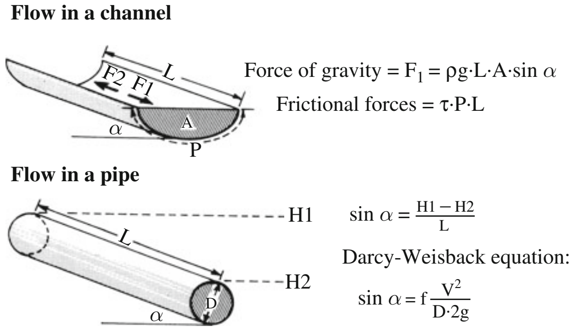

Fig. 2.7 Flow of water in channels is controlled by the ratio between the gravitational forces and the shear stress against the bottom of the channel

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:

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

or

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

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 ($\tau = c \cdot v^2$).

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 Cherzy equation and C is the Cherzy 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:

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

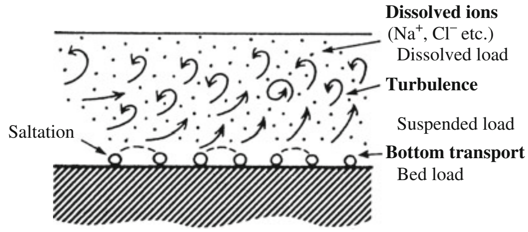

Fig. 2.8 Different forms of transport in water. Sediment grains may be carried in suspension if the vertical component of the turbulence is equal to the falling velocity of the grains. Larger grains are carried along the bottom due to the shear stress

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), which 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 sediment particles along stream

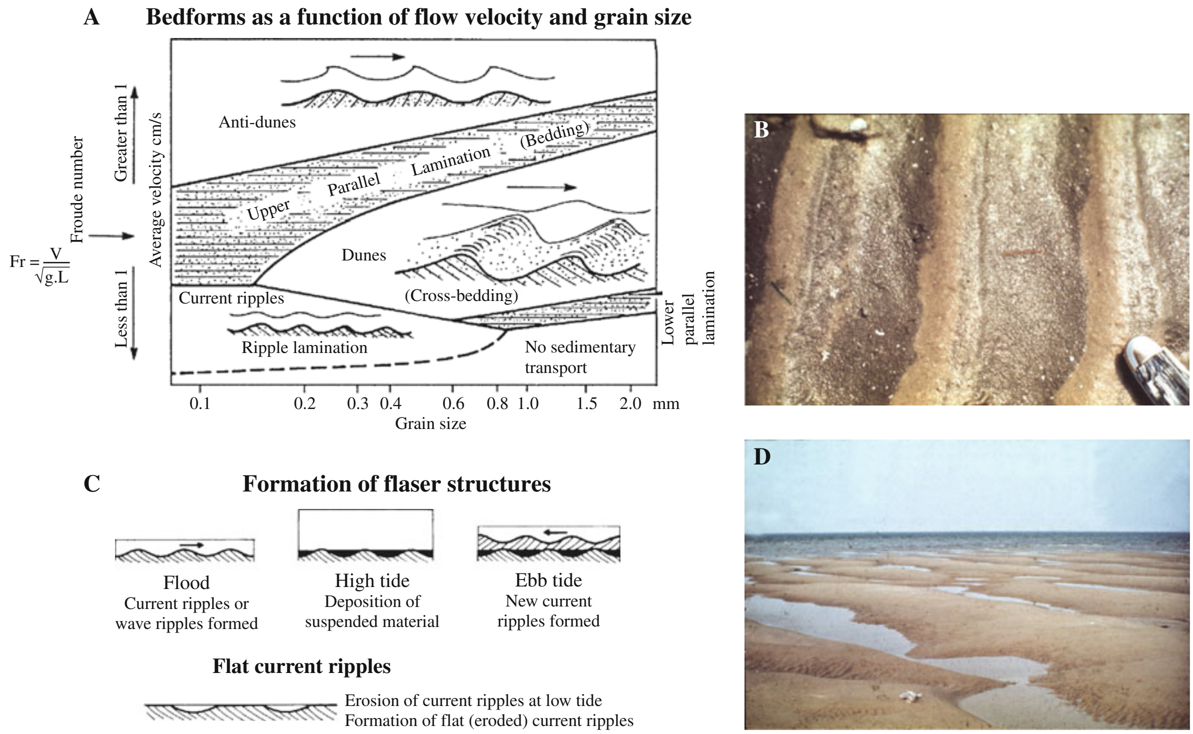

Fig. 2.9 (**a**) Sedimentary structures as a function of flow velocity, grain size and water depth. The Froude Number (F) is an expression of the velocity as a function of depth. (**b**) Ripples with clay pellets accumulating between the ripple crests. (**c**) Formation of current ripples and truncated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

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 (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.

