Chapter 2
PHYSICS OF SEDIMENTATION

1. A FLUID DYNAMICS FIELD TRIP
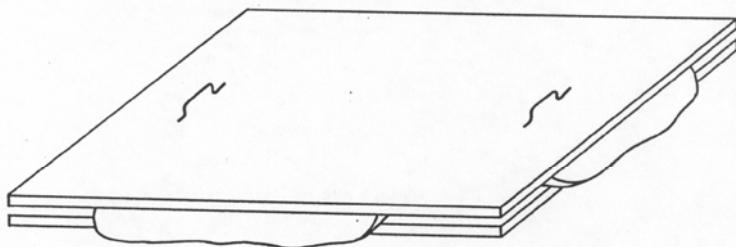
1.1 Introduction

1.1.1 In dealing with the production and significance of flow-produced sedimentary structures and textures, it helps to know something about the flows in which they were produced. On the other hand, we can't assume that you've had courses in fluid dynamics. So here's some relatively painless background on fluid dynamics.

1.1.2 What we've done here is imagine that we are out looking at the flow in a large river and provide some qualitative material on the most important aspects of the flow that you could observe for yourselves. That should go a long way toward providing important background without spending inordinate time.

1.2 Variables

1.2.1 Two of the most important physical properties of fluids are density and viscosity. The ***density*** of a fluid, denoted by ρ, is *the mass of fluid per unit volume of fluid*. The density of water is very close to one gram per cubic centimeter (1 g/cm^3) or one thousand kilograms per cubic meter (10^3 kg/m^3). The density of air at sea level and room temperature is far smaller, by a factor of about 800.

1.2.2 The ***viscosity*** of a fluid, usually denoted by μ, is *a measure of the resistance of the fluid to deformation*. Imagine that you sheared a fluid by placing it between two parallel plates and then moving the plates parallel to each other (Figure 2-1). The greater the relative speed of the plates, the greater the force needed to maintain the shearing of the fluid. The viscosity is *the coefficient associated with the ratio between the force (measured per unit area in the planes of shear) that resists the shearing, on the one hand, and the rate of shearing, on the other hand*.
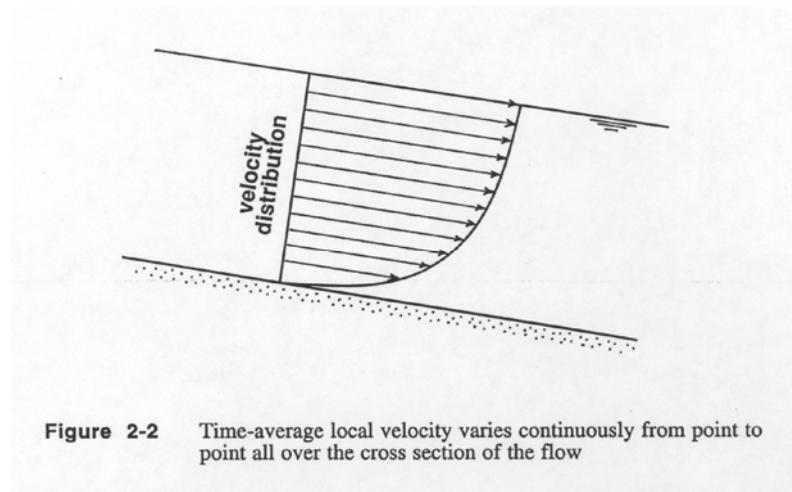

Figure 2-1   Shearing a viscous fluid between two parallel plates
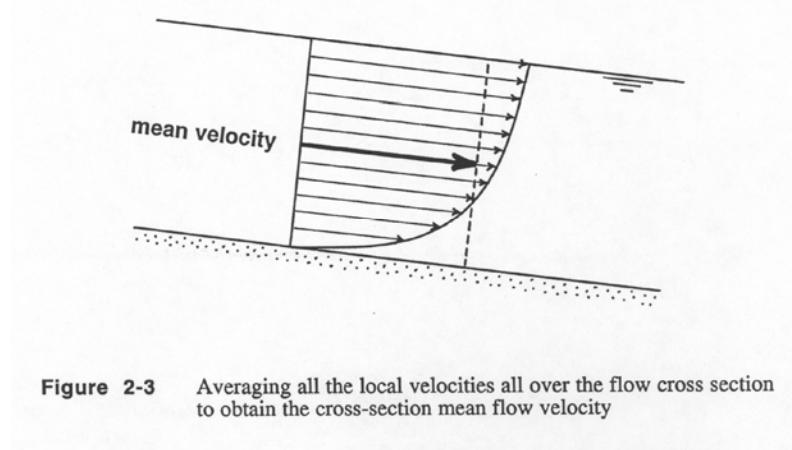
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1.2.3 One more fluid property, the specific weight, usually denoted by γ, is important when the fluid is in a gravitational field, as on the Earth's surface. The *specific weight* is *the weight of the fluid per unit volume of fluid*.

1.2.4 There's a wealth of variables you can use to describe the flow. Some are more concrete and easy to understand than others. On the one hand, you can think about the *velocity of the flow*, and on the other hand, you can think about the *force per unit area the flow exerts on its boundary*.

