CHAPTER 10

MOVEMENT OF SEDIMENT BY WATER FLOWS

INTRODUCTION

1 A simple flume experiment on sediment movement by a unidirectional current of water in a flume serves to introduce the material in this chapter. Place a layer of sediment in the flume, level it to have a planar surface, and establish a uniform flow at a certain depth and velocity. Gradually, in steps, increase the strength of the flow beyond the condition for incipient movement. The magnitude of the flow strength relative to what is required for incipient movement of the bed sediment is conventionally called the ***flow intensity***, and is usually taken to be the ratio τ_o/τ_{oc} (or, what is the same, u_*/u_{*c}), where the subscript c denotes the threshold ("critical") condition.

2 At first the particles move as ***bed load***, by hopping, rolling, and/or sliding. Particle movement is neither continuous nor uniform over the bed: brief gusts or pulses of movement affect groups of particles locally, and seemingly randomly, on the bed. Particles move a short distance, stop, and then move again. Even when they are moving, they are generally not moving as fast as the fluid near the bed surface.

3 As the flow becomes stronger, some of the particles moving near the bed are lifted upward by upward-moving turbulent eddies and travel for more or less long distances downstream as ***suspended load***. The stronger the flow and/or the finer the sediment, the greater is the concentration of suspended sediment, the higher it can travel in the flow, and the longer it moves downstream before returning to the bed. Of course, the particles are not really suspended in the way that a picture is suspended on the wall by a nail; they are continuously settling through the surrounding fluid, and eventually they return to the bed. If the sediment is fine and the flow is strong, however, the particles are likely to travel for the entire length of the flume.

4 If you introduce a small quantity of very fine clay-size sediment into the flow, you would find that it too travels in suspension, but the essential difference between this part of the suspended load and the coarser part you observed before is that even if you add large quantities of it to the flow, it would not be represented in the bed. Fine sediment of this kind is called ***wash load***. Extremely fine particles, in the size range of small fractions of a micrometer, can be kept in effectively permanent suspension, because their mass is so small that they can be moved about by the random bombardments of the molecules constituting the fluid itself. These random motions are a manifestation of Brownian motion.

5 For flow intensities not much above threshold, it is fairly easy to observe the particle motions in the bed load, provided that you have clear water, good lighting, and sharp eyes (close-up slow-motion vision would be a big help), but as the flow intensity increases, the concentration of particles in motion as bed load increases, and it becomes difficult or impossible to observe the motions of individual particles. Unfortunately, no one yet seems to have devised a good way to see into the dense layer of moving bed-load particles at high flow intensities to study its characteristics. This important aspect of sediment transport remains contentious and inadequately studied.

6 To gain an appreciation of a rather different mode of sediment movement, you need to resort to a wind tunnel. It is not difficult to build one: all you need to do is construct a rectangular duct resting on the floor, leading from a flared entrance at the upwind end to a large empty chamber at the downwind end, with an exhaust fan in the side of the chamber to create a wind through the duct. A louver just downwind of the fan lets you adjust the wind velocity. Especially when the ratio of sediment density to fluid density is very large, as with quartz sand in a wind tunnel, sediment particles are entrained impulsively by the flow at middling to steep take-off angles and move downstream in long arching trajectories little affected by the fluid turbulence to make impact with the bed at low angles. This characteristic mode of movement, known as ***saltation***, is especially important in the transport of sand by wind. Its manifestation in transport of particles that are not much denser than the transporting fluid, however, is much less striking or distinctive.

THE BED, THE FLOW, AND THE LOAD

7 The aggregate of sediment particles being transported by a flow at a given time is called the ***load***. At the very outset, it seems appropriate to define what is meant by the bed, the flow, and the bed load (Figure 10-1). (I am not sure that the following definitions, intuitive as they seem to me, would be approved by all specialists in sediment transport.) The *bed* comprises all of the particles that at a given time are motionless and in direct contact with the substrate, and the *load* comprises all of the particles that are in motion in a given flow, whether or not they are in contact with the bed. That leaves the less certain definition of the *flow*: all the material, fluid and solid, that at a given time are in motion above the bed.

8 The load can further be subdivided in two different ways. On the one hand, the load can be divided into ***bed-material load***, which is that part of the load whose sizes are represented in the bed, and ***wash load***, which is that part of the load whose sizes are not present in the bed in appreciable percentages. The wash load, which if present is always the finest fraction of the load, is carried through a reach of the flow without any exchange of sediment between the bed and the flow. On the other hand, the load can also be divided into ***bed load***, which travels in direct contact with the bed or so close to the bed as not to be substantially affected by the fluid turbulence, and ***suspended load***, which is maintained in temporary suspension above the bed by the action of upward-moving turbulent eddies. I hope that it is clear from these definitions that bed load is always bed-material load, and suspended load is likely to be partly bed-material load and partly wash load, although in particular cases it could be all wash load, or all bed-material load. Confused? Figure 10-2 may or may not be of help.

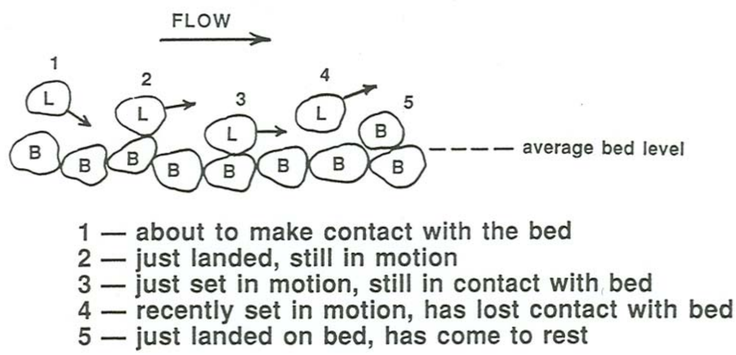

Figure 10-1. The flow, the bed, and the load.

9 The movement of bed load is sometimes called ***traction***. Bed-load movement can be by rolling, sliding, or hopping. Words like those three are not entirely adequate for the task of describing the nature of bed-load movement, however, because the moving particles commonly partake of all three "modes", which vary in importance from movement event to movement event, and from instant to instant during a movement event. It is not easy to observe bed-load movement in great detail, but when you have the chance to watch a carefully made high-speed close-up motion picture of bed load (in flows where the load is not yet so abundant as to obscure one's view) you see that the particles characteristically take occasional excursions downstream, by rolling and hopping along irregularly, and then come to rest for some time before being moved again.

10 Particle shape has a substantial influence on mode of bed-load movement: disk-shaped particles have a much greater tendency to slide or bulldoze, whereas equant or spheroidal particles have a much greater tendency to roll or hop. Discoidal particles can under some conditions be seen to roll like cartwheels!

11 There is clearly a problem in distinguishing between bed load and suspended load: how far can a particle move up into the flow and still be considered bed load? The standard criterion is whether or not fluid turbulence has a substantial effect on the time and distance involved in the excursion. It is important to keep in mind that there is no sharp break between bed load and suspended load: a given particle can be part of the bed load at one moment and part of the suspended load at another moment, and not moving at all at still another moment. The consequence of this is that at any given time there is an appreciable overlap in the size distributions of the bed load and the suspended load, although obviously the suspended load tends to be finer than the bed load. Moreover, there seems to be no sharp break, or jump discontinuity, in the volume concentration of sediment upward from the bed-load layer into the suspended-load layer (although accurate observations are not easy to make). That is to be expected, because in a sense the bed-load layer acts as the "lower boundary condition" for the suspended-load concentration; see the later section on suspension.

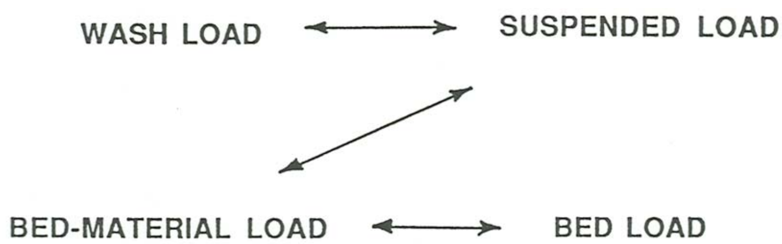

Figure 10-2. Relationships among the various kinds of sediment load.

12 The relative proportions of the bed material that moves as bed load and as suspended load depend upon the characteristics of the bed material, especially its size, and on the flow conditions. Very coarse bed material in rivers (gravel) generally moves as bed load, whereas fine to medium sands move predominantly as suspended load. The mode of movement of the coarser sand sizes generally varies depending on the hydraulic conditions: at low flow intensities the coarser sand fractions move predominantly as bed load, whereas at high flow intensities they are taken into suspension. Even at the same average discharge, sand of a given size may alternate between suspension and traction, as it is caught up by powerful eddies (for example, the separation eddies formed on the lee sides of major bed forms) or returns to the bed in less turbulent parts of the flow. Most sand sizes do not travel in continuous suspension; the very fact that these sizes constitute a major part of the bed material in most rivers indicates that they are taken into suspension only intermittently. To distinguish this mode of transport from the almost continuous suspension typical of wash load, which is generally composed of particles finer than fine sand, we could call the coarser part of the suspended load the ***intermittent suspension load***.

