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 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

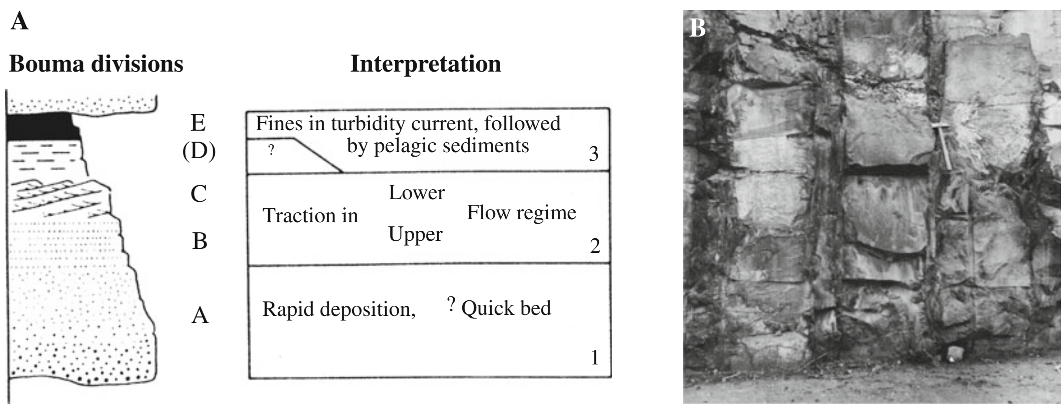

Fig. 2.12 (**a**) Bouma sequence in turbidites. (**b**) Turbidites in a Late Precambrian sequence at Lillehammer, Norway. The sequence is younging to the *right*

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 clays" 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 8000 years ago, involving about 3500 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:

$$\tau = C + (\sigma_{\mathrm{v}} - P)\tan\varphi,$$

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°). $(\sigma_{\mathrm{v}} - P)$ is the effective stress. When the pore pressure, P, approaches the weight of the overlying sediments (σ_{v}), the friction component $(\sigma_{\mathrm{v}} - P)$ 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 due to cohesion in the rest of the clay. 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.

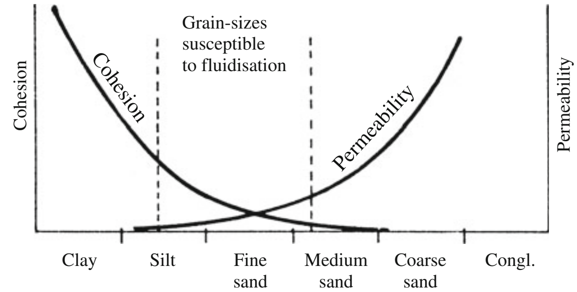

Fig. 2.13 Relation between permeability and cohesion in sediments. In coarse-grained sediments water escapes quickly, preventing build-up of overpressure, and in fine-grained clayey sediments the cohesion prevents mobilisation

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.