6 Sediment grains in fluids: settling, transport and feedback

Full from the rains, but the flood sediment gone;
Under the brace of the glancing current
Each pebble shines with a life of its own,
Electric, autonomous, world-shaking-divergent.

Hugh MacDiarmid, 'The Point of Honour', *Complete Poems*, Vol. 1, Carcanet

6.1 Introduction

Having established some basic fluid properties and principles of fluid motion, we turn now to the interaction between fluid and sediment grains. In the 1960s R.A. Bagnold gave the name 'loose-boundary hydraulics' to this field of study, distinguishing it from the analysis of pure fluid motion alone. We distinguish two types of sediment bed: granular/cohesionless and cohesive. The first type includes all boundaries made up of solid grains that are kept in contact with adjacent grains purely by gravitational effects. The second type applies most commonly to clay mineral aggregates on mud beds where the tiny clay mineral flakes are mutually attracted by electrostatic forces, which may be large compared to gravitational ones. We shall concentrate in this chapter on the former case.

6.2 Fall of grains through stationary fluids

Sediment particles falling through static or very slowly moving water and air masses are common in Nature, from silt in river-mouth plumes to dead (and dissolving) pelagic shells sinking through the oceans. If we introduce sediment grains of density σ singly into a static liquid of density ρ such that $\sigma > \rho$, then the spheres will initially accelerate through the fluid, the acceleration decreasing until a steady velocity known as the terminal or fall or Stokes velocity (U_S) is reached. A typical graph of experimental data for smooth quartz spheres in water is shown in Fig. 6.1, showing clearly that the fall velocity increases with increasing grain diameter but that the rate of increase gets less.

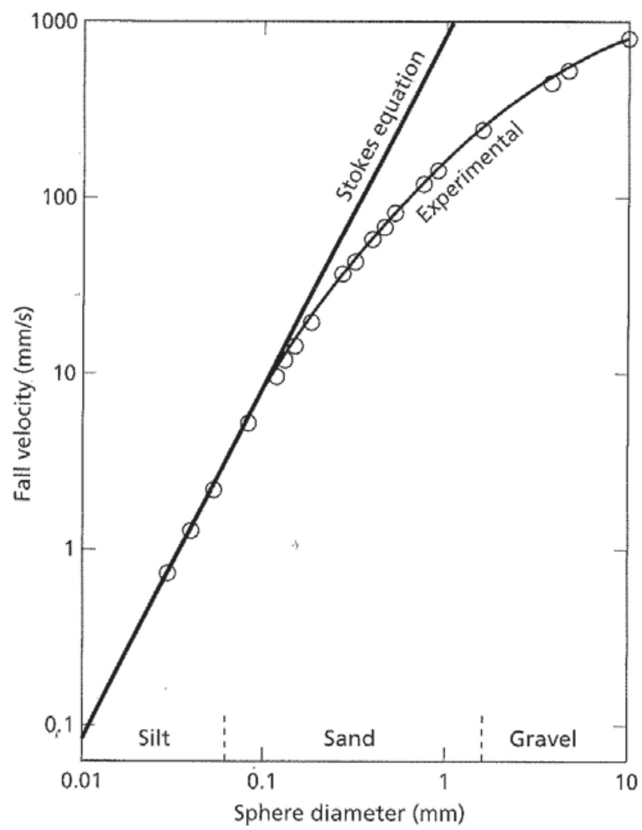

Fig. 6.1 Fall velocity as a function of grain diameter for quartz spheres in water at 20 °C compared to the predictions of Stokes' law. (After Gibbs *et al.*, 1971.)

Accurate prediction of the rate of fall of solids through stationary fluids is only possible for one highly specialized case—that of the *steady* descent of *single, smooth, insoluble spheres* through a *still Newtonian fluid* in *infinitely wide and deep* containers when the *grain Reynolds number is low* (< 0.5) and hence viscous flow separation does not occur. The prediction was first made by G. Stokes, the 19th-century English physicist. The reader should have noted the eight or more restrictions to the Stokes analysis (*italicized* above), perhaps concluding that

