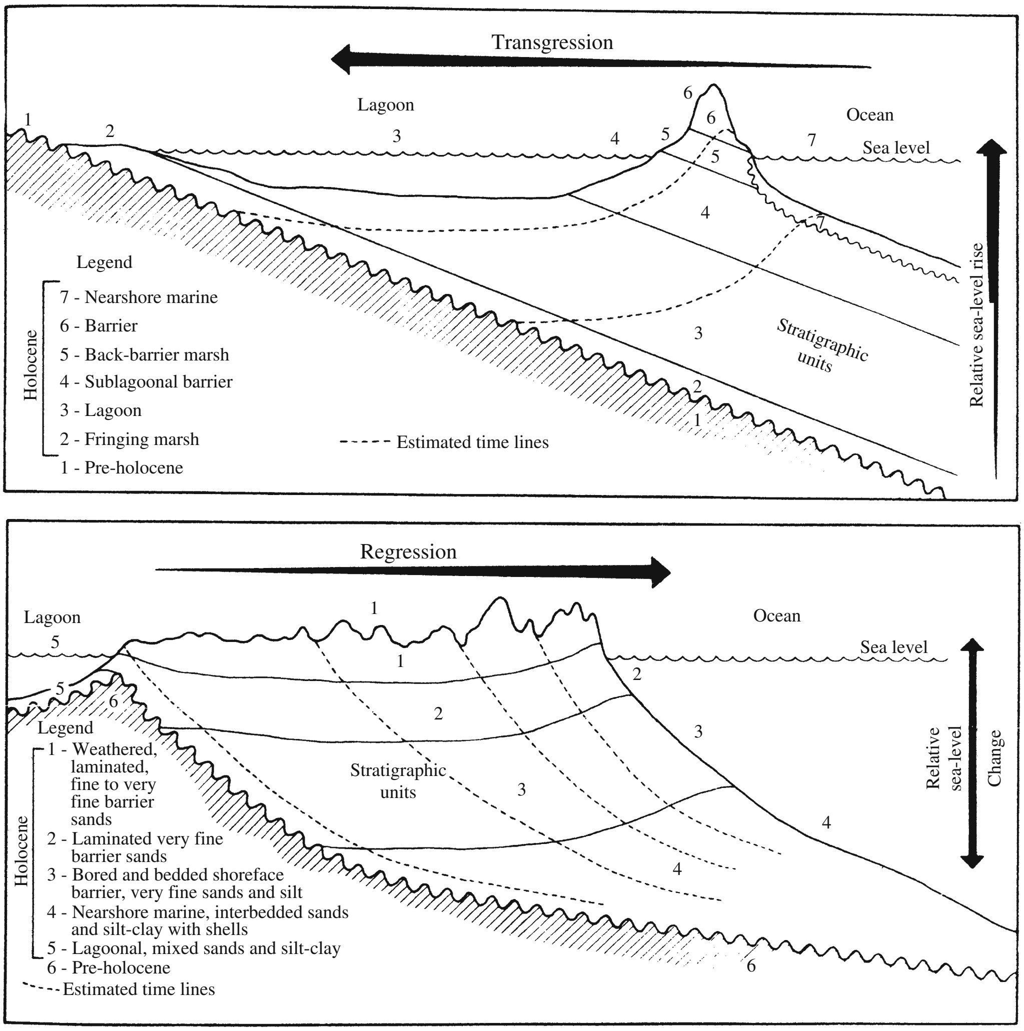


Fig. 2.40 Transgressive and regressive barrier systems. (from Kraf and John 1979)

be transported across the lagoon, and must be added along the length of the islands parallel to the coast (strike-feeding from deltas or eroding coastlines).

During storms or hurricanes the sea level may rise due to wind stress, and waves may break over and through the barrier island. An erosion channel is then formed through the island, and a *washover fan delta* develops at the rear, out into the lagoon. A delta of this type can form in a matter of hours during a hurricane. Barrier islands may extend for tens of kilometres, but they will not form a continuous belt along the coast. Water has to circulate between the lagoons and the ocean through gaps between barrier islands. The gaps are called *tidal inlets*, and the distance between them will be a function of the tidal range. Both seaward and landward of the inlets small sandy deltas may develop in response to ebb and flood currents respectively (Fig. 2.41). Flow at ebb tide will normally be stronger than that at flood tide. This is due to the profile of the lagoons. At high water in the lagoon the volume of water which must flow out to compensate for a specific lowering of sea level is greater than the volume needed to raise the water level in the lagoon correspondingly at low tide when the area of the lagoon is smaller. The tidal inlets are therefore capable of transporting more sediment out during the ebb, and structures indicating this flow direction may predominate (Fig. 2.41). Inlets are characterised by an erosional base, and lateral migration of inlets produces a characteristic fining-upwards sequence. The strong currents in tidal inlets