13 The distinction between bed load and suspended load can be made either on a practical observational basis or on a more theoretical basis with reference to support mechanisms. The practical definitions, those given above, are based on the observation that the bed load is carried in direct contact with the bed or very close to the bed whereas the suspended load is carried far above the bed. The more theoretical definitions are based on the concept (not easily applied, in practice!) that the suspended load is the part of the load that is supported entirely by fluid turbulence, and the bed load is the part of the load that is supported in one way or another by the bed itself, not by fluid turbulence. Bed-load particles that are moving in direct contact with the bed are supported, at least in part, directly by the bed, if the possible contribution of fluid lift forces is left out of account. By this definition, bed-load particles that are temporarily not in direct contact with the bed are either following a path that is largely unaffected by fluid turbulence, in consequence of having parted contact with the bed by a momentarily stronger fluid force (saltating particles fall naturally into this category) or are maintained in motion above the bed, perhaps at a distance of many particle diameters, by collisions with other particles that are part of a thick bed-load layer at high flow intensities. There will be more to say about the existence and nature of these latter bed-load layers later in this chapter.

14 Finally, here are a few words about how sediment concentration might be measured. Measuring the concentration of suspended sediment is fairly straightforward: you could imagine capturing a volume of the flow, in a snap-close bottle of some kind that does not disrupt the flow very much, and measuring the volume or mass of sediment per unit volume of the fluid–solid mixture. Provided that the measured volume is small relative to the characteristic spatial rate of change of "average" concentration (for example, you would not want your sample to integrate over a large fraction of the flow depth) but large enough relative to small-scale variations in sediment concentration related to the details of local eddy structure, your sample should provide a representative measure of the local average sediment concentration. Measuring the concentration of bed load, however, is a different matter. The bed-load layer is by its very nature thin. People attempt to measure the transport rate of the bed load (that is not a trivial matter either; see the later chapter on transport rates) but ordinarily not the bed-load concentration. Conceptually, however, it is reasonable to think about a kind of "area concentration" of bed load: the volume or mass of bed load, per unit bed area and as a time average, above a small area of the sediment bed. But I cannot provide any helpful ideas about how to measure such a quantity.

TRANSPORT MODE VERSUS FLOW INTENSITY

15 Before we go into more detail about how sediment particles move, as bed load or in suspension or in saltation, it is worth developing a rational framework for relating the various modes of movement to one another. As with so many aspects of sediment transport, it is valuable to think in terms of *regimes*: distinctive ranges of the phenomenon, characterized by modes of particle movement that differ from other ranges. In this case, such regimes have been called ***transport stages***. To develop a good framework for visualizing and assessing the results of experiments on transport stages, start by making a list of the variables that are likely to be important in determining the transport stage. The flow strength is best defined by the bed shear stress, just as it is for the threshold of movement. In contrast to the problem of movement threshold, however, the flow depth, which reflects the possible effects of outer-layer flow phenomena like large-scale turbulence (see Chapter 4), might not be ignorable, but as a first approximation suppose that the flow is characterized only by τ_o. Both particle size D and particle density ρ_s need to be included. The submerged specific weight of the particles, γ', must be included as well as the particle density

ρ_s, because of the effect of particle weight in settling, aside from the effect of particle inertia when the particles experience accelerations caused by fluid turbulence. The fluid properties ρ and μ have to be included for the usual reasons. Then

$$\text{transport stage} = f(\tau_o, D, \rho, \mu, \rho_s, \gamma') \tag{10.1}$$

and we should expect that everything about the transport stage, expressed in dimensionless form, should be expressible in terms of three dimensionless variables. Examples of such things are: the positions of boundaries or boundary zones between qualitatively different transport stages; lengths or heights of particle trajectories, nondimensionalized by dividing by the particle diameter D; or particle velocities, nondimensionalized by dividing by the shear velocity u_*.

16 One such set of dimensionless variables might be:

$(\tau_o)^o = (\rho/\gamma'\mu^2)^{1/3}\tau_o$, a dimensionless form of τ_o

$D^o = (\rho\gamma'/\mu^2)^{1/3}D$, a dimensionless form of the particle diameter D

ρ_s/ρ

The advantage of this set is that the leading variables, τ_o and D, are segregated into different dimensionless variables. An alternative would be to replace the dimensionless boundary shear stress with the flow intensity, u_*/u_{*c}. In either case, one could attempt to plot experimental or theoretical results in two-dimensional graphs for certain values of ρ_s/ρ (most importantly, quartz-density sediment in water-density fluid).

17 Figure 10-3, a very generalized version of a graph of boundary shear stress vs. particle size, makes a start at representing transport stages. In Figure 10-3, the axes are labeled in two ways: the dimensionless versions of τ_o and D mentioned above, and also actual values of τ_o and D at a water temperature of 10°C, to give a more concrete appreciation of conditions. We know at the outset that one boundary has to be present in the graph: the curve for threshold of particle motion. That is readily obtained by transforming the Shields curve (see Chapter 9) into these coordinates. Another boundary, which we consider next, is the curve for the onset of suspension in addition to bed-load movement.

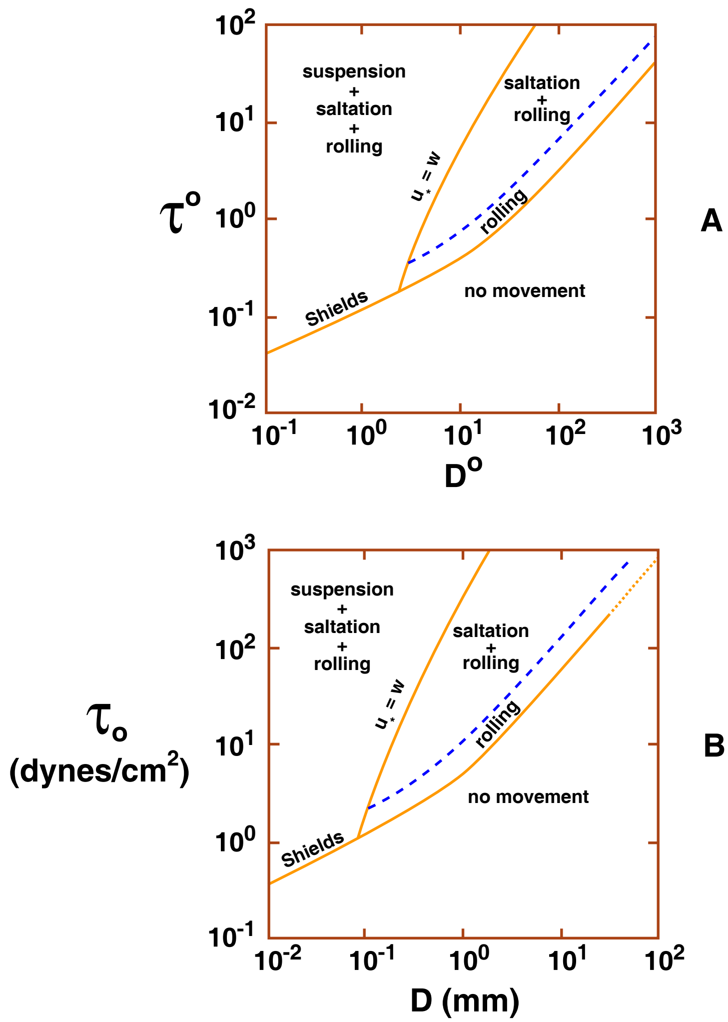

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Figure 10-3. Transport stages in **A)** a dimensionless graph of boundary shear stress vs. particle size and **B)** the same graph, but with actual values of boundary shear stress and particle size standardized to a water temperature of 10°C.

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18 The natural criterion for suspension is that the vertical turbulent velocities are at least as large as the settling velocities of the sediment particles; otherwise, particles could never be carried any higher above the bed than the entraining forces permit. The problem is that although for a given sediment size the settling velocity is fairly well defined (if effects of sorting and particle shape are ignored), the vertical turbulent velocities are distributed over a wide range of values. Should we use the very largest but very uncommon values, or smaller but more frequent values? What has commonly been done is to assume that the root-mean-square (rms) value of the vertical turbulent velocities is a good measure to use. Measurements in turbulent boundary-layer flows past both smooth and rough boundaries have shown that there is a maximum close to the bed and that the maximum values reached are proportional to the shear velocity u_* (Blinco and Patheniades, 1971). The data of McQuivey and Richardson (1969) and Antonia and Luxton (1971) show that the maximum value of (rms v)/u_* is approximately equal to one and that the value does not depend strongly on the type of roughness. An approximate criterion for the onset of suspension is then

$$u_* = w \tag{10.2}$$

19 For values of u_* less than w, there should be no suspension, and for values of u_* greater than w, some of the sediment should be traveling as suspended load. There is no reason to expect, however, that the coefficient of proportionality in Equation 10.2 is exactly equal to one; the coefficient would presumably need to be adjusted somewhat in light of actual observations on the onset of suspension. Middleton (1976) has argued that the criterion $u_* > w$ is also supported by a comparison of hydraulic measurements with the settling velocity of the largest particle sizes present in the suspended load of several rivers.