Further information on Stokes' law

There are two ways of approaching this problem. An engineer's or experimentalist's approach is to group together the variables that might control the velocity of fall and to do dimensional analysis. Thus we might surmise that the fall velocity will vary directly with density contrast, $\sigma - \rho$, size as diameter or radius, d, and gravity, g. We might expect it to vary inversely with fluid viscosity, μ. Thus we could write a proportionality relationship of the general form:

$$U_S \propto \frac{(\sigma - \rho)dg}{\mu} \quad \text{(B6.1)}$$

This leaves the determination of the proportionality constant and any nonlinear behaviour to experimentation. The approach of the mathematical physicist (originally G. Stokes) is to solve the problem theoretically by recourse to the fundamental equations of flow and potential flow theory, both topics we have lightly touched upon in Chapters 4 and 5. Stokes considered viscous fluid resistance forces only and arrived at a theoretical relation that expresses fall velocity as a function of grain and fluid properties. During steady fall, all acceleration terms in the equations of motion vanish, leaving a balance between the solid grain's immersed weight force F_w and the viscous drag force F_v acting over the solid grain surface. (Derivation of F_v from first principles is a lengthy business and beyond the scope of this book. Faber (1995) gives a good account, although the reader will need some higher mathematics to understand it.) We now have:

$$F_w = F_v \quad \text{(B6.2)}$$

or

$$\tfrac{4}{3}\pi r^3(\sigma - \rho)g = -6\pi\mu r U_S \quad \text{(B6.3)}$$

where r is the grain radius, μ is the molecular viscosity and U_S is the fall velocity. This may be rearranged to find the fall velocity, frequently called the Stokes velocity in honour of its discoverer:

$$-U_S = \frac{2}{9}\left(\frac{r^2(\sigma - \rho)g}{\mu}\right) \quad \text{(B6.4)}$$

Note that U_S changes sign for buoyant systems where $\sigma < \rho$. As noted previously Stokes' law only accurately predicts the fall velocity of particles whose particle Reynolds number (where the velocity term is the full veocity) is < 0.5. This corresponds to silt-sized and finer, quartz-density particles in water. At higher grain Reynolds numbers the viscous resisting force $-6\pi\mu r U_S$ becomes an underestimate, as marked accelerations of the bounding fluid and flow separation effects (Fig. B6.1) begin to occur. No satisfactory theoretical solution is now possible, so recourse has to be made to experimental data on rates of fall and on drag coefficients. In the Stokes range we may equate the two forms of expression for the surface drag force. Thus:

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Further information on Stokes' law [*continued*]

Fig. B6.1 Visualization of flow around a sphere to show the onset of turbulent flow separation. (From van Dyke, 1982.)

$$6\pi\mu r U_S = C_D \pi r^2 \rho \frac{U_S^2}{2} \quad \text{(B6.5)}$$

or

$$C_D = \frac{24}{Re} \quad \text{(B6.6)}$$

for the drag coefficient. Outside the Stokes range C_D must be gained from experiment.

the formula is practically useless. Thus natural silicate grains or calcareous biological debris are not spherical, usually fall in a group and the latter are soluble. Nevertheless, we may make use of Stokes' formula as a good approximation in some natural situations.

In many natural systems, other problems arise because of the presence of rough, nonspherical particles and multigrain settling. The problem of nonsphericity must be solved by recourse to specific experiments, particularly when fragments of biological grains at high grain Reynolds numbers (Re_g) are involved. In multigrain settling the fluid streamlines relative to the individual grains interact. Increased drag then results in a decrease of the grain fall velocity relative to its velocity in a grainless fluid. It has been proposed (Richardson & Zaki, 1958) that the fall velocity, U'_S, of a spherical particle in a dispersion of other falling grains varies as

$$U'_S = U_S(1 - C)^n \quad (6.1)$$

where U_S is the fall velocity of a single grain in an otherwise grainless fluid, C is the volume concentration of grains in the falling dispersion, and n is an exponent varying between 2.32 (for $Re_g > 1$) and 4.65 (for $Re_g < 1$). This relation tells us that the fall velocity of a grain in a dispersion will be smaller than that in an otherwise grainless fluid and strongly dependent upon concentration. For fine sediment, when $n = 4.65$, and at a high value of C around 0.5, U'_g may be only a few per cent of U_g. This consideration is of great importance in understanding the settling behaviour of dense, sediment-laden flows such as turbidity currents. Further interesting complications arise:

- when a grain is soluble such that its size and mass decrease (but not necessarily linearly) during fall;
- when a grain falls through a density-stratified fluid;
- when a grain falls through or into a Bingham fluid whose yield strength may exceed the applied gravitational force due to the grain itself.

Each of these situations is common in Nature. Further examples will doubtless spring to the reader's mind.

6.3 Natural flows carrying particulate material are complex

The sedimentologist studies natural flows and the majority of these are complex in that they contain various types of particulate matter that influence fluid behaviour under conditions of shear. Three end-members may be defined, although natural 'dirty' flows may feature several kinds of behaviour:

1 In fresh water or seawater transporting cohesionless, non-electrically-charged grains of silicate minerals, additional resisting stresses may arise through grain-to-grain interactions, leading to an increase in the apparent viscosity of the two-phase system (Bagnold, 1956, 1966b).