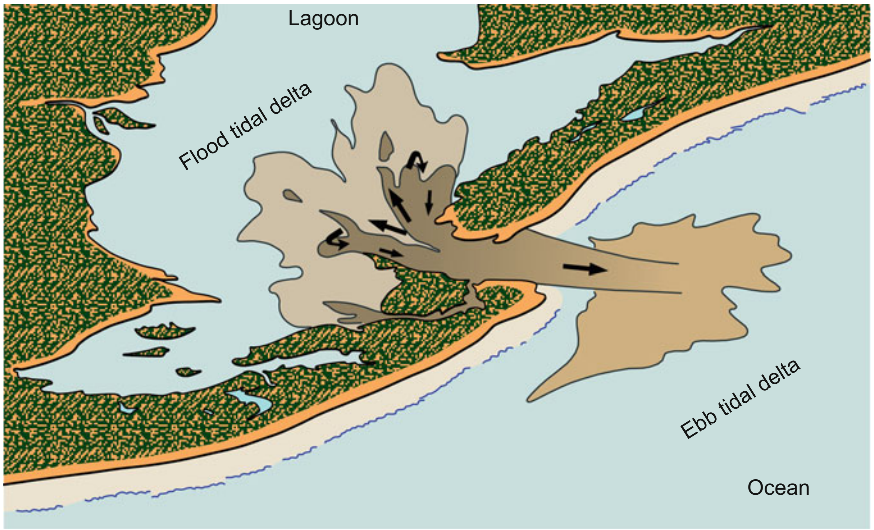


Fig. 2.41 Tidal channels with tidal deltas forming between barrier islands. The barrier islands and the channels will migrate laterally and deposit channel facies sediments by lateral accretion. Note that the ebb-tidal delta outside the barrier is much more exposed to waves than the flood-tidal delta in the lagoon

often generate sand waves which tend to migrate in the ebb direction, but they may also be modified by flood currents.

Ebb-tidal deltas consist of a channel dominated by ebb currents with smaller flood-tide channels on the sides. At the ocean end of the channel sediment is deposited in a sand ridge which is similar to a channel mouth bar in an ordinary delta. This sand ridge, which is called *a terminal lobe, is* subject to wave erosion, and smaller *swash bars* may form, which reach above sea level. In areas with strong wave power, ebb-tidal deltas will be less obvious because of erosion and further transport along the barrier ridges. Ebb-tidal deltas will be characterised by greater water depths than flood tidal deltas.

Flood tidal deltas form inside the lagoon and are well protected from wave erosion. Here the water flows into flow channels which branch inwards in a flood-tidal delta, where the sediments are deposited on a tidal flat. Ebb currents move back along the edges of the outer side of this delta and may form small *spill-over lobes* when ebb-currents penetrate over the edge of the flood-tidal delta. Flood-tidal deltas are associated with shallower channels than ebb-tidal deltas and are not eroded very much by waves.

Tidal channels fill with sand which forms an upward-fining sequence overlain by tidal flat sediments (Fig. 2.42). In areas with carbonate sediments or cohesive clays, erosion due to lateral migration of tidal channels results in intraformational breccias.

Barrier island deposits thus consist of a long, thin body of sand. The thickness of the sand layer will correspond to the depth of the wave base plus a few metres which correspond to the height it builds up to above sea level.

In areas with a larger tidal range, this lateral migration will be rather pronounced, and fining-upwards sequences will also be common.

If the barrier islands are drowned by a transgression, a carpet of clay and silt will be deposited over these sandstone deposits. This represents the ideal stratigraphic trap for oil and gas. Compaction or tectonic tilting will cause the sandstone deposits to interfinger with mud from the lagoon deposits, which are a good source rock. Oil will be able to collect in the top of the barrier ridge sand or in flood-tidal delta deposits (or *washover fans)* which represent *pinch-outs* in the muddy lagoon sediments.

2.39 Tidal Sedimentation

Tidal range is an important factor in coastal sedimentation. We distinguish between:

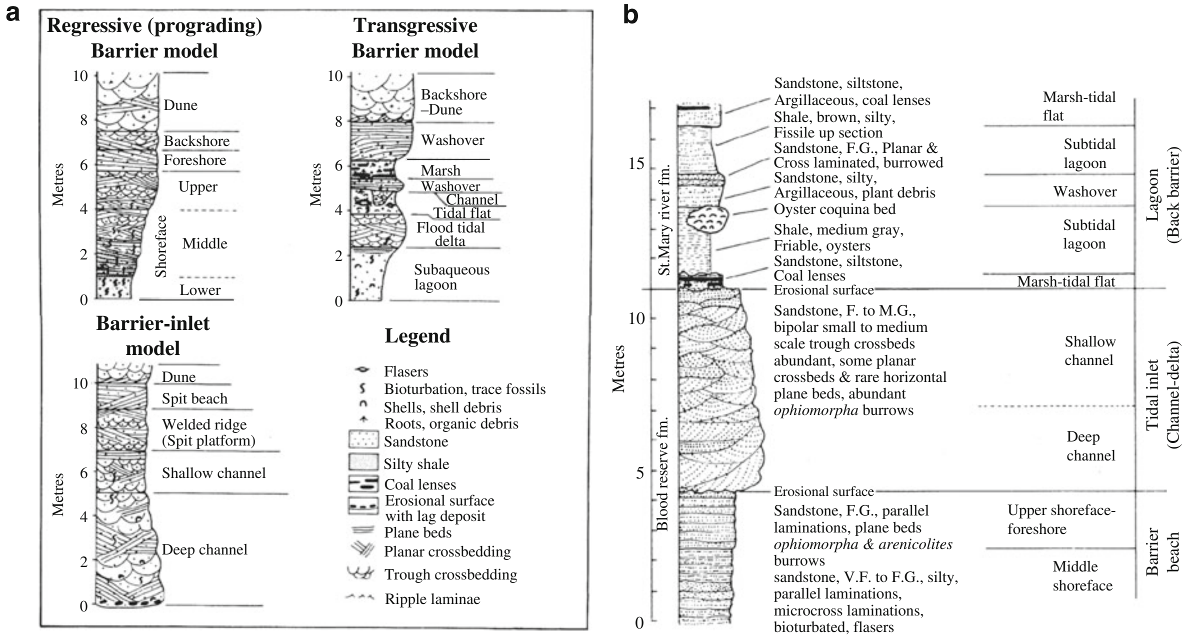


Fig. 2.42 (**a**) Interpretation of a vertical section in a tidal environment (From Walker 1979). (**b**) Section through a barrier beach cut by a tidal inlet channel into the lagoonal facies

1. Microtidal environment (tidal range <2 m).
2. Mesotidal environment (tidal range 2–4 m).
3. Macrotidal environment (tidal range >4 m).