20 What remains is to convert the suspension criterion in Equation 10.2 to a corresponding curve in Figure 10-3. To do this, first write Equation 10.2 as

$$\left[(\tau_0)^0\right]^{1/2} = w^0 \tag{10.3}$$

by use of the definition of u_*. Then use the definition of the dimensionless boundary shear stress $(\tau_o)^o$, given above, and a corresponding definition of dimensionless settling velocity, $w^o = (\rho^2/\gamma'\mu)^{1/3}w$ (see Chapter 2) to obtain an expression for τ_o in terms of $(\tau_o)^o$ and an expression for w in terms of w^o:

$$\tau_0 = \left(\frac{\gamma'^2\mu^2}{\rho}\right)^{1/3}(\tau_0)^0$$

$$w = \left(\frac{\gamma'\mu}{\rho^2}\right)w^0 \tag{10.4}$$

Now substitute the expressions for τ_o and w in Equation 10.4 into Equation 10.3:

$$\left[\left(\frac{\gamma'^2\mu^2}{\rho}\right)(\tau_0)^0\right]^{1/3} = \rho^{1/2}\left[\left(\frac{\gamma'\mu}{\rho^2}\right)^{1/3} w_0\right] \qquad (10.5)$$

and simplify to obtain

$$\tau_0^{\ 1/2} = \rho^{1/2} w \qquad (10.6)$$

The final step is to use the curve for w^o as a function of D^o (Figure 3-38 in Chapter 3) to obtain the relationship between $(\tau_0)^o$ and D^o corresponding to the criterion for suspension:

$$\left[(\tau_0)^0\right]^{1/2} = f\left(D^0\right) \qquad (10.7)$$

Keep in mind that over most of its range, for settling-velocity Reynolds numbers greater than the Stokes range, the function in Equation 10.7 has to be determined by observation.

21 We see that the curve that represents the suspension criterion slopes more steeply than the curve for incipient movement. This is just a manifestation of the fact that, qualitatively, the shear stress needed to make rms *v* equal to the settling velocity with increasing particle size increases more rapidly than does the shear stress needed for incipient movement with increasing particle size. The consequence is that the two curves intersect at a certain small value of dimensionless particle size. (The suspension-inception curve does not extend downward below the movement-inception curve, because the flow there is not strong enough to move any sediment in the first place.) To the left of the intersection point, the fall velocity of the sediment particles is exceeded by the magnitude of the turbulent velocity fluctuations in the flow even at flow strengths just sufficient for sediment movement, so that sediment particles can be put into suspension as soon as they begin to be moved. Keep in mind, however, that at this and even finer sediment sizes, some of the sediment moves as bed load as well as suspended load. The existence of current ripples in sediments effectively at least as fine as medium silt is a good indication of this, inasmuch as ripples owe their existence to bed-load transport; see Chapter 11.

22 Finally, Figure 10-4, which shows transport stages in a graph of u_*/w, the ratio of to shear velocity u_* to settling velocity w vs. flow intensity u_*/u_{*c} is an equivalent form of Figure 10-3; it is just a rubber-sheeted D^o–τ^o diagram. It is neater and more synthetic than Figure 10-3, although perhaps less useful. Because $w = f(D)$ and $w \neq f(u_*/u_{*c},)$, there is a one-to-one correspondence between sediment diameter D and points on a vertical line in this graph. The movement-inception curve becomes the left-hand vertical axis ($u_* = u_{*c}$), and the suspension-inception criterion, $u_* = w$, becomes a horizontal line. The area below the line for the suspension-inception criterion represents only bed-load transport, and the area above the line represents bed-load and suspended-load transport together.

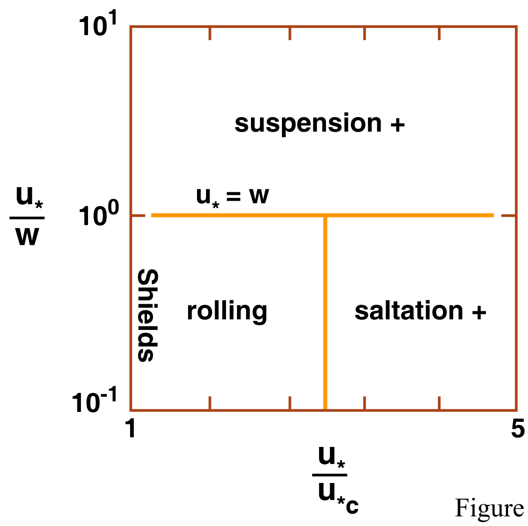

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Figure 10-4. Transport stages in a graph of u_*, the ratio of shear velocity to particle settling velocity, vs. flow intensity u_*/u_{*c}.

BED LOAD

Styles of Bed-Load Movement at Low Flow Intensities

23 Transport of the surficial particles on a locally planar bed produces a distinctive microtopography, consisting of small-scale irregular and discontinuous ridges and depressions oriented approximately parallel to the flow. The spacing of these features on a sand bed is of the order of several millimeters, and the relief is very small, generally only a few particle diameters. This lineated relief is a characteristic feature of transport, for particles ranging from silt sizes (Mantz, 1977) up to at least very coarse sand, and for hydraulic conditions ranging from just above the threshold of particle movement to high flow intensities that produce upper-regime, plane-bed conditions. It is the manifestation of the low-speeds streaks associated with the burst–sweep cycle, described in Chapter 4.

24 It does not take special experimental conditions to see manifestations of this lineated microtopography. Here are two everyday (well, almost) examples. It is a cold, gray day, and the snow has just begun to fall. Before the paved surface of the road is completely whitened, you see distinctive, shifting white streaks of snow aligned with the wind blowing across the road. Or you are standing at the kitchen sink, washing root vegetables fresh from the garden. The fine fraction of the loosened sediment is carried in suspension down the drain, never to be seen again, but the coarser fraction is immediately formed into small-scale streaks on the surface of the sink, beneath the fast-flowing water headed for the drain. Go back to the final section in Chapter 4, on coherent structures in turbulent flow, for the dynamics behind these bed-load streaks.

25 The ridges and depressions are produced by the action of small but strong turbulent eddies on the bed. Exactly how this takes place is becoming clearer as the structure of turbulence close to the bed becomes better understood as a result of numerous laboratory studies. The relevance of these observations to sediment movement has been discussed more thoroughly by Grass (1971, 1974), Karcz (1973), Jackson (1976), Sumer and Oguz (1978), Bridge (1978), Sumer and Deigaard (1981), and Leeder (1983a).

26 Most experimental observations have been made on boundaries that are dynamically smooth or transitionally rough, i.e., on boundaries characterized by a viscous sublayer that is at least poorly developed. But observations of dynamically rough boundaries (e.g., by Grass, 1971) show that even for flows without a viscous sublayer there exists a region close to the boundary, which Grass called the "inner zone", for $y^+ < 40$, characterized by distinctive low-speed longitudinal streaks and a quasi-cyclic alternation of events that has come to be known as the burst–sweep cycle (see Chapter 4). As strong vortices with axes transverse to the flow approach the boundary, they produce pressure gradients that tend to lift up the streaks and eject them into the turbulent boundary layer. This "burst" of slow-moving fluid is capable of carrying small particles away from the bed to distances a few centimeters above the bed. Sumer and Oguz (1978) found that particles whose settling velocity was of the order of $0.5\,u_*$ were carried "in a single continuous motion" up to dimensionless elevations y^+ of 100–200. The slow-moving fluid in the burst then mixes with, and is accelerated by, the fluid in the outer zone, and some returns to the bed as a fast-moving vortex or "sweep", which in turn creates a new burst, and so on. The process is not strictly periodic, although on the average it displays a period and scale controlled mainly by the velocity and thickness of the turbulent boundary layer. If the period of bursts is T and the velocity far from the bed is U, then $UT/\delta = 5$, where δ is the thickness of the boundary layer (in an open-channel flow, the depth of the flow).

27 The existence of the burst–sweep cycle suggests an explanation of the phenomena of particle movement described above. The gusts of movement of particles along the bed described by many workers since Vanoni (1964) probably correspond to fluid sweeps in the burst-sweep cycle, when velocity gradients close to the bed, and therefore shear stresses, are locally high. Particles tend to be swept to one side or the other of fast-moving fluid streaks to gather under slow-moving streaks and produce the characteristic current lineation observed on plane beds.

28 For sand finer than about 0.6 mm, movement of particles over a plane bed eventually results in minor accumulations of particles that grow to form ripples (Chapter 11). The ripples then change the pattern of flow at the bed, and therefore also the interaction of the flow and the moving sediment particles. Separation of flow takes place at the ripple crest, and the boundary layer is reestablished downstream of where the flow reattaches to the bed, on the stoss side of the next ripple downstream. Although the bed surface is now much more irregular, the same lineated surface seen on a plane bed can generally be observed on the upper stoss sides of ripples or larger bed forms, if the water is clear and there is good low-angle lighting.

29 Perhaps the least difficult way of observing how bed-load particles move as the flow intensity increases beyond the threshold of movement is to study the movement of a single particle over a bed composed of similar particles that are held in place by an adhesive. The only problem with such experiments, although not a serious one, that is that there is no exchange between moving particles and the sediment bed: while not in motion, the test particle always rest high, in a very exposed position, and can never move into a low position (often called a ***pocket***) recently vacated by another entrained particle. Such experiments have been reported by Meland and Norrman (1966), Francis (1973), Abbott and Francis (1977), and Nakagawa et al. (1980). Valuable as these single-particle studies are, however, there is no substitute for watching the motion of bed-load particles on real sediment beds. The observational difficulties are formidable: one needs clear water (no suspended load to obscure the view), a magnified close-up view, and, ideally, high-speed filming or video recording, because the rapidity with which the state of particle motion changes makes it not very productive to watch particle motions in real time.