2 In freshwater flows transporting cohesive particles, Bingham-like behaviour occurs, with significant increases in apparent bulk viscosity directly related to the amount of transported clays and rate of shear (Wan, 1982). Such behaviour is to be expected if the cohesive particles behave as rigid aggregated particles analogous to colloids. The fluid must flow around the rigid aggregate rather than through it and hence the viscous dissipation of the flow must increase (Witten, 1990).

3 Seawater flows transporting dilute (0.15–10 g/L), flocculated, organic-rich aggregates of clay minerals have been found to maintain a Newtonian flow structure and show no noticeable gradient of sediment concentration towards the wall (Gust, 1976). The thickness of the wall layer may be enhanced by a factor of 2–5 and the friction velocity decreased by up to 40%. This drag-reducing behaviour is the opposite of that expected if the clay particles had caused a small increase in fluid viscosity and has been postulated to be a naturally occurring analogue of plain water flows to which have been added small amounts of polymer molecules (Gust, 1976; Best & Leeder, 1993).

Any reduction in fluid drag exerted adjacent to the bed by smooth boundary flows must come about by decreases in some or all of the following:

- apparent molecular viscosity;
- velocity gradient in the viscous sublayer;
- Reynolds stress production.

The first effect is the opposite to that expected if small amounts of clay are added to a flow since a small increase in molecular viscosity will result. The other causes of drag reduction have been widely used to manipulate turbulent boundary layers. Studies with drag-reducing polymers (the Toms effect; e.g. Tiederman *et al.*, 1985) have revealed that drag reduction is achieved through modification of the near-wall turbulence structure within the buffer region of the flow, and that the viscous sublayer plays a relatively passive role. This indicates that drag reduction is inherently linked to changes in the mechanisms of turbulence production within the buffer region; turbulent momentum exchange is evidently reduced in subsequent interactions with the outer regions of the boundary layer.

6.4 Fluids as transporting machines

The transport of sediment grains by shearing fluid must be due to exchange of momentum between grain and fluid, i.e. forces are set up during the transport

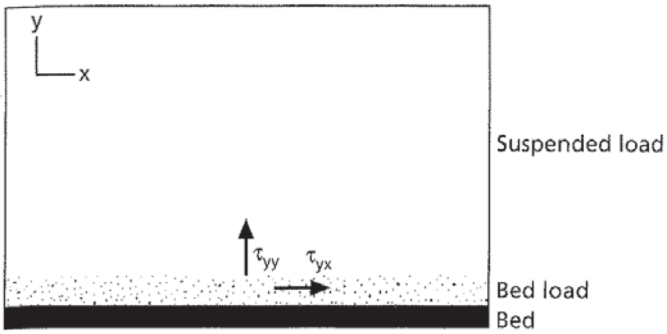

Fig. 6.2 Simple definition sketch to show the two stresses responsible for bedload and suspended-load transport.

process. Two components of the stress tensor τ_{ij} are chiefly involved: τ_{yx} drives the transport of grains close to the bed via the production of lift and drag forces; τ_{yy} supports the load in the remainder of the flow. Figure 6.2 illustrates this division of labour.

Working from fluid dynamic principles as discussed in earlier chapters, we might expect the following:

1 In order to move a layer of stationary grains as bedload, it must be sheared over the layer below. This process involves lifting the immersed mass of the topmost layer over the underlying grains as a dilation, and hence work must be done to achieve this result.

2 The energy for the work of bedload transport must come from the kinetic energy of the shearing fluid boundary layer; whether the fluid is in a state of laminar or turbulent flow is immaterial.

3 Close to the bed it might be expected that the fluid momentum transferred to the moving grains will in turn be transferred to other stationary or moving grains so that a dispersion of colliding grains will eventually evolve. The efficacy of grain collisions will depend upon the immersed mass of the grains, the dynamic friction coefficient and the viscosity of the moving fluid. We will thus expect major differences between wind and water transport.

4 If grains are to be transported in the body of the fluid, then some fluid mechanical mechanism must act to effect the transfer of grains from the bed layers. This mechanism must be sought in the processes of turbulent shear, chiefly in the burst and sweep motions outlined in Chapter 5.