The average tidal range in the open sea is only about 50 cm. Along the coasts, however, we often get increased interference by tidal currents. This is particularly true around large islands where tidal waves converge on the lee side and can build up, and also in bays along the coasts. In long, narrow bays we may get a high degree of *resonance*. This occurs if the bay has a length and depth which cause tidal waves which are on the rebound to reinforce the next incoming tidal wave. The highest tidal range which has been measured is 16.3 m in the Bay of Fundy in Canada. In some areas around the British Isles the tidal range may be up to about 12 m, and there are also large ranges in the German Bight and adjacent parts of the North Sea. Tidal waves also move anticlockwise in N hemisphere around centres with zero tidal range (amphidromic points), with the tidal range increasing radially with the distance from the centre. This is typical of the tidal pattern in the North Sea.

The width of the continental shelf plays a major role in determining tidal ranges. When tidal waves enter shallow water, their velocity is reduced due to friction against the bottom. When the water depth and the velocity decline, the height of the tidal wave will increase, so that the total energy flux is maintained. The tidal range, which is thus the height of the tidal wave, therefore increases inwards across the shelf. Where there are embayments along the coast, tidal waves become focused, so that the tidal range increases towards the middle of the bay. This is true of the east coast of the USA, off Georgia, and of the German Bight in Europe.

If the shelf becomes even wider, as in an epicontinental sea, tidal power will gradually be exhausted due to the high frictional resistance. This is the case on the present Siberian continental shelf where the tidal range is very low (<10–20 cm). Wind stress may set up rather large storm tides. When the water is very shallow, as on the Bahamas Bank, friction damping takes place over much shorter distances.

The great epicontinental seas, in Ordovician and Cretaceous times for example, were also characterised by low tidal ranges.

About 1/3 of the world's coasts have tidal ranges greater than 4 m (macrotides), 1/3 have mesotides (2–4 m) and 1/3 microtides (less than 2 m). Small inland seas, like the Mediterranean, the Black Sea and the Baltic Sea, are too small to keep pace with the attraction of the moon and the sun, and have small tidal ranges. This is also true of lakes.

1. *Tidal Channels*. The infills may resemble fluvial channels or submarine channels in that they form fining-upwards sequences. Channels formed in estuaries in fact are often connected to fluvial channel systems. Tidal channels which are not part of a river delta, however, tend to be filled with sandy sediments from the surrounding tidal flat, because they have no supply from land. Channels will often erode their banks and cause them to collapse, and this may result in the formation of intraformational conglomerates if the sediments are slightly lithified.

Lateral migration of tidal channels may produce typical epsilon cross-bedding which is the result of lateral accretion of point bars in the tidal channel.

Tidal channels often contain a bed of marine fossils at the base, and marine trace fossils. Channels on tidal flats which are not associated with deltas (i.e. those in estuaries) will not receive much clastic material from land. Conglomerates and breccias in these tidal channels will therefore typically be of the intraformational type, derived by local reworking of tidal flat sediment. Because of their early lithification, carbonate beds in particular can be reworked to form intra-formational conglomerates and breccias.

Tidal cycle duration is about 12 h and 25 min, with currents switching direction every 6 h, and we sometimes find good examples of cross-bedding with opposite current directions. This is not always the case in tidal environments, however. Some tidal channels are dominated by ebb flow and others by flood currents. This is because the ebb and flood often find different dominant pathways. Bipolar cross-bedding is therefore not an essential feature of tidal channels. Storms with opposing wind directions may also produce some form of bipolar cross-bedding.

In a regressive sequence tidal channels will be overlain by lagoonal sediments (Fig. 2.42).

2. *Flaser Bedding*. Consists of clay laminae in a matrix of sandstone with ripple cross-lamination. The clay occurs mainly as infill in ripple troughs, but may also drape over the ripple crests as well. Flaser bedding forms as a result of alternating periods of currents or wave activity, and slack water. The clay settles out in the slack water periods in a tidal environment, though this type of bedding may also form in other environments where there is rhythmic sedimentation, such as in certain fluvial environments. On tidal flats, clay and silt will settle out at high tide to be deposited between ripples formed by the ebb and flood currents. Here the fine sediment may consist partly of clay pellets (faecal pellets from marine organisms) which settle out faster than clay-sized particles. Flaser bedding belongs to a type of structure which we get with mixtures of sand and clay. *Lenticular bedding* represents isolated laminae or lenses of sand in mud.

The inner parts of a tidal shelf often have embayments consisting of very muddy sediments, usually with abundant bioturbation. Mollusc shells in the mud (Fig. 2.43) are often eroded and deposited as shell lag.

3. *Tidal Bundles*. The best identifying feature for tidal environments is regular lamination consisting of fine sand and mud, making up tidal couplets that each represent a tidal cycle. Both modern and ancient tidal sediments show a regular variation in the thickness of such couplets, reflecting the energy levels of spring and neap cycles. Regular laminations reflecting tidal cycles are often called tidal bundles (Fig. 2.44).

2.40 Shallow Marine Shelves

The shelf extends from the nearshore environment to the shelf edge, where there is a rather abrupt increase in slope, usually at a depth of 200–500 m. The width of the shelf varies considerably, and may exceed 1,000 km. Continental shelves are generally very flat areas which may be cut by deeper channels transporting sediments across the shelf from nearshore or deltaic environments.