30 Meland and Norrman (1966) distinguished three stages of particle motion:

(1) In the stage of **"stop-and-go" movement**, the particle is for a part of the time trapped between other particles on the bed. This stage follows initiation of motion and is especially typical of particles smaller than the average bed material. At this stage a very small increase in u_* above the critical value produces a marked increase in the average rate of particle movement, which however is controlled mainly by the bed and particle size.

(2) In the stage of **continuous movement in contact with the bed**, the particle rolls or skims over the surface of the bed.

(3) In the highest stage, particles begin to be **lifted up above the bed level or to make long jumps**. Increases in particle velocity are roughly proportional to increases in shear velocity.

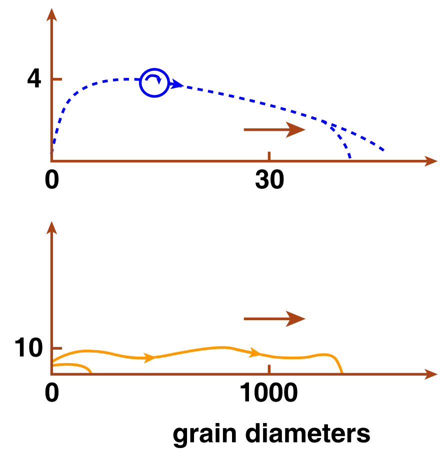

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Figure 10-5. Sketches of typical particle trajectories observed by Francis (1973). Above: saltation trajectory. Below: suspension trajectory (wavy line, with saltation trajectory shown to same scale for comparison. (From Francis, 1973.)

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31 Francis (1973) also distinguished three modes of movement, although they do not correspond exactly with the stages described by Meland and Norrman:

(1) **rolling** of particles in contact with the bed (roughly equivalent to stages 1 and 2 of Meland and Norrman);

(2) **saltation**, with particles rising up to heights of about two to four particle diameters above the bed and then falling back along "ballistic" paths, as illustrated in Figure 10-5; and

(3) **suspension**, in which at first particles move in "leaps" that are distinguished from saltation by their length and sinuous trajectories (Figure 10-5), but as suspension becomes better developed the particles rise farther above the bed and return to it less often.

32 Francis (1973) distinguished saltation from suspension on the basis of a qualitative assessment of the particle trajectory. Abbott and Francis (1977, p. 229) suggested a more rigorous definition: a particle is in saltation when it "jumps away from the bed and follows such a trajectory that its vertical acceleration is always directed downwards between the upward impulses sustained while in contact with the bed." If at any time the vertical acceleration is directed upwards, then the particle is regarded as being in suspension. According to this definition, whether or not a particle is in suspension cannot be determined simply from qualitative observation: a detailed analysis of the vertical component of its acceleration, based on high-speed photography of its trajectory, is needed. Further discussion of the nature of saltation (which is better developed in air than in water) will be deferred until we have examined some of the experimental results on the relation between flow intensity and rate of sediment movement. The mode of movement at shear velocities just above the threshold value u_{*c}

predicted by the Shields diagram was investigated by Abbott and Francis (1977), with the results shown in Figure 10-6.

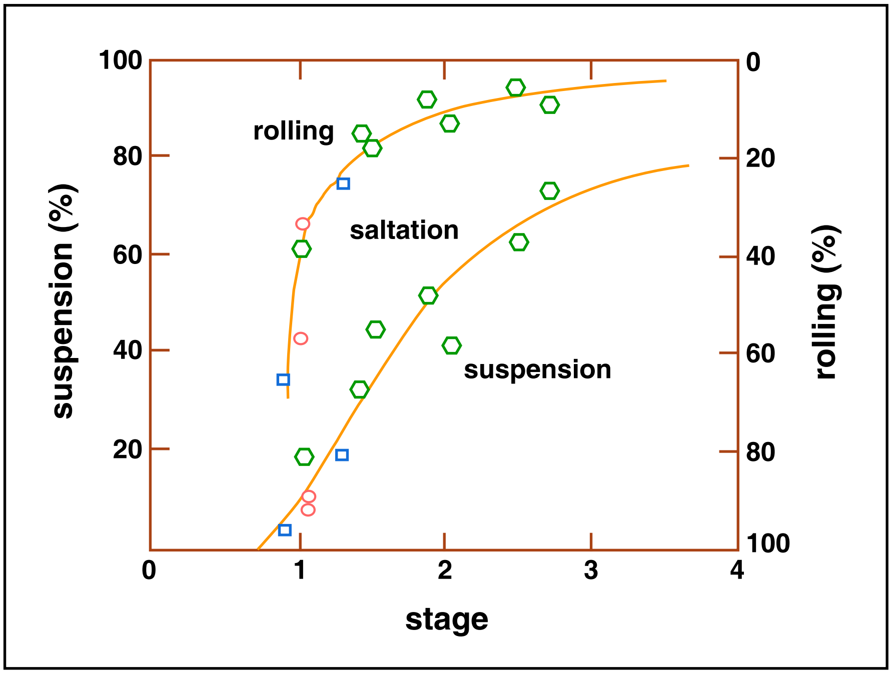

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Figure 10-6. Trajectories of a saltating glass sphere calculated for the case of drag only (non-rotating sphere) and drag plus lift (a sphere rotating 275 times per second) compared with the observed trajectory. It can be seen that spinning produced a large effect both on the shape of the trajectory and on the maximum height of rise. (From White and Schultz, 1977.)

33 The earliest particle movements, at u_*/u_{*c} of about 1 (corresponding to u_*/w about 0.15), are by rolling, but as the flow intensity increases, saltation rapidly becomes the dominant mode of particle movement. By the time u_*/w has reached values of only about 0.3, about 50% of the particle trajectories are classified as being in the suspension mode, using the strict definition of Abbott and Francis (1977), but the particles follow paths that are still very close to the bed, and the average speed of the particles, U_G, is roughly proportional to the shear velocity. (U_G/u_* is about 6 to 8, indicating that the speed of the particle is almost equal to that of the flow close to the bed) At $u_*/w = 0.5$ most trajectories would be classified by Abbott and Francis (1977) as in the suspension mode, but the particles are still moving mainly close to the bed in a mode that might be subjectively described by most observers as being more like saltation than true suspension.

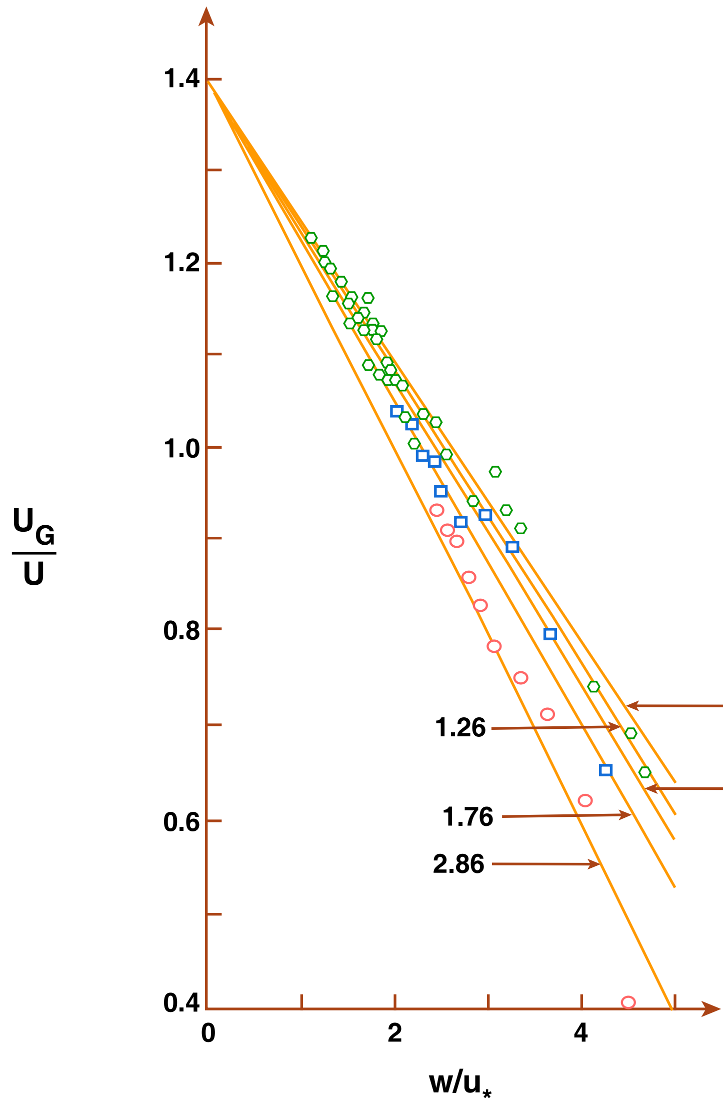

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Figure 10-7. Plot of dimensionless particle speed U_G/U vs. the ratio of particle settling velocity to shear velocity, w/u_*. Different symbols and lines refer to experimental particles of different specific gravities. (From Abbott and Francis, 1977.)