1.2.5 With regard to velocity, if you occupied any point in the flow that's fixed relative to the boundaries and used a magic little velocity meter to measure the local fluid velocity, you'd get some stable *local time-averaged velocity* if you continued your measurement for a fairly long time. This time-average local velocity varies continuously from point to point all over the cross section of the flow (Figure 2-2); more on what that distribution of local velocity looks like presently.
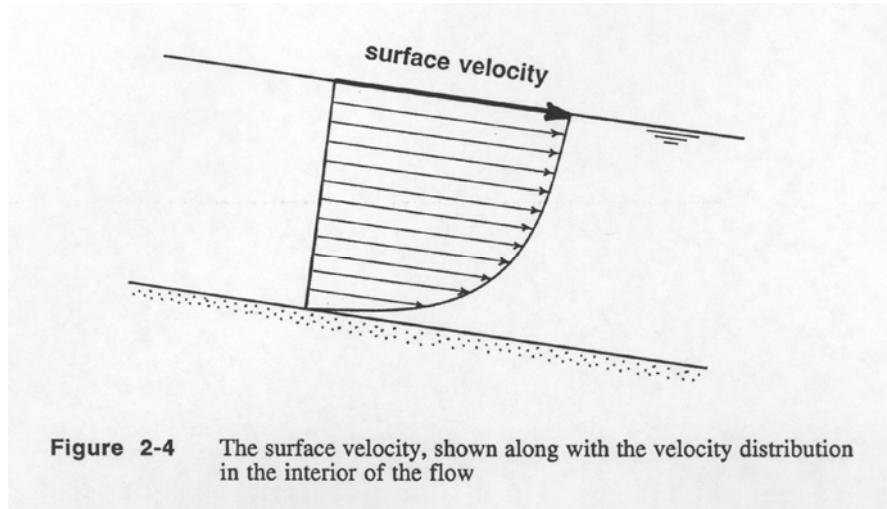

Figure 2-2   Time-average local velocity varies continuously from point to point all over the cross section of the flow

1.2.6 You can take the average of all the local velocities all over the flow cross section to get a *mean flow velocity U* (Figure 2-3). In real rivers this is how it's actually done, but in laboratory channels, where it's easy to know the water discharge (the volume rate of flow), you can use the relationship $Q = UA$, where Q is the discharge and A is the area of the flow cross section, to find U.

Figure 2-3   Averaging all the local velocities all over the flow cross section to obtain the cross-section mean flow velocity

1.2.7   You could also measure the *surface velocity U_S* along the centerline of the flow (Figure 2-4). This is greater than the mean velocity, by a factor of something like 1.2 or 1.3; it's related to U in a rather complex way that needn't concern us here. Incidentally, unless the channel is very wide compared with the depth of flow *the position of maximum velocity on the cross section lies a little below the surface*.

Figure 2-4   The surface velocity, shown along with the velocity distribution in the interior of the flow

1.2.8 The *boundary shear stress τ_O* is a little trickier to deal with than velocity, and is less concrete to visualize, but it's important for the obvious reason that it's what exerts forces on the sediment particles on the bed.

1.2.9 Let's backtrack a little to point out something about the flow near the boundary that may seem counterintuitive to you, but stands as a firmly established fact of observation: *fluid in direct contact with a solid boundary has exactly the same velocity as that solid boundary*. That's called the **no-slip condition**. In the context of our river flow, it means that the water right at the bed has zero velocity. But of course there's an increase in velocity upward from the bed, so all the water above the bed, no matter how small a distance above, generally has a nonzero velocity.

1.2.10 The reason we've brought in the no-slip condition here is this: *the origin of shear stresses in fluids is just the existence of shearing itself*. The stronger the shearing, the greater the shear stress that's generated, other things being equal. And foremost among those other things is the viscosity, introduced above: the viscosity is just the coefficient that relates the shear stress to the shear in a given fluid.

1.2.11 So, back to the boundary shear stress τ_O. It's true that if you could get right down close to any point on the bed, you could (with the right instrument, which doesn't actually exist!) measure a local shearing force per unit area the flow exerts on the solid boundary. That fits the term "boundary shear stress" well, but it's not what's conventionally meant by the boundary shear stress. What you have to do is *take a spatial average of the local downstream-directed component of the fluid force acting on the bed over an area large enough to include a large number of representative "roughness elements"* (sediment particles or bed forms).
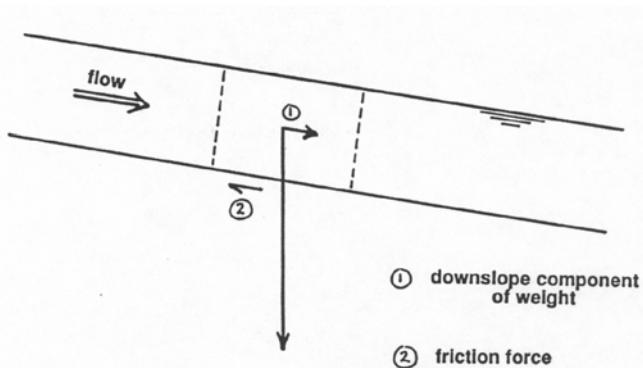
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1.2.12 Finally there's the matter of *flow power*. There's a school of thought (with which I don't agree) that the flow power is the most fundamental determinant of sediment transport. We'll just state to you, without further development, that in a channel flow like a river *the flow power comes out to be equal to the product of the boundary shear stress τ_O and the mean velocity U.* Those of you so inclined might try your hand at figuring that out from the first principles embodied in Newton's laws; it's not as formidable as it seems at first thought.