The fact that fluids may do useful work is obvious from their role in powering waterwheels, windmills and turbines. In each case the kinetic energy becomes the mechanical energy of the machine in question. Energy losses occur, with each machine operating at a certain efficiency as a result. For each case:

work rate = available power × efficiency

Applying these basic principles to a river or delta channel—perhaps the simplest natural systems to begin with—the concept of useful work is now replaced by natural work. The river will try to transport the sediment grains supplied to it by hillslope processes, tributaries and bank erosion. How much sediment can be carried? This will depend upon the power available to the river, the local power supplied to the channel boundaries and the efficiency of the energy transfer between fluid and grain. As we saw in Chapter 5, flow power is made available as the fluid potential energy is converted to kinetic energy down a gravity slope. This power is available to erode and transport sediment grains. However, as with any machine, only some portion of this power is utilized in the mechanical process of doing work, i.e. the river may be said to operate at some efficiency, *e*.

Useful as the concept of flow power is for channelized water flows, there are problems in other systems:

1 The wind has no readily definable upper flow limit.

2 The availability of sediment is not uniform, and it depends upon delivery.

3 Nonequilibrium effects may be very important. Clays or very fine silt grains have settling velocities so low that they may be carried far downstream by a flow before they can settle to the bed. Thus the flow may be said to be oversaturated with sediment.

6.5 Initiation of particle motion

As fluid shear stress over a levelled plane bed is slowly increased, there comes a critical point when grains begin to be moved downstream with the flow. Sediment transport has begun. A particular fluid shear velocity above the threshold for motion may be expressed as a ratio with respect to the critical threshold velocity (u_{*C}) for the grains in question. This is the transport stage (Francis, 1973), defined as the ratio u_*/u_{*C}.

A great deal of attention has been paid to the determination of the critical threshold for grain movement since it is an important practical parameter in civil engineering schemes (canals, irrigation channels, model experiments). A knowledge of the threshold value for different grain types and sizes is also of

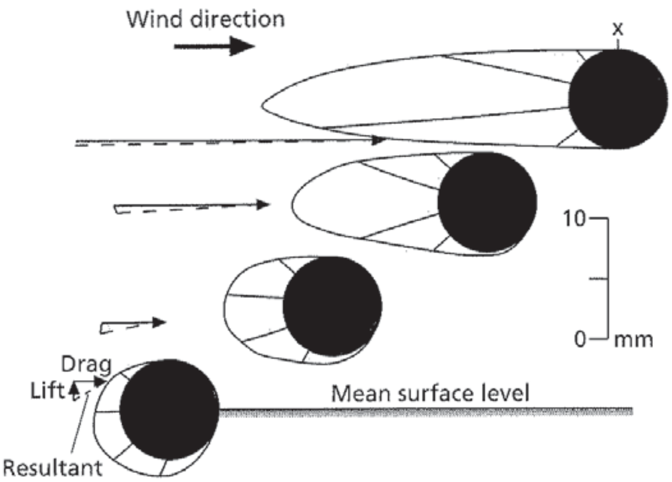

Fig. 6.3 The pattern of approximate pressure differences measured in experiments between position X on top of a 7.5 mm sphere and other positions on the sphere at various heights in a windstream. The lengths of the lines in the shaded areas outside the spheres denote the relative differences in air pressures. Both lift and drag forces act on the sphere, but lift decreases rapidly with height whereas drag increases because of the direct pressure of the wind. The wind velocity at 20 mm above the surface in the experiments was 7.7 m/s; shear velocity was 0.98 m/s. (After Chepil, 1961.)

special interest to the sedimentologist interested in determining conditions when erosion or deposition will occur and in establishing estimates for the magnitude of ancient currents.

Grains of silt, sand or gravel comprising natural sediment beds are acted upon by a pervasive surface shearing stress as any fluid, be it laminar or turbulent, moves over them. If τ_0 is the mean bed shear stress, then the mean drag per grain is given by τ_0/n, where n is the number of particles over unit bed area. A lift force due to the Bernoulli effect also exists. Fluid streamlines over a projecting grain will converge, the velocity will increase and therefore, to maintain the energy equilibrium, the pressure must decrease above the grain. Chepil's (1961) (Fig. 6.3) and later results leave no doubt that the lift force is comparable to the drag force when the grain is on the bed. The lift force rapidly dies away and the drag force rapidly increases as the grain rises from the bed.

Both fluid forces considered above will try to move bed grains; they are resisted by the grain's normal weight force. A useful way of nondimensionalizing these threshold forces is to express the ratio of applied shear to normal weight forces, grain motion occurring when the fluid force exerts a critical moment about the pivot point of the stationary grain with its neighbour (Fig. 6.4).