The study of sedimentation on modern continental shelves is complicated by the fact that sea level was more than 100 m lower only 10,000 years ago. This means that most shelf areas have not yet reached an equilibrium with respect to the modern environment. Sandy shelf deposits are much more difficult to core than muddy sediments and the commonly used gravity corer has to be replaced by vibro-core equipment.

Most shelf areas are below the wave base for normal waves (fairweather wave base) and sedimentation is governed largely by tidal currents and storms.

When the wind is landward, waves will usually approach the beach obliquely, resulting in wave refraction effects, particularly on relatively steep beaches with high wave energy. This will produce *rip currents*, where the water piling up against the

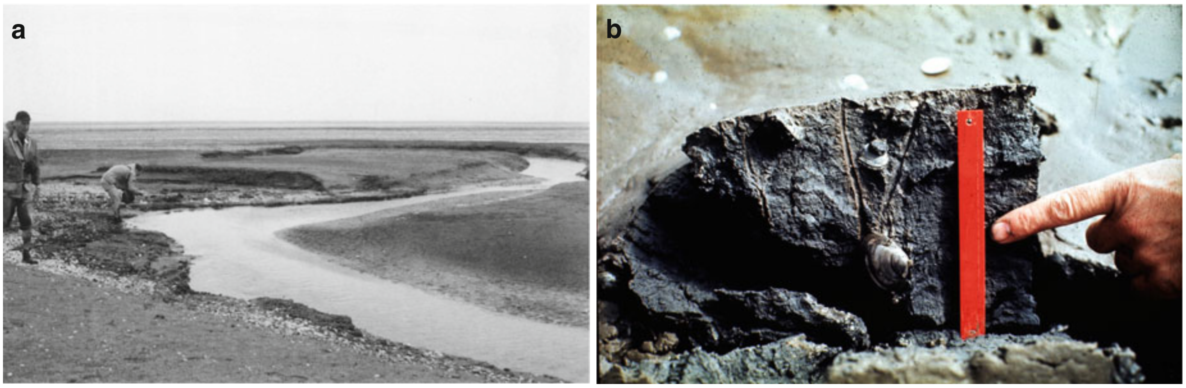


Fig. 2.43 (**a**) Tidal channel on a tidal flat in the muddy facies of the inner part of a tidal flat (near Wilhelmshaven, Germany). Erosion by tidal channels in this mud produces a lag of mollusc shells. (**b**) Mollusc in living position

Fig. 2.44 Tidal bundles. These are bundles of laminae which reflect the tidal cycles between spring tides. From the Late Precambrian Wonoka Formation, Patsy Spring, Flinders Ranges, Australia

beach has its return flow seawards. Rip currents are often rich in suspended material and transport material from the beach towards the shelf. Another component of wave energy is transmitted parallel with the beach as longshore currents, contributing to the longshore drift of sediment transport.

During onshore storms the sea level near the coast may be raised by several metres due to the combined effect of wind stress, tides and the low barometric pressure associated with storms. The increased potential (elevation) of the coastal water due to these *storm surges* will result in strong bottom currents which are capable of transporting sediment further out onto the shelf. Storm surges may transport fine sand and mud in suspension, but not as true turbidity currents. In this case the increased potential of the coastal water is the driving force and not the density difference between the current and the surrounding water, as with turbidites. Hummocky cross-bedding is a characteristic sedimentary structure produced by the deposition of coarse particles from suspension during storms, and rounded, undulating sand surfaces tend to form. Finer particles will be transported further, into areas with lower energy.

In many modern shelf areas tidal currents are important transport mechanisms, in combination with rare strong storm currents. Sediments deposited in this environment are characterised by rather abrupt transitions from well-sorted sand to mud and from bioturbated to non-bioturbated strata.

The main characteristic of tidal shelves is the mobility of the clastic sediments, on scales ranging from the diurnal tidal cycle to annual (storm augmented) cycles and gradual long-term movement of the largest bedforms. In shelf areas with relatively strong tidal currents (>150 cm/s) we may get *furrows* and *gravel waves*. At velocities of less than 1 m/s *sand ribbons* may be deposited – longitudinal bedforms developed parallel to the currents. Sandwaves are large-scale transverse bedforms, generally 2–15 m high, with a wavelength of 150–500 m. Sandwaves require current velocities exceeding 60–70 cm/s. Cross-stratification may be symmetrical or asymmetrical on both sides of the sandwave, depending on the relative strength of the opposing currents.

Relatively low-energy environments (<50 cm/s) are characterised by *sand patches* and mud. The sand forming the patches probably only moves during storms.

Shelf sediments characteristically accumulate at relatively low overall sedimentation rates (1–10 mm/1,000 years).

2.41 Continental Slopes

Continental slopes are the areas between the edge of the mostly very flat continental shelf which commonly lies at a depth of 200–500 m and the *continental rise*, where the ocean deep begins at a depth of 2–4,000 m. The continental slope gradient is typically 2–6°, and it is 20–100 km broad. The gradient is a function of a number of different factors, but the stability of the shelf edge constitutes a major control factor. The steepest slopes are therefore to be found off carbonate banks with well-cemented coral reefs and carbonate beds which have high shear strength. In areas with rapid sedimentation, loose sediments have little shear strength and submarine slides, slumping and formation of turbidites occur, maintaining relatively gentle slopes (1–2°). Where sedimentation is slower the sediments have more time to consolidate, and will be more stable. The steepest submarine slopes in clastic sediments (greater than 10°) are therefore found in submarine canyons, where erosion cuts into older, well-consolidated sedimentary strata.