1.3 Why Do Fluids Move?

1.3.1 The movement of fluids is part of the experience of our daily lives. But it's important to think for a moment about *why* fluids move. According to Newton's first law, a body moves in a state of uniform motion (changing neither speed nor direction) unless it is acted upon by some force. Moving fluids are always acted upon by friction, so they come to rest unless some other force is available to offset the friction and keep them in motion.

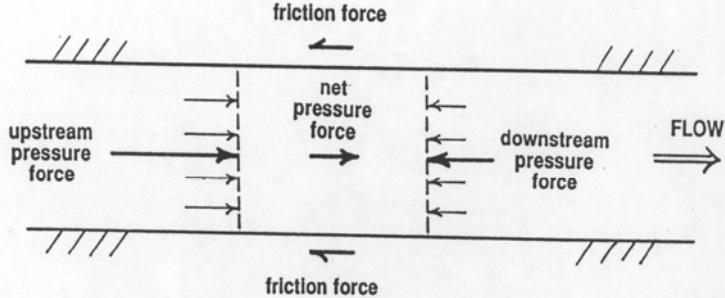**1.3.2** Two important kinds of forces drive fluid motions:

θ   *The force of gravity*. In a flow down a sloping channel, the weight of an element of fluid has a downslope component (Figure 2-5). This downslope component of fluid weight counterbalances the frictional resistance exerted on the fluid by the bottom boundary.

Figure 2-5   In a flow down a sloping channel, the weight of an element of fluid has a downslope component that's balanced by the friction force the boundary exerts on the flow

θ   *Pressure gradients*. In a flow in a horizontal pipe or conduit, the fluid moves only if there's a downstream decrease in fluid pressure. To see why this causes the fluid to move, look at an element of fluid in the pipe (Figure 2-6). If there's a downstream pressure gradient (that is, the pressure decreases in the downstream direction), then the force on the upstream face of the element is greater than the force on the downstream face. The difference in forces on the upstream and downstream faces is a force directed downstream. This downstream

force counterbalances the frictional resistance exerted on the fluid by the walls of the pipe.

Figure 2-6 In a flow in a horizontal pipe or conduit, the fluid moves only if there's a downstream decrease in fluid pressure

1.3.3 In a general fluid flow, *both of these kinds of forces (downslope gravity components and downstream pressure gradients) can be present at the same time, and the fluid moves under their combined effect.*

1.4 Channel Flows

1.4.1 Now let's think about what the flow really looks like in a large channel flow like a river. The basic picture is as shown in Figure 2-5: there's a balance between the downchannel pull of gravity on the fluid, which acts throughout the flow (in some physics course you may have heard gravity described as a "body force") and the resistance force the bed and banks exert on the flowing fluid, which is equal and opposite to the boundary shear stress we discussed above. A natural consequence of the action of these two different forces—gravity acting throughout the fluid and the resistance force acting only at the boundary—is that *the velocity increases upward in the flow, from zero at the boundary, by the no-slip condition, to a maximum at or near the surface.*

1.4.2 That's only the crudest first-order picture, though: the question comes naturally to mind: *what are the details of the distribution of time-average local flow velocity over the entire cross section of the flow?* That's much too intricate a question for us to pursue here, so we'll just hit the high points.

1.4.3 It's simplest to think about a river with a large width-to-depth ratio, which is usually the case. Then we can forget about the effect of the banks. A straightforward application of Newton's laws shows that the distribution of velocity should be *parabolic* (Figure 2-7A), with the vertex of one limb of a parabola located at the surface.
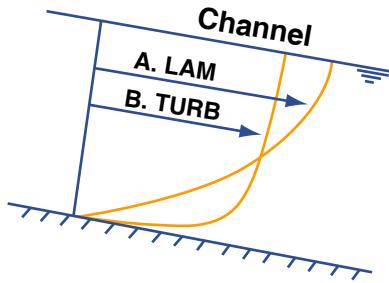
Figure by MIT OCW.

Figure 2-7: Laminar and turbulent velocity profiles in channel flow

1.4.4 But if you went out and measured the local time-average velocity at a series of points along a vertical from the surface down to the bottom, you'd find the actual distribution to be far from parabolic: *the distribution would be much more uniform over most of the flow depth, and the necessary decrease to zero at the boundary would be in a very thin zone right next to the boundary*. The distribution would closely approximate a *logarithmic* distribution (Figure 2-7B), for reasons that go way beyond the scope of this course.