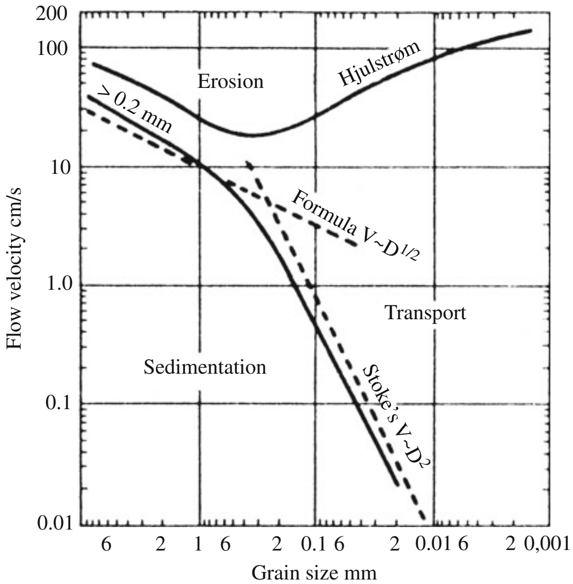

Fig. 2.10 Hjulstrøm's curve showing relations between grain size, flow velocity, erosion and sedimentation. Fine-grained sediments stay in suspension and are transported at low velocities, but require higher velocities to erode than silt due to cohesion

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.

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

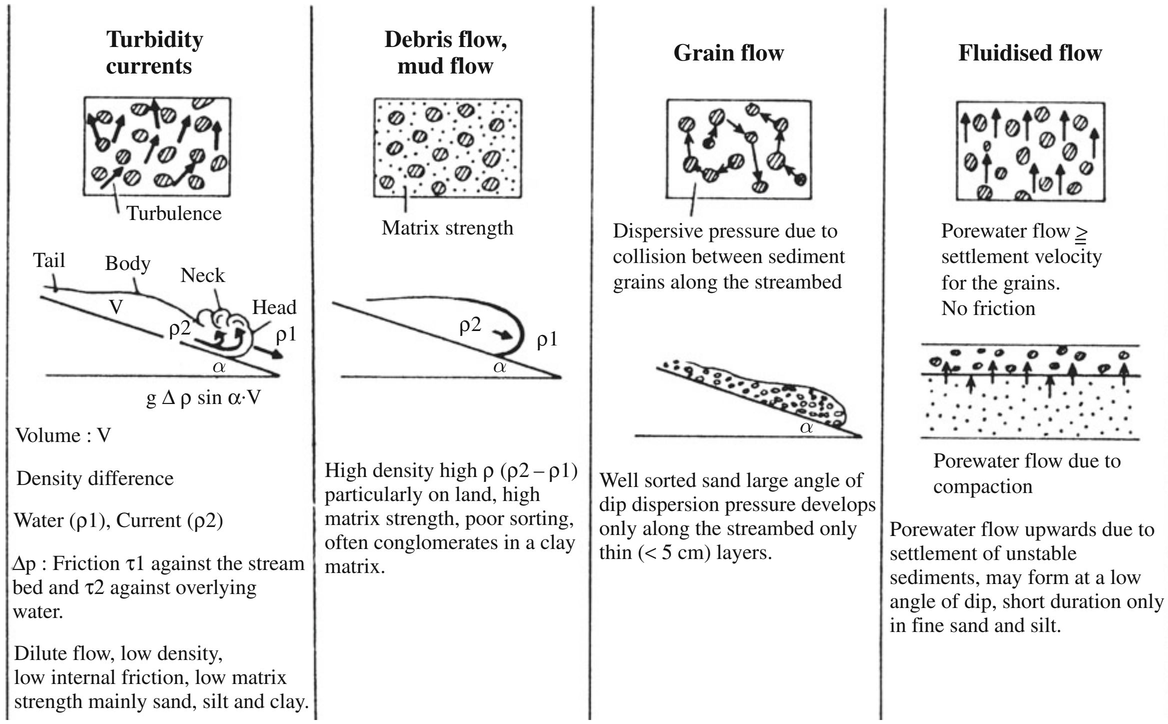

Fig. 2.11 Sketch showing different types of transport on slopes

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 $\left(\tau = cv^2\right)$.