Along passive continental margins the continental slope is associated with the transition from continental crust to oceanic crust. In areas with a large supply of sediment, the shelf may have prograded beyond this boundary.

2.42 Organic Sedimentation on the Slope of the Continental Shelf

The continetal slopes are enriched in organic matter compared to the shelf and the deep ocean.

This is because the slope is where we have the greatest upwelling of nutrients from the deep. We also find low oxygen content in the water column on continental slopes, allowing much of the organic matter produced to be retained in the sediments. There will also be high productivity on shallower slopes in front of deltas because of the large nutrient supply from river water, but the organic matter may be greatly diluted by rapid clastic sedimentation. On the continental shelf the supply of nutrients is small and the prevalence of stronger currents and turbulence means that most of the organic production there will be oxidised.

Out in the open ocean basin organic production is relatively low, due to a limited supply of nutrients. Much of the planktonic organic matter is oxidised near the seafloor by deep currents of cold, oxygenated water from the polar areas.

Sediments deposited on the continental slope are therefore more promising as source rocks for oil than shelf and deep-water facies.

2.43 Sediment Transport on Submarine Slopes

Gravitational processes are naturally important on submarine slopes.

Gravity forces can be represented by forces (vectors) normal to, and parallel to, the slope. The component parallel to the slope consists of shear forces which may overcome the shear strength of the sediment, causing slumping. Sliding of large volumes of sediments downslope produces extensional faulting in the upper part of the slope and compression in the lower part. Gravitational instability on the slopes may also develop into debris flows and turbidity currents. Collapse of the sediment grain framework may cause sudden compaction and liquefaction of the slope sediments.

Traction currents may, however, also play a part on the submarine slopes, particularly in canyons but also near the toe of the slope, where we may have *contourites* – deposited by currents flowing parallel to the slope contours.

2.44 Submarine Canyons

These are valley-shaped depressions which extend from the top to the bottom of the slopes, down to 2,000–4,000 m. In some cases they may start in shallow water near the beach, in others close to the edge of the shelf. The height from the bottom of the canyon to the top of the slope on each side may be up to 2,000 m. We are dealing with enormous topographical features, which would have been very impressive indeed if they had been on land, towering structures on the scale of the Grand Canyon.

Shepard et al. (1979) systematically gathered data on currents and sediment transport in submarine canyons and found that tidal currents are of great importance *also* at great depths in submarine canyons. Current meters have shown that currents flow *both up and down* the submarine canyons, and that they switch every 6 h like tidal currents in shallow water. Current velocity is often only 10–20 cm/s, but in many canyons velocities of up to 40 cm/s occur sometimes, powerful enough to transport fine to medium-grained sand. The flow velocity tends to be greatest in the upper part of the canyon and diminish downvalley. Most sediment transport takes place during these episodic and unusually high flow velocities which may be linked with storms which create wind-induced shear forces which sweep the water up against the coast *(storm tides)* and may cause currents to develop along the bottom and down the submarine canyons. However, high flow velocities have also been measured without it being possible to associate them with storms or wind stress. The most powerful flow velocity is most commonly directed downvalley, but upvalley-directed streams have been observed with velocities of up to 90 cm/s.

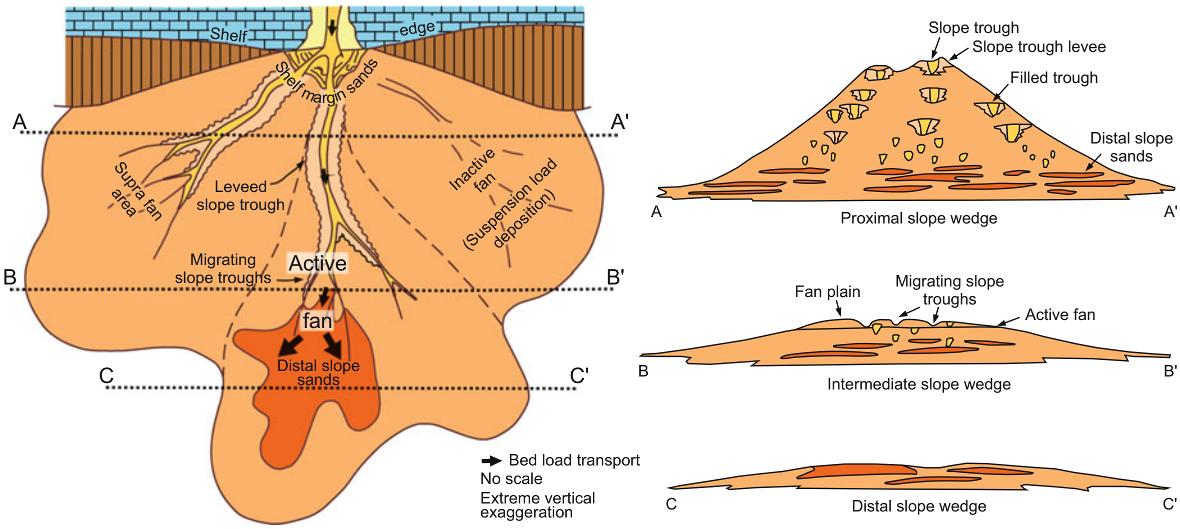
Detailed measurements in the submarine canyons off the coast of California reveal that the highest flow rates are oriented up the canyon in a way which seems to indicate that they are generated by internal waves from the ocean basin, and not by gravitational forces. Currents may then develop when the waves "break" against the coast or the continental shelf.