1.4.5 This discrepancy has to do with the difference between **laminar flow** and **turbulent flow**, which we're sure you've all heard about one way or another. Channel flows that are very shallow and/or very slow-flowing (like sheet flow or overland flow on the land surface during and after a rain) are characterized by the regular and locally rectilinear flow trajectories of laminar flow (Figure 2-8A), whereas channel flows that are deeper and/or faster-moving (almost all channelized flows fall into this category) show the irregularly sinuous flow trajectories characteristic of turbulent flow (Figure 2-8B).
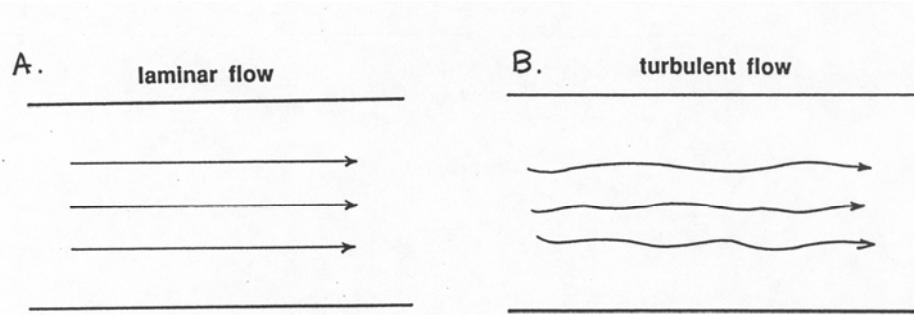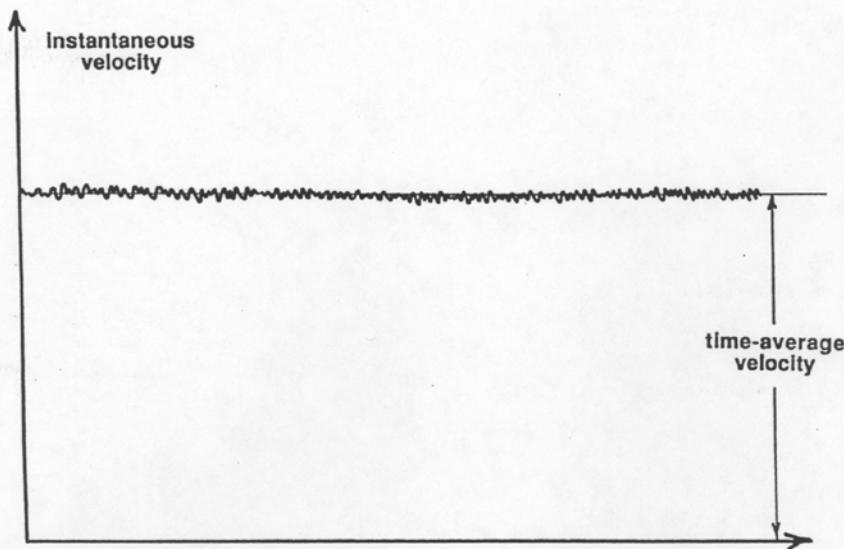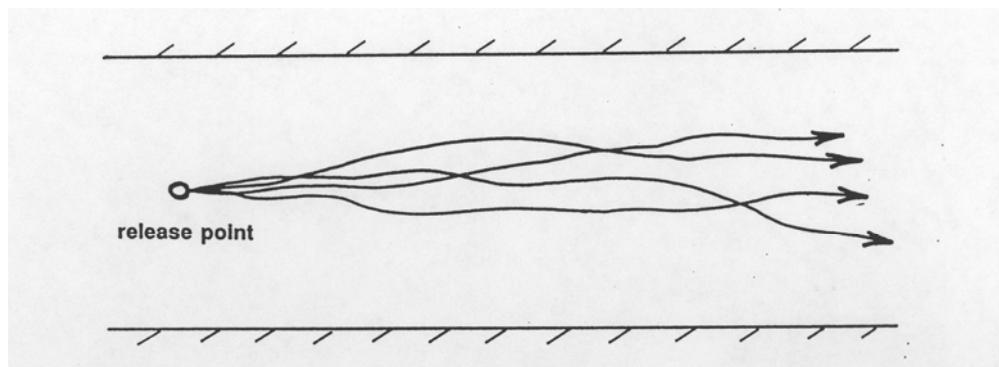

Figure 2-8   Paths of fluid elements in laminar and turbulent flow

1.4.6 Turbulence isn't easy to describe. But because it's such an important aspect of sediment transport, we'll attempt to give you a qualitative picture of what turbulence is like in a flow like in a channel flow. One way of studying turbulence is to *make a continuous measurement of velocity as a function of time at a fixed point in the flow*. You'd obtain a record that looks something like the graph shown in Figure 2-9: velocity fluctuations with a range of magnitudes and time scales

would be present around the time-average velocity. Another way of seeing the turbulence is to *release small neutrally buoyant markers from some fixed point in the flow and watch them move downstream* (Figure 2-10). The trajectories would be irregularly sinuous, with angles of deviation from the downstream direction amounting to no more than several degrees. Each trajectory would be different in detail from all of the others.
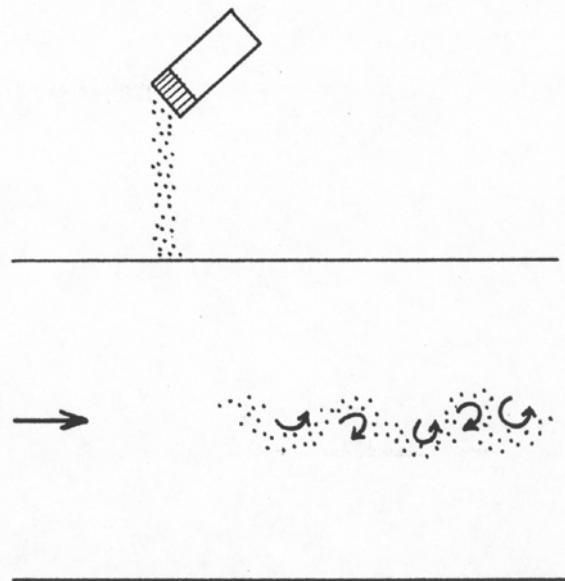