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34 As flow intensities are increased further, particle trajectories become longer and more irregular and the particles are carried higher into the flow. At these higher intensities it seems more reasonable to normalize the average speed of particle movement by dividing by mean flow velocity U (a property of the main flow) than by dividing by shear velocity u_* (a property of the flow just above the bed). Abbott and Francis (1977, and see Francis, 1973) found that for particles of equal density U_G/U was directly related to w/u_*, the reciprocal of u_*/w (Figure 10-7). At u_*/w greater than 0.5 the average particle speed was actually higher than the mean flow velocity, because most trajectories carried the particles up into the higher and therefore more rapidly moving parts of the flow. (Earlier experiments reported by Francis, 1973, suggest that in most cases U_G/U does not approach 1 until u_*/w approaches 1.) This is consistent with the suspension criterion, $u_* > w$, introduced in an earlier section.

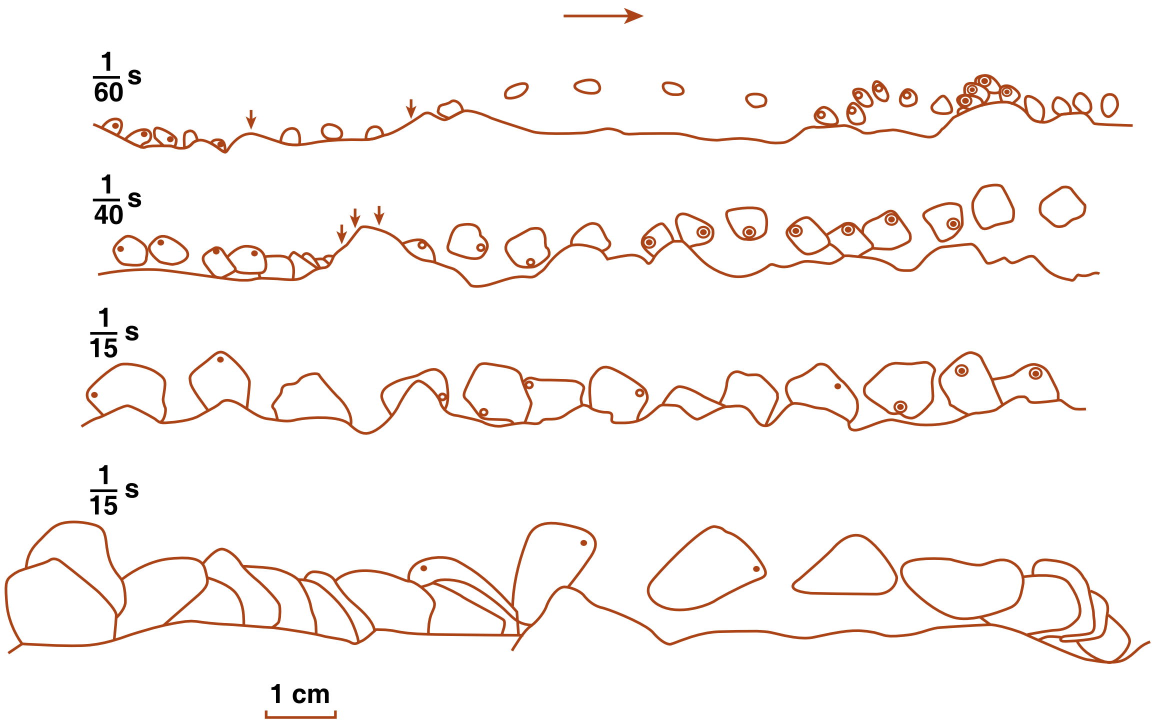

Figure by MIT OpenCourseWare.

Figure 10-8. Typical bed-load trajectories of four particles of differing sizes, traced from side-view high-speed motion pictures in a stream with median bed-material size of 4 mm. The filming speed, in frames per second, is given for each trajectory. (From Drake et al., 1988.)

35 In his pioneer stochastic model for particle movement, Einstein (1950) postulated that the average distance traveled by a particle moving as bed load does not depend upon the flow intensity. Fernandez Luque (1974; see also Fernandez Luque and van Beek, 1976), in observations of particles moving over a loose planar bed at shear stresses only slightly larger than critical, found from direct observation that the average length of individual particle "steps" (or saltation jumps) was a constant equal to 16 times the particle diameter. Particles accelerated slightly at the beginning of each jump and decelerated upon returning to the bed but generally did not come to rest. On the average a particle jumped about 18 times, for a total step length of 288 particle diameters, before coming to rest. The average velocity while moving was reduced by collisions with the bed surface to about 85% of the maximum velocity achieved in each jump.

36 There have been few observational studies of the motions of bed-load particles in flows over loose rather than immobilized, sediment beds. Drake et al. (1988), in a study that shows how much information about particle motions can be obtained by carefully arranged observation, recorded the movements of bed-load particles on the bed of a clear-water stream by means of high-speed cinematography. The stream was 6.45 m wide and 0.35 m deep, and the bed consisted of moderately sorted sand and gravel with a median size of 4 mm. During filming, the bed shear stress was about 6 Pa, which was about twice the threshold for movement. There was active bed-load movement, but the concentration of bed load was not so great as to obscure the motions of the particles. The particles moved mainly by rolling, although the finest moved by saltation, and large, angular particles moved by brief sliding, pushing smaller particles out of the way. Displacement times for individual particles, which lasted for a few tenths of a second, were much shorter than repose times. Figure 10-8 shows representative trajectories of four particles, of various sizes, as seen in side view, taken from the motion-picture frames.

Effect of Shape

37 The movement of particles on the bed is strongly affected by their shape. Particles tend to become oriented on the bed by pivoting around other particles or resting against them, and they do not necessarily orient themselves with their maximum projection area normal to the flow, as they generally do during settling. Certain shapes—notably prolate forms (rollers) but also disks—roll more easily than others.

38 An early experimental study of the effect of shape was made by Krumbein (1942) using artificial ellipsoidal particles, all with the same nominal diameter, in a flume with a smooth bed. Depth was held constant at 0.3 m and velocity was varied by changing the slope. Krumbein found that, for a given fluid velocity, spheres and rollers moved fastest. Within any one shape class (e.g., rollers), particles velocity increased with increasing sphericity; the shape effect was greatest at low fluid velocities and particle velocities, and was less important as particles tended to be taken into suspension.

39 Lane and Carlson (1954) found that pebbles lining the beds of Colorado drainage canals were sorted by both size and shape. In a given sample of bed pebbles the disk-shaped pebbles had substantially smaller volumes than the more spherical pebbles—the opposite of what would be the case if the pebbles had the same settling velocity—indicating that spherical pebbles rolled more easily and were more easily set in motion than disk-shaped pebbles, which tended to assume more stable, imbricated orientations on the bed.

40 Bradley et al. (1972) studied the effect of shape both in the field (Knik River, Alaska) and in the laboratory. They detected downstream sorting of shapes, with platy pebbles being the most easily transported, then elongate pebbles (rollers), and more equant pebbles being the least easily transported. They recognized that the anomalies in the shape effects observed in different field and laboratory investigations are probably caused by the different shape-sorting effects of particles moving by traction and by suspension. The readier transport of bladed pebbles can probably be explained by their observed “erratic saltation” type of motion, which tends to lift the bladed particles up into the flow, so that at sufficiently high fluid velocities their low settling velocity is more important than their poor rollability.

Bed-Load Movement at High Flow Intensities

41 As the flow intensity increases, and bed load becomes more abundant, the bed-load layer becomes thicker and the separation distance between bed-load particles becomes smaller. The difficulties of observing the details of particle motions in such thick, high-concentration bed-load layers become formidable—one might even say insurmountable. It seems fair to say that the ratio of hard observational data to theoretical deduction is probably lower in this area of sediment transport than in any other. The literature is replete with speculation about the forces and motions involved.

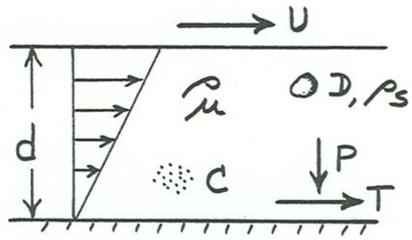

Figure 10-9. Variables governing dispersive stresses in a sheared sediment–fluid mixture, in the absence of gravity.

42 We start with the classic work of Bagnold (1954, 1956), which has played such an important role in subsequent thinking. Bagnold made pioneering experiments on interparticle forces in a strongly sheared mixture of water and solid particles. The experiments were made in a small, table-top apparatus that consisted of two concentric cylinders, with a thin annular space between. The inner cylinder was held stationary and the outer cylinder was rotated, giving almost uniform shear in the annular space, much like the hypothetical kitchen-table experiment described in Chapter 1. The annular space was filled with water containing a certain concentration of neutrally buoyant solid spherical particles. For a range of particle concentrations and rotation rates, Bagnold measured both the shear stress and the normal stress on the wall of the inner cylinder. He observed that both the normal stress and the shear stress were in excess of those for zero particle concentration, and he attributed these stresses, which he called dispersive stresses, to the intuitively reasonable effect of lateral forces engendered by particle interactions in the sheared mixture. Such interactions might be actual ballistic collisions, albeit cushioned to a greater or lesser degree by the ambient fluid, or they might only involve lateral particle motions caused by distortions of the local flow field by the presence of nearby particles moving at different speeds in the sheared medium.