Powerful currents in submarine canyons are capable of transporting sand, sometimes in rare instances even coarser material. These are frequently not turbidity currents, but *traction currents*, which transport and deposit better-sorted material. Turbidity currents have been observed in submarine canyons as well, but definite (observed) examples were only low-velocity, low-density turbidity currents with a maximum velocity of 70–100 cm/s. In submarine canyons we thus have both traction currents, which are controlled partly by tidal forces, and turbidity currents. Downward-moving currents driven by tidal forces may, if they contain much suspended material, turn into turbidity currents. The downward-moving currents have both a component of traction and gravitation.

The relief of submarine canyons is due partly to erosion down into the underlying sediments, and partly to lack of deposition in the canyon while the adjacent beds were being deposited. During low sea level stands rivers may prograde closer to the shelf edge and hence supply more sediment to the submarine canyons, thus feeding submarine fans. At sea level highstands, currents in the submarine canyons and the shelf may erode shelf and slope sediments and deposit pure sand onlapping an erosional unconformity at the toe of the canyon.

Most of the canyon itself is an area of sediment transport and erosion. Deposition takes place where there is a change of slope near the basin floor. Here the channel defined by the canyons splits up into several channels which build depositional lobes called *suprafan lobes* (Fig. 2.45).

As the lobes build up, the gradient of the slope is reduced and a new channel will form in a part of the fan where there is a steeper slope (Fig. 2.46). This produces lobe-shifting similar to that observed in fluvially dominated deltas. Each lobe will tend to build a fining-upwards sequence, with conglomerate and coarse sand near the base. On the sides of the channels fine-grained material in suspension is deposited as thin-bedded turbidites. The levee builds up on both sides of the channel, and resembles a river levee. On the submarine Amazon delta slope there are well-developed meandering channels. The distal fan is also dominated by fine-grained sediments deposited as thin, graded fine sand, silt and clay. Progradation of submarine fans may also produce an upwards-coarsening sequence (Fig. 2.47).



Slope fan model

Fig. 2.45 Model of submarine fans deposited at the foot of submarine slopes. Submarine fans have channels which sometimes meander. On the sides of the channels we find fine-grained levée deposits resembling those along fluvial channels. The pattern of shifting depositional lobes resembles that of deltas. (Brown and Fisher 1977)

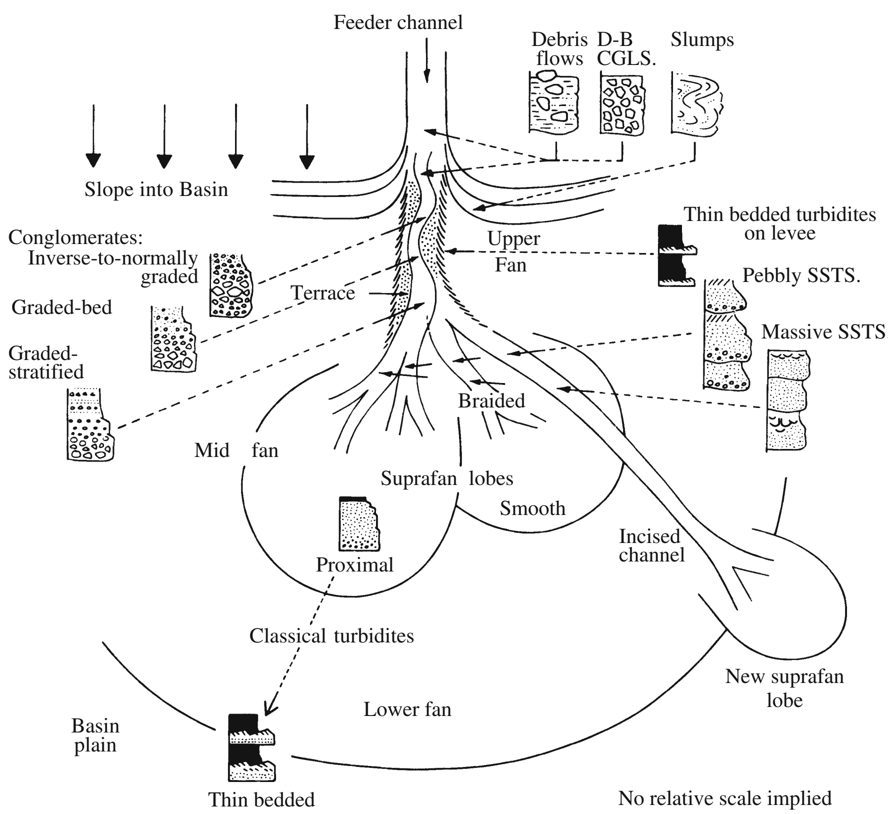


Fig. 2.46 Submarine fan model showing progradation and shifting of lobes similar to delta lobe shifting. One of the main differences between submarine fan and delta facies is the absence of wave reworking. (Walker 1984)