Figure 2-9   Typical record of a continuous measurement of velocity as a function of time at a fixed point in the flow

Figure 2-10   Trajectories of small neutrally buoyant markers released from some fixed point in a turbulent channel flow

1.4.7   The best way to see the turbulence, however, is to *sprinkle the flow with magic powder that allows you to see the eddies* (Figure 2-11). You'd see *rotating swirls of fluid that grade continuously into one another and that change their shape continuously with time*. Individual eddies retain their identity for a certain period of time (larger eddies live longer than smaller eddies) but any given

eddy eventually becomes unrecognizable and is replaced with newly developed swirls. The flowing surface of a river carrying fine sediment in suspension shows eddies moving up to the water surface to spread out and flatten themselves.
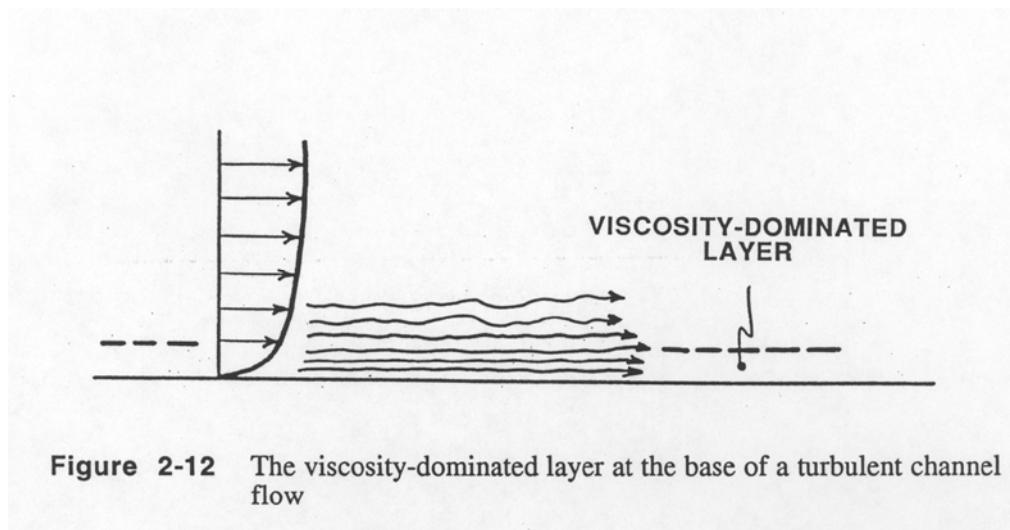

Figure 2-11   Sprinkling the flow with magic powder that allows you to see the eddies

1.4.8 The maximum vertical scale of eddies in the flow is not much smaller than the depth of the flow, and the maximum cross-flow scale is even larger. There's a continuous range of sizes down to very small eddies, of the order of fractions of a millimeter. The smaller eddies are superimposed on the larger eddies. As the large eddies sweep along the boundary they cause periods of stronger flow and weaker flow. When you're standing on a broad plain or on the deck of a ship at sea, the wind gusts you feel are a manifestation of these large eddies affecting the flow boundary. Have you ever watched a broad field of tall grass or grain in a strong wind from a point high above? You'd see a striking pattern of wind gusts moving along in the direction of the wind as they sway the grass.

1.4.9 Here's an interesting but incidental aspect of turbulence in a channel flow. In terms of the *kinetic energy of the flow*, the total kinetic energy is held partly in the mean flow and partly in the turbulent eddies that are superimposed on that mean flow. As the large eddies keep forming from the mean flow, they take some of the kinetic energy of the mean flow with them. The natural tendency is for the bigger eddies to subdivide themselves into smaller eddies—have you ever seen it work the other way around?—so *the kinetic energy of the turbulence is inevitably passed down to smaller and smaller eddy scales, finally to be dissipated into heat mainly at the smallest eddy scales*, where (it turns out) the local shearing is strongest. This is picturesquely called the ***energy cascade***. (In case you're wondering how the mean flow replenishes its mechanical energy, it gets it from the loss of potential energy as the fluid moves downslope.)