Image removed due to copyright restrictions.
Bagnold, R. A. "Experiments on a Gravity-free Dispersion of Large Solid Spheres in a Newtonian Fluid Under Shear." Royal Society [London], Proceedings, vol. A225, 1954, pp. 49-63.

Figure 10-10. Experimental results on dispersive normal stress and dispersive shear stress in experiments by Bagnold (1954). σ is the fluid density, and λ is the "linear concentration", a measure of sediment concentration, the ratio of sediment particle diameter to mean free distance between particles.

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43 It is easy to develop a dimensionless framework in which to evaluate the results of Bagnold's experiments. Imagine that you are weightless, high above the Earth in a space station. You are equipped, somehow, to do the experiment described in Chapter 1, shearing a fluid between parallel plates, but without having to worry about leakage of fluid around the edges. You are at liberty to use particles of any density, because you do not have to be concerned that the particles will settle under their own weight or be centrifuged in a rotating device.

44 Which variables would govern the dispersive normal stress and dispersive shear stress of the kind that Bagnold observed in his experiments (Figure 10-9)? Plate spacing L and relative plate velocity V are not important by themselves but only in combination to give the shear rate V/L; call that R. The others are straightforward: density and viscosity of the fluid, and density, size, and concentration of the particles. You can nondimensionalize the stresses in a way similar to a particle Reynolds number: $(T/\rho)^{1/2}D/\mu$, and $(P/\rho)^{1/2}D/\mu$. One obvious independent dimensionless variable is ρ_s/ρ, the density ratio, and another is the concentration C itself, if it is taken to be a volume concentration. The third dimensionless variable needs to involve the shear rate; it is most natural to construct a variable in the form of a Reynolds number, $\rho RD^2/\mu$.

45 Bagnold's experiments were more restricted than your space-station experiment, because to avoid centrifugation he had to use neutrally buoyant particles. The implication of that is that the dispersive effects he found would have been even greater if ρ_s/ρ could have been greater, as with natural sediment in water. Bagnold's experimental results, cast in the dimensionless form developed above, are given in Figure 10-10.

Figure 10-11. The "gravity-bed case": a turbulent shear flow in a gravity field, transporting sediment particles denser than the fluid medium.

46 The dispersive stresses Bagnold observed are now known to be important in a wide range of what are called ***grain flows***: flows of loose solid particles caused by the direct force of gravity, without the necessary involvement of a fluid medium. Grain flows are important in certain natural environments, as in snow avalanches, certain kinds of landslides, and, on small scales, sand flows down the lee faces of eolian sand dunes, and in technology as well.

47 Bagnold (1956) took the further step of applying the concept of dispersive stresses to what he called the "gravity-bed case": a flow of fluid in a channel or conduit in a gravitational field, transporting denser particles near the bed, as what we would call, in the context of this chapter, bed load (Figure 10-11). The idea is that if the flow is strong enough there can be a lowermost layer of transported sediment, with a thickness of many particle diameters, in which the shear is sufficiently strong that a dispersive normal stress makes its appearance and acts to maintain the bed-load layer in a dispersed state. Bagnold theorized that within this sheared and dispersed bed-load layer the fluid turbulence is unimportant, in the sense that it is not the principal agent that maintains the particles in the dispersed state. These ideas were later elaborated by Dzulynski and Sanders (1962), who applied the term ***traction carpet*** (which is in common use to this day; see, for example, Hiscott, 1994, 1995, and Sohn, 1995) to the concept, and by Moss (1972), who introduced the term ***rheological layer*** for essentially the same concept. A quotation from Moss (1972, p. 162) captures the essence of the phenomenon well:

> As bed-load motion becomes more intense in sand-sized materials, a stage is reached wherein collisions between particles become inevitable and thereafter the load proceeds as a dense mass of colliding particles, buoyed up by the dispersive pressure thus generated.... This moving mass of particles behaves like a viscous fluid, but has a remarkably sharp boundary with the flow above and maintains almost constant thickness over quite large bed areas.... (It) will be called the "rheological layer."

48 The problem is that although the concept of a dispersion layer is consistent with the well-established importance of dispersive stresses in certain ranges of shearing of particle–fluid mixtures, no one has ever seen inside one, owing to the obvious experimental difficulties. (One investigator—the writer of these notes!—once tried to overcome the observational difficulties by means of the seemingly ingenious technique of using monochromatic illumination of transparent sediment particles being transported in a concentrated bed-load layer by a transparent liquid with exactly the same index of refraction as the particles, in order to have a clear and unobstructed view of a few opaque fluid and sediment marker particles and record details of particle motions in the interior of the bed-load layer using high-speed cinematography. He could never get it to work well enough, though.) Until the importance of dispersive stress in concentrated bed-load layers is established by observation, rather than merely deduced, the concept is best regarded as hypothetical rather than as proven. Of course, traction carpets or rheological layers can still exist; it is just a matter of whether dispersive stresses or other effects like small-scale fluid turbulence are the more important factor in their dynamics.

Saltation in Water

49 In Chapter 11 you will learn that in the wind, saltation is the principal, and very characteristic, mode of particle movement. In saltation, particles take long, arching trajectories above the bed, little influenced by the turbulence in the flow. Here we address the question: What is the nature and relative importance of saltation in water? The importance of saltation in air is clear, but there is much less agreement on its importance in water. Saltating particle rise much higher above the bed in air (commonly a large fraction of a meter) than in water (only a few millimeters) because of the much greater effect of fluid drag and the reduced effect of particle inertia in water. Kalinske (1943) calculated that the height to which saltating particles would rise, for given particle size and shear velocity, should be inversely proportional to the fluid density, i.e., particles should rise 800 times higher in air than in water. Also, the criterion for suspension developed in a previous section, $u_* = w$, implies that particles should be relatively easily taken into suspension in water, because of the much lower settling velocity of particles in water than in air. Therefore most engineering writers (Einstein, 1950; Einstein and Chien, 1955; Ippen, 1971; Vanoni, 1975) have assumed that suspension by turbulence is a much more important mechanism of sediment transport in rivers than saltation, even quite close to the bed. In contrast to this view, Bagnold (1956, 1973) has argued that true saltation is independent of turbulence, and that high concentrations of particles close to the bed tend to suppress turbulence and make saltation (and particle collisions) the dominant mechanism of sediment transport.

50 Certain observations by Gordon et al. (1972), Fernandez Luque (1974), Fernandez Luque and van Beek (1976), and Abbott and Francis (1977) suggest that simple ballistic movement of particles, and movement by particles impacting on the bottom, may not be as important in water as some authors have held. Gordon et al. (1972) studied the saltation of spheres of diameter 6.6 mm and specific gravity 1.3 in a flow of water. Particle movement was made essentially two-dimensional by confining the flow within a flume only 7.9 mm wide. Observed trajectories were typical of saltation except that take-off angles were rather low, generally in the range of 10° to 35°. One reason for the low lift-off angle was that a saltating particle did not simply bounce off the loose particles on the bed; instead the moving particle rolled around the particle on the bed before lifting off to make another saltatory jump. There was a clear correlation between the fractional loss of kinetic energy and the angle of incidence in the collision, but the collisions were not simple elastic collisions; it seems clear that a combination of particle inertia and fluid drag forces was involved. Both Fernandez Luque (1974) and Abbott and Francis (1977), studying saltation in water, found that very few apparent saltations could be explained entirely as simple ballistic trajectories; some other kind of lift force was involved in most trajectories. These authors did not investigate Magnus effects, but it seems probable (particularly for the data reported by Francis) that the main lift was provided by turbulence. Abbott and Francis (1977, p. 253) found that “there appears to be an effective elastic rebound between the bed and a moving grain impinging on it.” Very few observed saltations immediately followed the return of a particle to the bed; most were preceded by some rolling. Furthermore, there seemed to be no difference in take-off velocity between particles rebounding from the bed and particles beginning a saltation from rest or rolling.

51 Murphy and Hooshiari (1982) studied the saltation of marbles 15.7 mm in diameter on a bed of similar but fixed marbles. The settling velocity, about 0.8 m/s, was much higher than the shear velocities needed to produce continuous saltation (0.08–0.11 m/s), so there is no doubt that saltation rather than suspension was the dominant mode of movement. In this case, particles appeared to be rebounding directly from the bed, though the exact mechanism of initial rise from the bed could not be studied by the stroboscopic technique used. Analysis of the trajectories indicated that they could be satisfactorily accounted for by a model that took into account gravity (and buoyancy), horizontal and vertical components of drag, and the added-mass effect that is produced by accelerating a solid through a fluid (Hamilton and Courtney, 1977). Magnus effects were not significant for particles of this size and shape in water. The observations suggest that bed impact forces are sufficient to produce the upward rise, and that lift forces are not necessary. If this is true (and it is not proven, because the bed was rigid in the model, not loose as it would be in nature) then there is the possibility that saltation of larger (gravel-size) particles in water may be different from that of sand in water. It may be that the saltation of gravel in water is more like that of sand in air than that of sand in water.