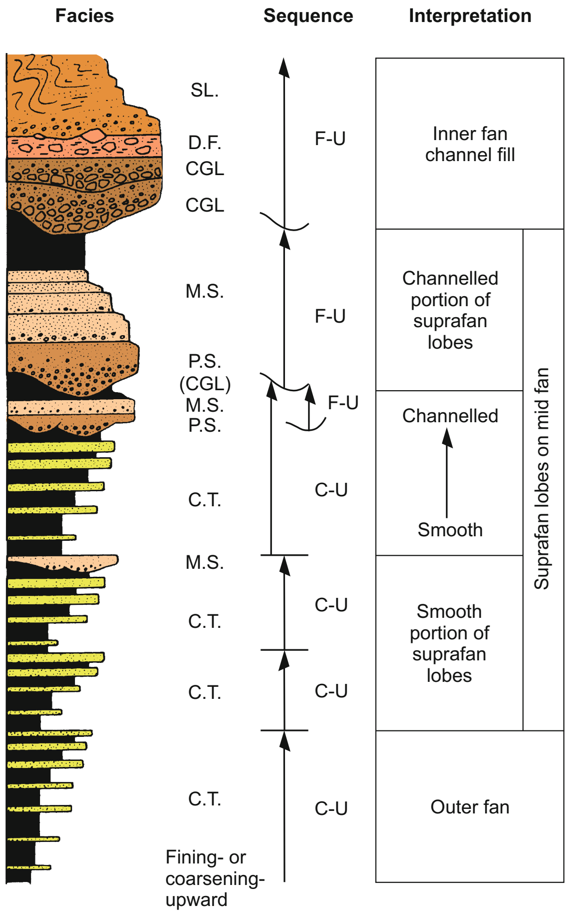


Fig. 2.47 Vertical sequence through a submarine fan (Walker 1984)

2.45 Sedimentation Along Continental Margins and in Epeiric Seas

As discussed above most of the sediment derived from land is deposited in river deltas and distributed relatively close to the shore line by longshore transport. There is, however, also significant sedimentation further away from the deltas and the coastlines. Turbidites and debris flows can in some cases transport sediments hundreds of kilometres offshore. Debris flows represent a very efficient type of sediment transport which can carry sediment far beyond the continental slopes because there is little internal shear deformation, only at the base and the top of the flow.

In fine-grained shales we may find evidence of ripples indicating traction currents. Storms and spring tides may produce rather high velocities (20–40 cm/s) on the seafloor far away from land and these currents are capable of transporting silt and fine sand. Clay should not produce ripples, but the clay is often composed of pellets of clay aggregates, often also with some organic matter. These clay particles behave like silt but they have lower densities and may be transported by traction currents forming small-scale ripple cross-lamination.

There may also be significant contributions of aeolian dust from deserts and of volcanic ash. In Palaeozoic times, before there were many land plants, aeolian dust was more important, but even today much of the sedimentation in the South Atlantic and Pacific Ocean is aeolian.

2.46 Sedimentation Along Island Arcs and Submarine Trenches

Submarine trenches form along converging plate boundaries where oceanic lithospheres are disappearing into a subduction zone. Along these converging plate boundaries sediment basins with very special deposition environments are formed. There are three main sources of sediment:

1. From the continent.
2. From island arcs, which may consist of continental crust, oceanic crust and/or volcanic rocks.
3. Pelagic sediment, including biogenic sediment and wind-blown volcanic ash.

These have quite different compositions. Sediment which is added from the continent is deposited in deltas and in turn fills the basin behind the arc *(back-arc basin)*. Sediments which are formed on the island arcs are rich in volcanic material, and this will characterise deposits in small basins on island arcs and *fore-arc basins*. Back-arc basins may also receive volcanic sediments from the island arcs. During the initial subduction phase fore-arc basins will tend to be characterised by turbidites deposited in relatively deep water, but sediment filling may convert them into a shallow water environment which may include carbonate sediments. A structural high separates the fore-arc basin from the actual slope down to the submarine trench.

Beyond the deep sea trenches a significant amount of pelagic sedimentation takes place on a relatively flat ocean crust. Some of the sediments get scraped off the subducting volcanic crust and stacked up in what are called "accretionary prisms" (Fig. 2.48). Some

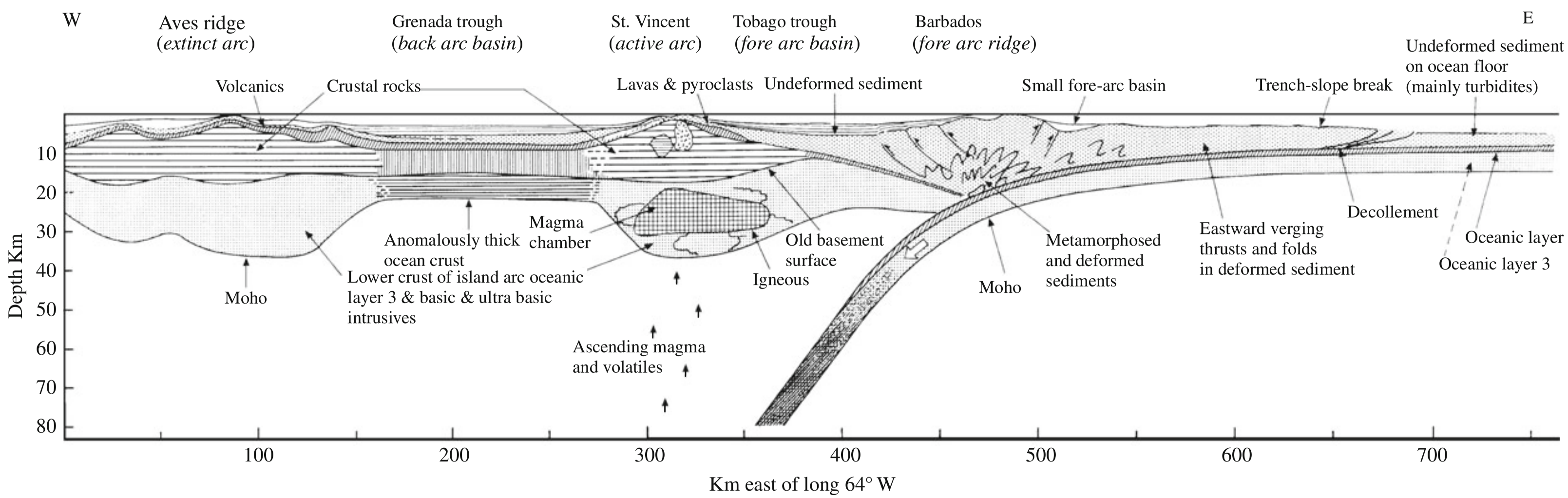


Fig. 2.48 Section through an area with converging plate boundaries in the Caribbean Sea. Note that sediments deposited on oceanic crust are scraped off the descending plate and imbricated in the accretionary prisms (see Fig. 2.49). From Leggett (1982)