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1.4.10 Returning now to the important qualitative difference between laminar and turbulent velocity distributions, *in turbulent flow the velocity distribution is much more uniform over most of the thickness of the flow, but changes much more sharply very close to the boundary.* It's easy to understand why this is so. In turbulent flow, over most of the flow depth the lateral motions of eddies tend to even out the differences in time-average velocity from layer to layer. This is essentially a matter of diffusion of fluid momentum. (Diffusion, remember, is the flux of some property by random motions in the present of a gradient in the mean value of that property.) As we move closer and closer to the boundary, however, the fluid motions perpendicular to the boundary are more and more restricted—because remember that right at the boundary the fluid velocity must match that of the boundary, so there can be *no* component of velocity normal to the boundary there. In a thin layer next to the boundary, therefore, the turbulence can't even out the differences in velocity from layer to layer, so the gradient of velocity is very steep. This *layer next to the boundary, where viscous shearing is more important than turbulence,* could be called the **viscosity-dominated layer** (Figure 2-12). Its thickness is typically a very small percentage of the flow depth, no more than millimeters to at most a few centimeters.
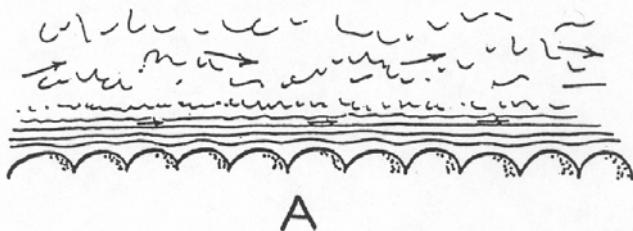

Figure 2-12   The viscosity-dominated layer at the base of a turbulent channel flow

1.4.11 But there's more to this matter than meets the eye. The above is a good picture when the roughness elements on the bed (sediment particles and/or bed forms) are small enough to be embedded within this viscosity-dominated layer. This is called **dynamically smooth flow**—even though the boundary may be physically rough (Figure 2-13A). But in many cases the roughness elements are taller than what the thickness of the viscosity-dominated layer would be if there were a continuous viscosity-dominated layer in the first place. You'll have to bear with that rather complicated verbal construction, because we're dealing now with a situation in which the roughness elements stick up into the turbulent part of the flow and there's no longer an overall, continuous viscosity-dominated layer. But keep in mind that there's still a local viscosity-dominated layer, or even just a film, at the surfaces of all the roughness elements themselves. This is called **dynamically rough flow** (Figure 2-13B).
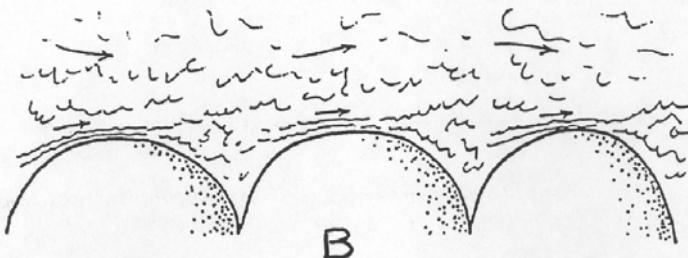
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DYNAMICALLY SMOOTH FLOW

particles embedded in
viscosity-dominated layer

A

DYNAMICALLY ROUGH FLOW

particles stick up through
viscosity-dominated layer

B

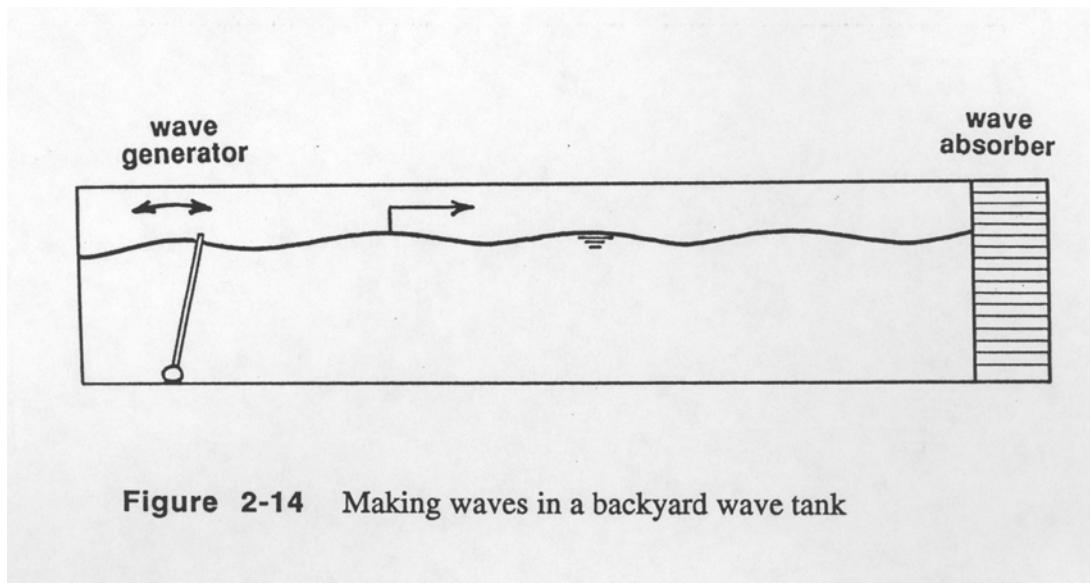**Figure 2-13**   A: dynamically smooth flow B: dynamically rough flow

1.5 Oscillatory Flows

1.5.1 Everything in the preceding sections is relevant to ***unidirectional flow***: *a current that moves in one direction only*, with velocity either unchanging with time (***steady flow***) or changing somehow with time (***unsteady flow***). Also important in the transport of sediments is ***oscillatory flow***: *a current that reverses its direction periodically with time*. The approach in this section will be somewhat different from the approach in the preceding sections: we'll concentrate more on the origin and description of oscillatory flows than on their dynamics.