It is difficult to determine theoretically the critical applied shear stress for grain motion, despite the initial attractions of using this 'moments of force' approach. This is because there are a large number of variables involved, one of the most unpredictable

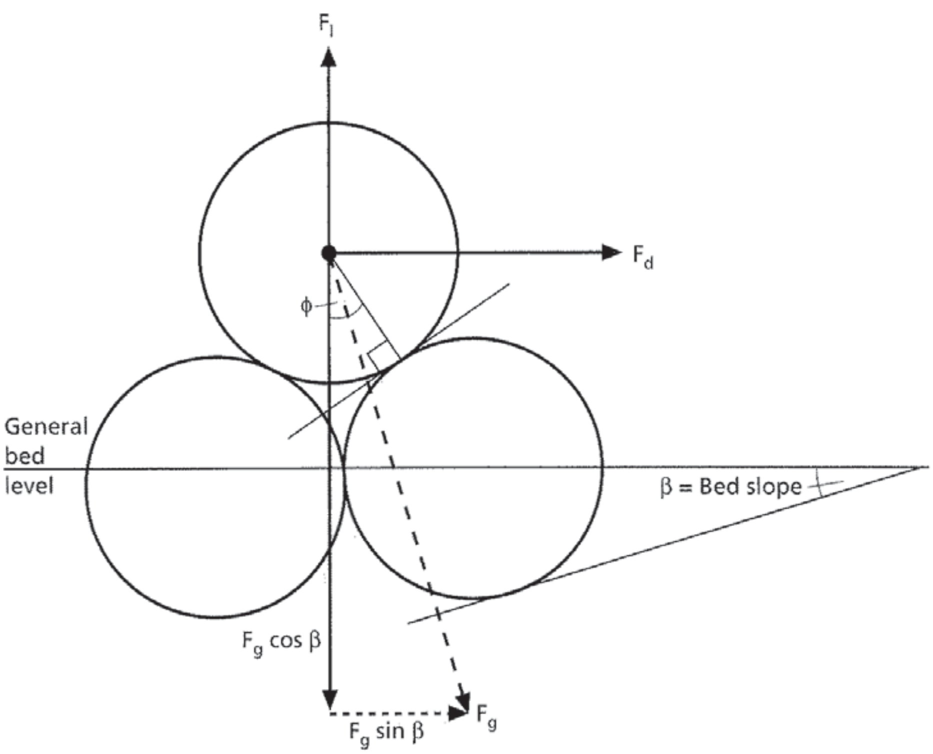

Fig. 6.4 Force moment balance diagram for the entrainment of a single sediment grain pivoting at angle ϕ about an adjacent like-sized grain on a bed sloping at angle β. F_d, F_l and F_g are the drag, lift and gravity forces. At equilibrium, $F_g \sin \beta + F_d = (F_g \cos \beta - F_l) \tan \phi$ See Bridge and Bennet (1992) for exhaustive discussion.

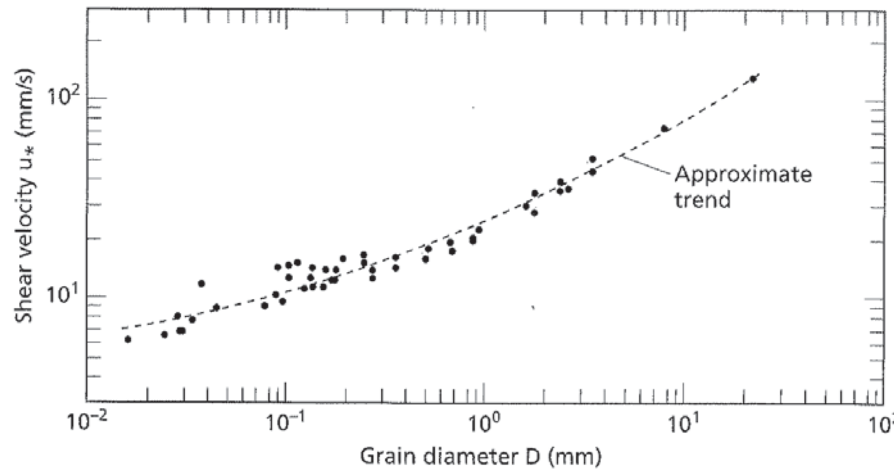

Fig. 6.5 The variation of threshold shear velocity necessary for the initiation of movement of quartz-density grains in water at 20 °C. (Modified after Miller *et al.*, 1977, by approximate trend.)