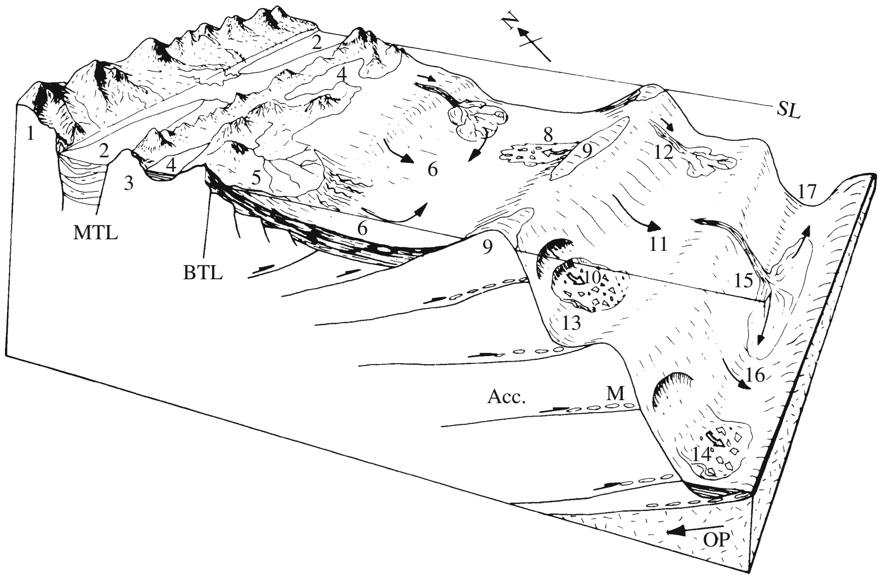


Fig. 2.49 Sedimentation in a submarine trench (accretionary prism) near Japan. Sedimentation is very much influenced by the relative movements (thrusting) of the rock units piled up in the prism (from Taira et al. 1982)

sediment may also be carried down with the subducting plates. The supply of sediment to the deep-sea trenches themselves is often very limited, which is why they do not fill up with sediment. They represent the greatest depths in the ocean (up to 10 km) and this can be explained isostatically by the fact that the oceanic plate which is undergoing *subduction* is cold, and therefore heavy. The downward movement acts against the direction of heat flow, resulting in low geothermal gradients and therefore dense crust.

The sediments may be pelagic oozes or distal turbidites. Since the oceanic plate is moving towards the island arc, the sediments on the oceanic crust have been deposited further away from the sediment source, and in consequence we do not normally have very thick sedimentary sequences in the subducting plate. The accretionary prism consists of a series of sliding faults which are steepest near the surface and have a lower gradient downwards. They are often draped with a blanket of pelagic sediments (Fig. 2.49).

Listric faults of this type are similar to those we find in plate boundaries with tension (rifting) but the relative movements are in the opposite direction (reverse faults). Sediments which are still not very consolidated tend to deform along the imbricated faults and develop various kinds of *drag folds*. Continued movements of the imbricated fault planes cause this slope to become very steep locally, and conglomerates and fan deposits may become unstable and slide. Lithified carbonates and sandstones will break up and form large blocks in a more clay-rich matrix. Volcanic rocks may also be included in this package of broken-up sediments and be incorporated into coarse conglomerates with large blocks called *olistostromes*. The blocks, which lie in a matrix of clay sediments, may be from a few metres up to several hundred metres across. The result is called a *tectonic melange*.

The sediments in an accretionary prism are subjected to strong tectonic deformation prior to deeper burial, and soft sediment deformation is a very characteristic feature of such deposits. If we look at the total package of imbricated wedges, there is a younging in the opposite direction, towards the subduction zone. This is a feature that can be used to recognise this depositional environment.

2.47 Summary

The study of sedimentary processes and facies relationships is important for the prediction of the distribution of different facies and rock properties.

We are interested in the geometry and distribution of sedimentary facies and also the internal properties of the sediments as they change during burial.

We therefore need to reconstruct sedimentary basins with respect to water depth, wave and tidal energy, and climate at different gelogical times.

The grain-size distribution and the mineralogy are to a large extent controlled by the source area being eroded and by transport processes.

The primary sediment compostion with respect to mineralogy and texture is very important with respect to the diagenetic process during burial. This will deterime the rock properties recorded seismically, and also reservoir quality. Sedimentology provides an important basis for basin analyses and for predicting rock properties, such as porosity, ahead of drilling. It is also very close to sequence stratigraphy. See chapter 7.

Much of the interest in applied sedimentology was traditionally concentrated on predicting the distribution of reservoir rocks, mainly sandstones and limestones, and their properties. Now with the increased interest in unconventional oil and gas, shales have become very important, and the mineralogy is very critical.

The changes from clay to mudstones and shales can not be understood without detailed clay mineralogical analyses and geochemistry. The primary mineralogy and textural relationships determine the physical properties during burial in sedimentary basins. This is important for the interpretation of seismic data.

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