1.5.2 Except for reversing tidal currents, which technically are oscillatory flows with very long oscillation periods but which are better considered as re-versing unidirectional flows, almost all of the important oscillatory flows in nature are caused by ***surface gravity waves*** which propagate on the surfaces of oceans and large lakes. Such waves are called surface waves because they're associated with the water surface, and they're called gravity waves because the dominant force involved in their propagation (the force which attempts to restore the deformed water surface to a planar condition) is the force of gravity.
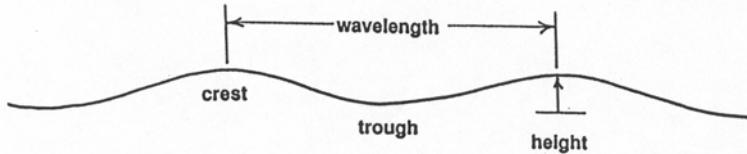
1.5.3 The generation of wind waves is a complex process, even now not fully understood, that involves both the shear stress of the moving air on the water surface and on the unequal distribution of air pressure over the windward and leeward sides of geometrical irregularities on the water surface. We'll note here only that the size of the waves depends on three factors: the *speed* of the wind, the *duration* of time the wind acts on the water surface, and the downwind *distance* (called the **fetch**) over which the wind acts on the water surface.

1.5.4 The best way to obtain an idea of the basic nature of the water motions produced by surface gravity waves is to watch the propagation of a single train of waves in a wave tank (Figure 2-14). You can imagine building a long tank with rectangular cross section, perhaps several meters long, open at the top. Put a wave generator at one end. A board hinged to the bottom of the tank can be rocked or oscillated back and forth to produce a train of waves. It's best to install some porous material at the other end of the tank to absorb the waves, so that they're not reflected to interfere with the propagating wave train.
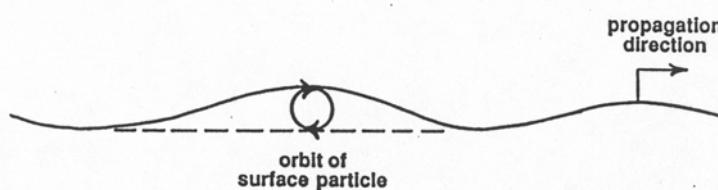

Figure 2-14   Making waves in a backyard wave tank

1.5.5 Here are some terms used to describe the waves in a simple train of waves (Figure 2-15). The **crest** of the wave is *the highest point on the wave profile*, and the **trough** is *the lowest point on the wave profile*. The **period** of the wave is *the time needed for one complete wave form to pass a point which is fixed relative to the bottom of the tank*. Surface gravity waves in natural environments have periods that range from less than a second to almost 20 s. The **wavelength** of the waves is *the distance from crest to crest or from trough to trough*. The **height** of the waves is *the vertical distance from trough to crest*. The heights of large waves generated by major storms are commonly in the range of five to ten meters, but they can be even larger.
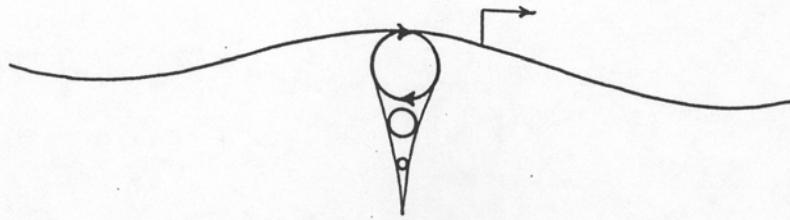
Figure 2-15 Some terms used to describe the waves in a simple train of waves

1.5.6 If you place a floating cork on the water surface and watch its motion through the wall of the tank as the waves pass by, you see that the cork travels in *an almost perfectly circular orbit,* making one circle during the passage of each wave (Figure 2-16). The sense of motion is such that the cork moves in the direction of wave propagation while the crest of the wave is passing, and opposite to the direction of wave propagation while the trough of the wave is passing.
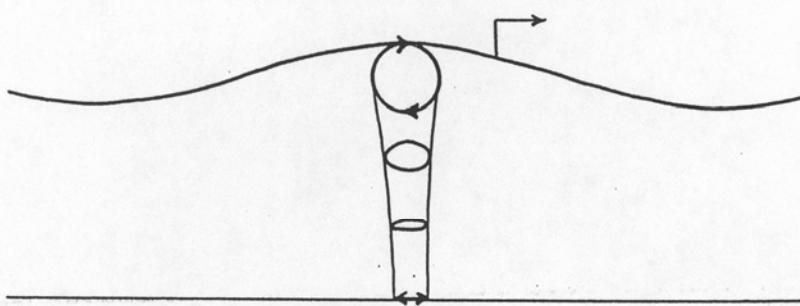

Figure 2-16 The circular orbit of a surface particle floating on the water surface as a wave passes