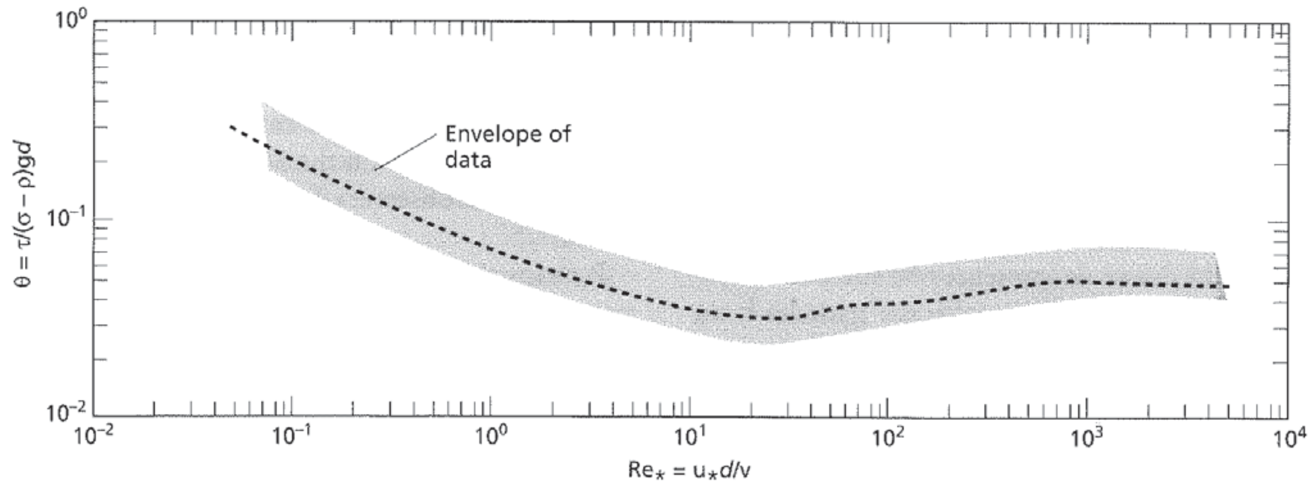

Fig. 6.6 The variation of threshold nondimensional shear stress vs. grain Reynolds number for water flow at 20 °C over a wide range of quartz-density grains. (Modified after Miller *et al.*, 1977, by envelope of data.)

being the degree of exposure of individual grains, no matter how carefully a flat bed is prepared. It is also difficult to estimate the lift force contribution. The critical conditions for the initiation of particle motion must therefore be determined experimentally. The simplest plot involves just a measure of flow velocity vs. grain diameter (Fig. 6.5), but, in order to have the greatest generality, experimental results must be applicable to a wide range of fluids and particles. As sedimentologists we are most interested in natural mineral grains in air and water, but these systems must be treated as special cases of a more general application. Working from first principles we might expect the critical conditions for particle motion, C_c, to be dependent upon gravity (g), grain size (diameter (d) or radius (r)), immersed weight $(\sigma - \rho)$, fluid kinematic viscosity (ν) and bed shear stress (τ_0). Thus:

$$C_c = f(d, g, (\sigma - \rho), n, \tau_0) \qquad (6.2)$$

Now, great generality will result if we arrange these quantities into dimensionless groups for the purpose of plotting experimental results:

$$\theta_c = \frac{\tau_0}{gd(\sigma - \rho)} = f\left(\frac{u_* d}{\nu}\right) = f\left(\frac{D}{\delta}\right) \qquad (6.3)$$

where θ_c is the critical dimensionless bed shear stress. As we surmised previously, this is the useful ratio of applied shear stress to resisting grain stress. The third expression is a grain Reynolds number, Re_g. A plot of θ vs. Re_g (for liquids, known as a Shields diagram after the German engineer A. Shields) is shown in Fig. 6.6.

Although considerable scatter is evident in Fig. 6.6, because of the many different sets of experimental conditions and the difficulty in deciding exactly when the threshold is reached, it appears that θ is nearly