52 All of the observations described above were made on flows in which very few particles were in saltation. Possibly fluid drag and lift play a much reduced role in initiating and maintaining saltation in a "traction carpet", but the observations certainly suggest severe limitations on a simple impact hypothesis for saltation at low concentrations in water, and they indicate a very significant role for turbulence in transport of sand as bed load.

SUSPENSION IN A SHEAR FLOW: THE DIFFUSIONAL THEORY OF SUSPENSION

53 Suspended particles are held up above the bed by the turbulent motion of the fluid. The weight of the particle is transmitted directly to the fluid, by way of the drag force exerted by the particles as they fall through the surrounding fluid, and increases the hydrostatic fluid pressure at the bed, in much the same way an airplane in flight increases the atmospheric pressure in an ill-defined circular area on the ground below. Suspended particles thus exert a force on the bed, albeit indirectly, in contrast to the direct forces exerted on the bed by moving bed-load particles.

54 It is theoretically possible for particles to move through the fluid close to the bed without actually being in contact with the bed, and yet not be in suspension. This happens in true saltation: the ballistic motion of the particles results from fluid lift forces and/or particles striking the bed, but it is not at all dependent on turbulence—and in fact Francis (1973) has described saltation of particles in a laminar flow. It has also been postulated that particles may be held in a dispersed state close to the bed by actual collisions between particles or by near-misses that produce viscous forces with vertical components that hold the particles above the bed. This is the "dispersive pressure" of Bagnold (1956), the effectiveness of which is still a matter for debate.

55 It was noted earlier in this chapter that particles first begin to travel in suspension when the vertical component of turbulence (or, more precisely, the normal-to-the-bed component of the turbulence) becomes about equal to the settling velocity of the particles (Equation 10.2). As noted earlier, there is no natural way to characterize the magnitude of this fluctuating component of the vertical fluid velocity, because it fluctuates over a wide range of values; the rms value is usually used to characterize its magnitude for this purpose. Contrary to a view that has sometimes been expressed in the literature, suspension does not depend on asymmetry in the frequency distribution of the vertical fluctuating velocities: provided that at least some of the vertical fluctuations are greater than the settling velocities of the particles, some of the particles experience suspension, even if the frequency distribution of fluctuating velocities is asymmetrical, because the conditions would still be conducive to diffusion (see Chapter 1): random motions of the medium, in combination with an upward gradient in sediment concentration, from nonzero in the bed-load layer to some smaller value, perhaps even zero, at some greater height above the bed. Such an asymmetry in the frequency distribution of vertical velocities might, however, affect the details of the concentration distribution.

56 Before dealing with the more important but more complicated case of suspension in a turbulent shear flow, we will look at suspension by homogeneous and isotropic turbulence. The characteristics of the turbulence do not vary from place to place within a certain region of the fluid, and neither do they vary with direction at any point within that region. Rouse (1939), the first to study sediment suspension in this way, produced a close approximation to isotropic turbulence by vertically oscillating an array of square grids in a large-diameter vertical cylinder ("turbulence jar").

57 The downward volume flux of particles by settling, from a region in the fluid having a concentration C of uniform-size particles, is $-wC$. It is reasonable to assume that the upward vertical diffusion of particles follows a Fickian diffusion law, like many other diffusion processes (see Chapter 1), so that the upward volume flux of particles by diffusion is $\varepsilon_S dC/dy$, where ε_S is a diffusion coefficient, which should be constant in a field of isotropic turbulence of any particular type and strength, and the positive y direction is upward. Equating the two fluxes gives an expression for the vertical distribution of the concentration of suspended particles:

$$wC + \varepsilon_S \frac{dC}{dy} = 0 \tag{10.8}$$

The resulting expression for suspended sediment concentration as a function of height y above the bed, developed below, is sometimes called the diffusional theory of suspended-sediment concentration. It also seems reasonable that the diffusion coefficient ε_S is proportional to, if not actually equal to, the corresponding coefficient for the diffusion of fluid momentum, i.e., the kinematic eddy viscosity, and therefore in a turbulence jar it should be proportional to the frequency of vertical oscillation of the grid. Rouse verified that this is the case, thus confirming the validity of the diffusion equation (see also experimental results reported by Antsyferov and Kos'yan, 1980).

58 In nature, homogeneous and isotropic turbulence is the exception; we have to deal with turbulence that typically varies in its characteristics with distance from the boundary, and at least to some extent with direction, mainly normal to the boundary at any point. In a turbulent shear flow, as for example, in a river, a tidal current, or a strong wind, where turbulence is not even approximately homogeneous and isotropic except perhaps at large distances from the bed, we should expect that the diffusion coefficient varies in the direction y normal to the bed, so we need an expression that tells us how it varies with y before we can make use of Equation 10.8 to predict how the sediment concentration varies with y.

59 To find such an expression we assume the sediment diffusion coefficient ε_s to be proportional to the eddy viscosity ε, given by

$$\tau = \rho\varepsilon \frac{du}{dy} \tag{10.9}$$

Assuming that $\varepsilon_s = \beta\varepsilon$, then

$$\tau = \frac{\varepsilon_s \rho}{\beta} \frac{du}{dy} \tag{10.10}$$

where β is a coefficient that is likely to be close to one. You already know that τ varies linearly with y in uniform open channel flow,

$$\tau = \tau_o \left(1 - \frac{y}{d}\right) \tag{10.11}$$

(see Chapter 4), so

$$\varepsilon_s = \frac{\frac{\beta\tau_o}{\rho}\left(1 - \frac{y}{d}\right)}{\frac{du}{dy}}$$

$$= \frac{\beta u_*^2 \left(1 - \frac{y}{d}\right)}{\frac{du}{dy}} \tag{10.12}$$

Using the law of the wall in differential form (Chapter 4),

$$\frac{du}{dy} = \frac{u_*}{\kappa y} \tag{10.13}$$

we have

$$\varepsilon_s = \beta u_* \left(1 - \frac{y}{d}\right) \kappa y \tag{10.14}$$

Equation 10.14 is the relationship between ε_s and y that we need in order to solve Equation 10.10. Combining Equations 10.8 and 10.14 gives

$$\frac{dC}{C} = \frac{-w\,dy}{\beta \kappa u_*(1-\frac{y}{d})y} \tag{10.15}$$

which can be integrated to give the equation first derived by Rouse (1937):

$$\ln C = \frac{w}{\beta \kappa u_*} \int_a^d \frac{dy}{(1-\frac{y}{d})y} \tag{10.16}$$

or

$$\frac{C}{C_a} = \left(\frac{d-y}{y} \frac{a}{d-a} \right)^z \tag{10.17}$$

where

$$z = \frac{w}{\beta \kappa u_*} \tag{10.18}$$

The exponent z is sometimes called the ***Rouse number***.

60 You can see from Equations 10-17 and 10-18 that the larger the value of z, the more rapidly the suspended-sediment concentration decreases with height above the reference level a. Equation 10.17, graphed in Figure 10-12, gives the concentration of suspended sediment of a given settling velocity w at a height y above the bed relative to its concentration C_a at an arbitrarily chosen "reference level" $y = a$ above the bed.

61 Ideally, the reference concentration C_a would be taken to be as close to the bed as possible but still far enough above the bed that a balance between downward settling and upward turbulent diffusion of suspended sediment is physically reasonable. The theory fails very close to the bed, because a balance between passive upward turbulent diffusion and downward settling is not applicable there: particle movements at and very near the bed are controlled by fluid lift and drag forces, and if concentrations are high these movements may be significantly affected by collisions or interactions between particles. The reference height a above the bed is most naturally just above the bed-load layer. This is consistent with the idea that the sediment concentration at the top of the bed-load layer acts as a lower boundary condition for the distribution of suspended sediment higher in the flow. This points up the problem of specifying the suspended-sediment concentration in absolute rather than relative terms: no successful theory has been developed yet for the bed-load concentration as a function of flow and sediment conditions. Because the structure of the flow and the dynamics of bed-load movement are so complex in the near-bed layer when the flow is strong enough to move sediment in suspension, no elegant way has been developed to put this appealing idea, that the bed load forms the lower boundary condition for the suspended load, into useful practice.

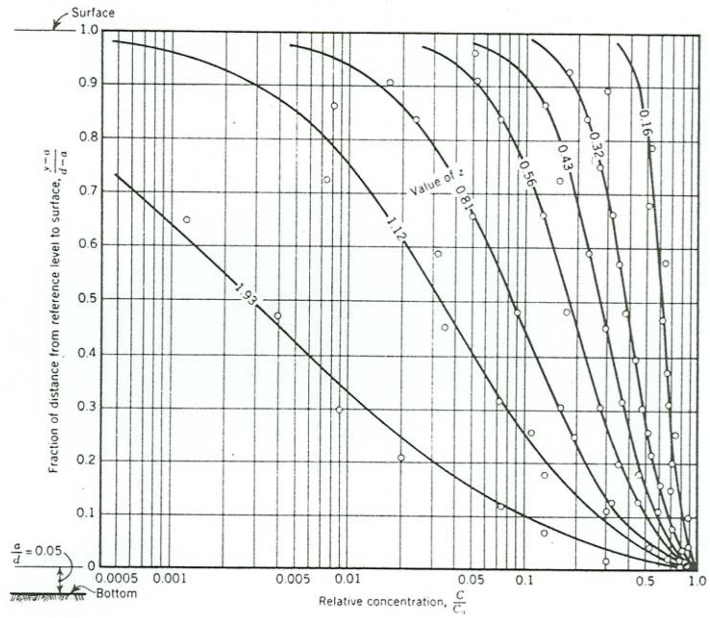

Figure 10-12. Distribution of the relative concentrations of suspended sediment with relative depth above the datum y = 0.011-3711-485d. (From Vanoni, 1975.)