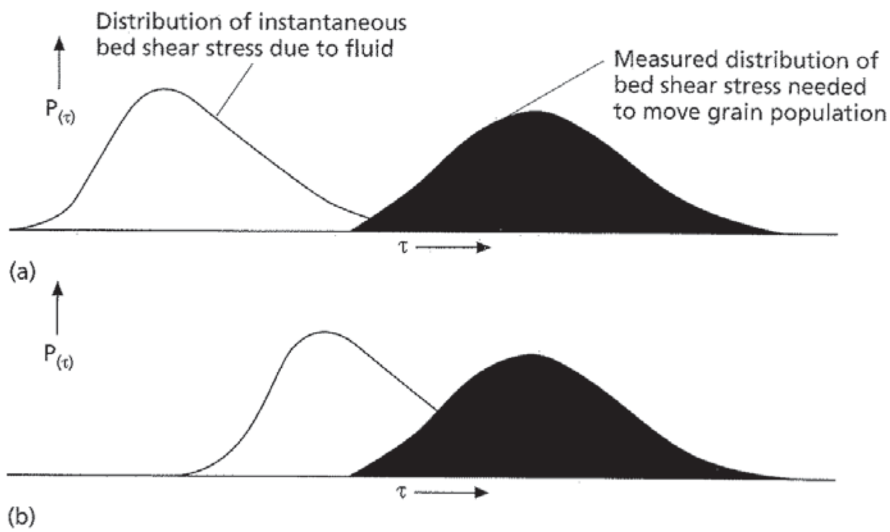

Fig. 6.7 Graphoids to show that the threshold of grain movement must be defined by the statistical overlap between the distribution of instantaneous applied shear stresses due to the turbulent fluid and the actual distribution of stresses needed to move the particular grain population. (a) Small overlap, little motion. (b) Larger overlap, general threshold exceeded. (After Grass, 1970.)

constant, at an average of about 0.05, for a wide range of grain diameters up to a grain Reynolds number of about 1.0. At low grain Reynolds numbers θ increases steadily to around 0.3. This increase is thought to be due to the presence of a smooth boundary to the flow (Chapter 5), with the particles lying entirely within the viscous sublayer. Here the instantaneous velocity gradients are less than in the lowest part of the turbulent boundary layer.

Important light has been thrown on the 'threshold problem' in water by regarding the onset of grain movement as resulting from the interaction between two statistically distributed variables (Grass, 1970). The first variable may be termed the initial movement characteristic of a given bed material in a fluid of given viscosity and density. Thus every grain on the bed is assumed to be susceptible to a local instantaneous stress, and because of the random shape, weight and placement of individual grains this τ_0, termed τ_c, has a probability distribution. The second variable is the local instantaneous bed shear stress caused by burst/sweep events (Chapter 5). These stresses also have a probability distribution, dependent on τ, fluid density, viscosity and flow boundary conditions. At the onset of grain motion the most susceptible particles (those with the lowest characteristic critical shear stress) are moved by the highest shear stresses in the shear stress distribution applied to the bed by the flow. Experimental results enable histograms to be constructed for each distribution (Fig. 6.7). These show that the critical shear stress necessary to move grains occurs when the two τ distributions overlap by a certain constant amount. Much of the scatter on the Shields plot evidently reflects different observers deciding on different degrees of overlap.

Perhaps the most intractable aspect of the 'initiation of motion problem' is the fact that many natural sediment beds are of radically mixed sizes, perhaps bimodal or even polymodal in their size distributions. A lot of experimental and theoretical research has gone into trying to come up with general expressions to cope with this common situation (Parker *et al.*, 1982; Wilcock & Southard, 1989; Paola *et al.*, 1992; Bridge & Bennett, 1992). It was initially pointed out that, for a pebbly sand, the motion of the pebbles occurred at lower critical applied fluid stress than for pebbles alone, because of the effects of greater exposure of the pebble to the shearing boundary layer and the tendency for the sand around the pebble to be scoured, thus elevating the pebble into the boundary layer and transporting sand at the same time. This has been named the condition of 'equal mobility' at threshold. However, the situation rarely reaches equilibrium because close to threshold the bed quickly 'armours' once a certain amount of the sand has been so scoured away, that is, a surface layer of interlocking and immovable pebbles protects the underlying sand–pebble mixture. It is only well above the threshold that equal mobility conditions seem to apply. The process of upstream armouring and a consideration of the great increase in turbulence intensity, and hence potential for suspension of sand, over mixed-size rough beds also serves to explain the almost universal trend found in rivers for downstream fining (see Paola *et al.*, 1992).

6.6 Initiation of motion by air flow

In air flows Bagnold (1954b) defined two types of threshold. At a critical value of air speed insufficient to move bed grains by fluid shear alone, grain motion could be started and propagated downwind by simply letting sand grains fall on to the bed. Other grains were bounced up into the airstream, which, upon falling, caused further movement as they impacted on to the bed, and so on. Grain motion ceased as soon as the introduction of artificial grains stopped. The critical wind speed necessary for this process was termed the *impact threshold*. Further increase of air speed enabled grains to be moved by the direct action of the wind at the fluid or *normal threshold*. The ability of natural sand grains in air to disturb and eject other grains after impact contrasts with the situation in water. This is because of the viscosity contrast between air and water, which controls resistance to motion, and the great effective density contrast between quartz and air (2000 : 1) compared to quartz and water (1.65 : 1). As we shall see later (Chapters 7 & 8), the types of ripples developed in the two fluids depend intimately upon these factors. For most sands in air flows the critical shear velocity is a function of the square root of the particle size.

6.7 Paths of grain motion

Once in motion, other contributions to fluid surface forces exist. Shear lift occurs due to relative grain motion through a velocity gradient. Spin lift occurs when a grain undergoing relative motion in a fluid rotates—the Magnus effect is responsible. Once the threshold for motion is exceeded, grains move downstream in three basic ways (Fig. 6.8).

1 *Rolling*. Rolling motion is simply defined as continuous grain contact with the bed and it includes the rarely observed 'sliding' motion. In air, rolling grains form a 'creep carpet' kept in motion by the kinetic

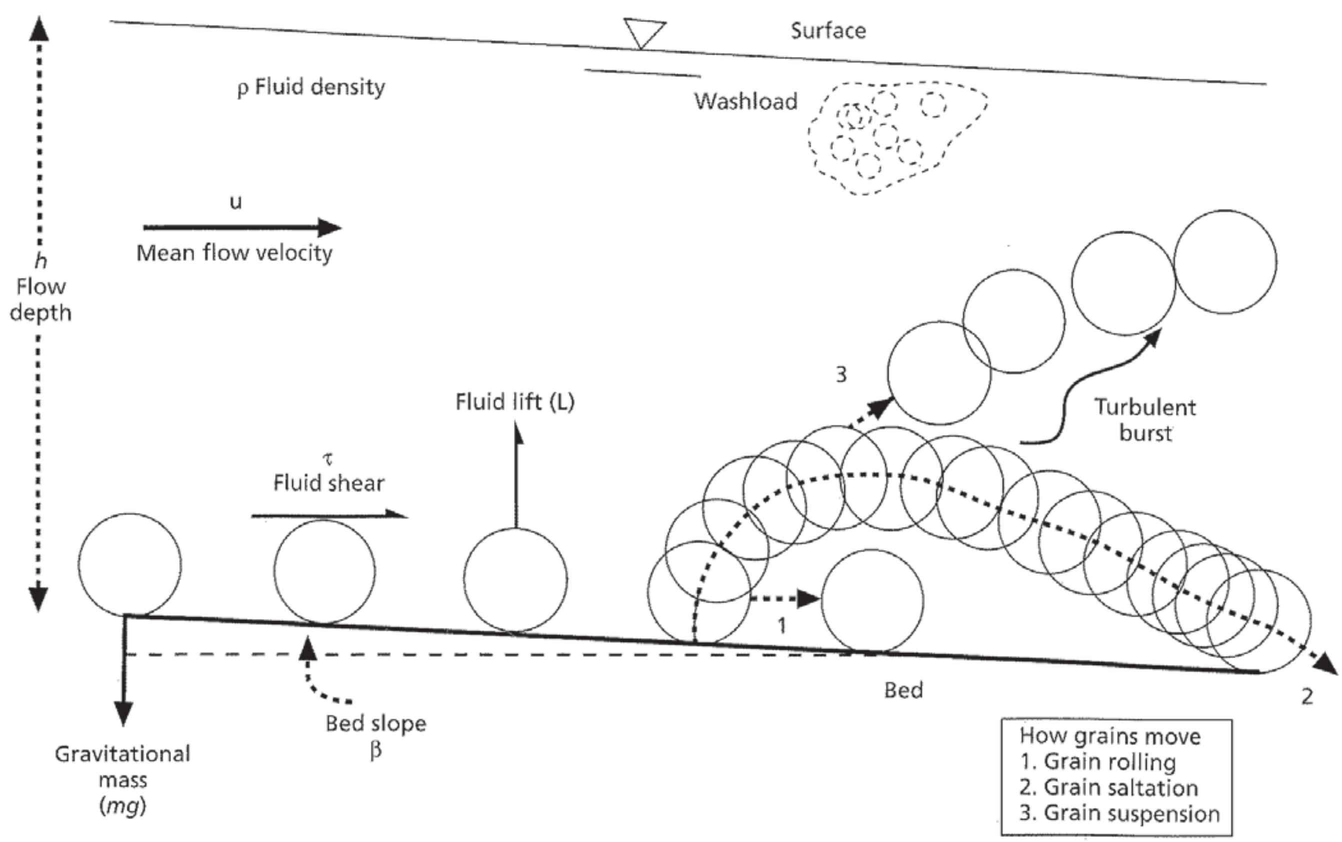

Fig. 6.8 Cartoon to show the various ways that grains are transported in water and air flows.