62 Experiments to test Equation 10-19 were reported by Vanoni (1946). These experiments were mostly made at relatively high velocities over a planar bed (either a sand bed or the rigid floor of the flume) and at varying concentrations of sand. Vanoni found a general agreement between predicted and observed sediment concentrations (Figure 10-12).

63 Because both β and κ are supposed to be constants, the main factor that determines the distribution of suspended sediment with height y above the bed should be the ratio of the settling velocity w to the shear velocity u_*. It was suggested in an earlier section of this chapter that a critical ratio of about one determines whether any particles will go into suspension: since $\beta \approx 1$ and $\kappa \approx 0.4$, w/u_* less than one corresponds to z less than 2.5. We can see from Figure 10-12 that at values of z greater than 2.5 any sediment in suspension would be concentrated in a zone very close to the bed—and this tends to confirm the choice of w/u_* as a suitable criterion for suspension.

64 Two factors in open-channel flows have a direct effect on the ratio w/u_* and therefore on the vertical profile of suspended-sediment concentration: viscosity, and friction. First the viscosity: for a given particle size and shape, w is reduced by an increase in the viscosity sensed by the settling particles. That can be brought about in two ways: a reduction in the temperature of the fluid, which increases the viscosity of the fluid itself, or an increase in the wash-load concentration. In the latter case, the viscosity of the fluid remains the same but the effective viscosity of the deformable medium (the fluid charged with wash load) that is sensed by the particles of the suspended bed-material load, which are much larger than the particles of the wash load, is greater. Both of these effects act to reduce the ratio w/u_*, and hence the Rouse number z, and make sediment more uniformly distributed in the vertical (Equation 10-19). The fluid-viscosity effect diminishes with increasing settling-velocity Reynolds number, however, and becomes unimportant when the range of Reynolds numbers for which the drag coefficient is approximately constant is reached; see Chapter 2.

65 Now for the effect of friction: for a given mean flow velocity, an increase in the coefficient of bottom friction causes an increase in bottom shear stress, and therefore in shear velocity. To see why, go back to the definition of the friction factor f (Equation 4.18 in Chapter 4): $\tau_o = (f/8)\rho U^2$, or $U/u_* = (8/f)^{1/2}$. So an increase in the shear velocity also results in a more uniform vertical distribution of suspended sediment, by decreasing the ratio w/u_*. In sand-bed rivers, changes in f are produced mainly by changes in the relative roughness, which depends mainly on the nature and size of the bed forms. Large bed forms, like dunes, produce large values of f, and therefore cause suspended bed-material sediment to be distributed more uniformly in the vertical than if the bed were planar. In fact, it is been observed in flume studies that the vertically averaged suspended-sediment concentration actually decreases somewhat in the transition from a dune-covered sand bed to an upper-regime plane bed, with its accompanying decrease in flow resistance, as the flow velocity increases.

66 The theory of suspension by turbulent flows outlined above is based on the assumption that the flow is steady. This is be a reasonable approximation for most rivers, but tidal currents change quite rapidly in both depth and speed over the tidal cycle. It has been shown that in experimental turbulent shear flows, decelerating flows have larger turbulence intensities, and produce larger shear stresses on the bed, than steady flows. Decelerating flows therefore might be expected to be more erosive, and to have a higher capacity for suspended sediment, than steady or accelerating flows. Wimbush and Munk (1970), Gordon and Dohne (1973), Gordon (1975), Bohlen (1977), and McCave (1979) have reported measurements suggesting that turbulence intensities are higher than normal during deceleration of flows on both flood and ebb tides. Gordon (1975) and Bohlen (1977) have commented on the implications for transport of suspended sediment by tidal currents, but convincing direct evidence of the effect of deceleration on sediment transport by tidal currents is still lacking.

67 The diffusional theory of suspension presented above is based on the assumption that turbulence diffuses sediment according to a simple ("Fickian") diffusion law. This assumption is in reasonably good accord with experiment, but it is not the only possible basis for a theory of sediment suspension. Alternative theories, based on different assumptions, are described by Nordin and McQuivey 1971), Drew (1975; see also Drew and Kogelman, 1975), Willis (1979), Herczynski and Pienkovska (1980), and McTigue (1981), among others.

68 Although the diffusional theory of sediment suspension has been described as "the brightest analytical achievement to date in the field of river hydraulics" (Hsu et al., 1980; see also Kennedy, 1984, p. 1257), in that it represents an elegant and rational theoretical approach, based on reasonably well understood physical effects, that does quite well in its predictions without relying upon any suspicious "fudge factors", it is subject to a number of criticisms:

• The theory takes no account of the details of how the sediment particles are actually handled by the eddies in the turbulent flow field. There are two different aspects to this. One has to do with the interesting and counterintuitive effect of the tendency of eddies to trap sediment particles (Tooby et al. 1977; Nielsen, 1984), discussed briefly in Chapter 3. The other is that the theory assumes turbulence that is isotropic in its vertical motions, i.e., that the frequency distribution of the vertical velocity is symmetrical. There is good reason to believe, however, that close to the bed the vertical component is anisotropic (Leeder 1983a, 1983b): the less common upward motions are stronger than the more common downward motions in this region, as would be expected from the semicoherent burst–sweep structure of the near-bed turbulence (Chapter 4). As first proposed by Bagnold (1966), and further developed by Leeder (1983a, 1983b), the anisotropy in vertical turbulent velocities is what maintains sediment in suspension—with the implication that without this anisotropy the concentration of sediment in suspension would be much less. The flaw in this concept is that, to maintain balance of fluid masses passing upward and downward in the turbulence field, the downward-moving eddies must cover a greater area, in any plane through the flow that is parallel to the bottom boundary, than the upward-moving eddies, thus maintaining a balanced exchange of sediment even in the face of the vertical anisotropy of turbulence. Despite some assertions in the literature to the contrary, such anisotropy is only a minor distorting effect on the diffusional theory, and is not a necessary condition for the maintenance of bed-material sediment in suspension.

• Vanoni (1946, and many subsequent investigations reported and analyzed in Vanoni, 1975) found that in some experiments, particularly those in which there was a high concentration of coarse sediment close to the bed, the value of the supposedly universal von Kármán constant decreased from its accepted value of 0.38 to values as low as 0.2. He interpreted this as indicating that the presence of sand moving close to the boundary changed the structure of turbulence in the flow. The von Kármán constant κ plays a fundamental role in the diffusional theory of suspended sediment, by virtue of its effect on the gradient of time-average flow velocity in the law of the wall (Equation 10-15); if κ is itself affected nonnegligibly by the presence of suspended sediment, then it becomes part of the problem rather than an independent input to the problem, and the theory would become much more complicated.

• Besides the uncertainty about κ, several authors have reported large deviations of β from the expected value close to unity. There are reasons to expect that solid particles are not diffused at the same rate as fluid momentum, and that the ratio of the two rates of diffusion is not a constant but varies with the properties of both the sediment and the fluid turbulence. At present there is no satisfactory way to predict the value of β. Prediction presumably will become possible only when there is a better understanding of the mechanism of diffusion.

• In the usual theory the sediment diffusion coefficient is assumed to be proportional to the eddy viscosity and the distribution with depth to be given by Equation 10-18. This equation predicts that ε_S (and ε) drop slowly to zero as the free surface is approached. Because sediment cannot diffuse through the free surface, ε_S must be equal to zero there. Coleman (1970) has, however, calculated ε_S directly from observed values of C and dC/dy using Equation 10-8. He found that there is a strong dependence on depth only near the bed; over most of the flow, and even quite close to the free surface, ε_S appears to be independent of depth.

69 For all of these reasons, the diffusional theory of sediment suspension, though it is a better theory than that available for most aspects of sediment transport, must still be regarded as somewhat less than completely satisfactory.

A NOTE ON THE EFFECT OF ACCELERATION OF GRAVITY ON SEDIMENT MOVEMENT

70 It is worthwhile to consider how sediment movement and bed configurations in water flows might differ where the acceleration of gravity is different. Back in Chapter 8, in the section on dimensionless variables Paragraph 47), a set of dimensionless variables was developed in which the leading variables in a sediment-transport system,—variables with dimensions of length, like particle size, or variables with dimensions of velocity—can be organized in such a way that each of the leading variables in sequestered in its own dimensionless version. In each such variable, the acceleration of gravity enters as well. If gravity is different, any length or velocity variable in a dynamically similar system must then also be different. Southard and Boguchwal (1990) show that, in the case of Mars, for which the acceleration of gravity is only about 04 times that of Earth, a length variable on Mars would be about 1.36 tomes that on Earth, and a velocity variable on Mars would be about 0.74 times that on earth, for a dynamically similar system.

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