1.5.7 The water below the surface also moves as the waves pass, but the size of the orbits of the water particles decreases very rapidly with depth below the surface (Figure 2-17). *The size of the orbits at a depth equal to half a wavelength is only about ten percent of the size of the orbits at the surface, and the size of the orbits at a depth equal to one full wavelength is negligible.* Such waves are called ***deep-water waves***. If the wavelength of the waves is much greater than the water depth, however, the size of the orbits is still appreciable at the bottom (Figure 2-18). Such waves are called ***shallow-water waves***, and they are common during storms. Downward from the surface the orbits become flattened into *ellipses* with greater and greater eccentricity, until at the very bottom the orbits are back-and-forth *straight lines*. This should make sense to you, because *at the bottom there can be no fluid motion perpendicular to the bottom*—if the slight flow into and out of the porous sediment is ignored.
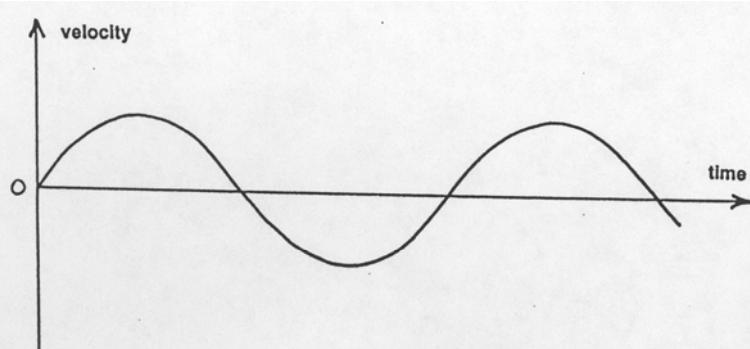
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Figure 2-17   Decrease in orbit diameter with depth beneath a deep-water wave

Figure 2-18   Decrease in orbit diameter with depth beneath a shallow-water wave

1.5.8 Now examine more closely the nature of the oscillatory water motion at the bottom. If you measured the fluid velocity near the bottom as a function of time at a point which is fixed relative to the bottom, you'd obtain a record which looks like the graph in Figure 2-19: the velocity varies *sinusoidally* with time. The nature of the motion can be characterized by its period T, or by the maximum velocity U_m which is attained during one oscillation cycle (this happens when the water is at the central point of its oscillation path) or by the distance d_o moved by a water particle during one full oscillation (this is called the **orbital diameter**). Only one of the latter two quantities is independent, however, because the period, the maximum velocity, and the orbital diameter are related by the equation $U_m T = \pi d_o$. This relation is easy to derive, but we won't do that here.

Figure 2-19   Fluid velocity near the bottom as a function of time at a point that's fixed relative to the bottom, as a shallow-water wave passes by overhead

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1.5.9 It's part of the physical nature of waves that *different trains of surface gravity waves can be superimposed on each other in such a way that the water-surface height and water motion at a given point is the sum of contributions from each superimposed wave train*. You might know from watching storms at sea or on large lakes that the sea state produced by a strong wind is complicated, because it's a superposition of a large number of separate trains of waves with slightly different periods, propagation directions, and heights. Also, *when the wind direction changes as a storm passes, the waves adjust their direction of propagation in response*. In such cases, the pattern of water movement at the bottom is no longer a simple back-and-forth movement. Even the relatively simple case of two regular trains of waves moving in different directions, the bottom water movement can be surprisingly complicated, to say nothing of the really complicated wave conditions produced during storms.

1.6 The Wind

1.6.1 The large-scale structure of the lower part of the atmosphere is obviously much different from that of even the largest river. The most obvious difference is *the absence of a free surface* in the atmosphere. An equally important difference has to do with *the much larger scale* of wind systems. This is important because the Earth's rotation plays a dominant role in determining wind direction but is a minor aspect of even the largest of rivers. A third important difference is related to the easy *compressibility* of air, in contrast to the incompressibility of water. This means that the thermodynamics of compressibility plays an enormous role in the behavior of the atmosphere.

1.6.2 Despite these far-reaching differences, it's fortunate for us as Earthbound sedimentologists that *the structure of the wind in the near-surface layer, up to hundreds or even a few thousands of meters above the surface, is usually not greatly different from that of large-scale channel flows*, in terms of such things as velocity distribution, turbulence structure, and boundary forces. Most of what we said in Section 1.4 above for channel flows of water is largely applicable to the wind. This is especially true of eolian sand transport and the shaping of sand dunes—certainly the most important aspect of eolian sedimentary processes for this course.

1.6.3 You might complain that the wind is so much more gusty than the ordinarily turbulent flow of rivers. We would counter that by saying that your conception of the gustiness of the wind is shaped by observations made for the most part down in among large and very irregularly shaped and irregularly distributed roughness elements like trees, hills, and buildings. To the extent things like that occupy river bottoms (hills are certainly there, and we'll have to deal with them as a really important part of this course), and you sensed turbulence down at their level, flow in rivers would seem pretty "gusty" too.   (There'll be more on that later, when we talk about flow over subaqueous bed forms.)   If you've ever experienced a strong wind at sea, you know that *the wind unimpeded by large roughness elements is surprisingly steady*.
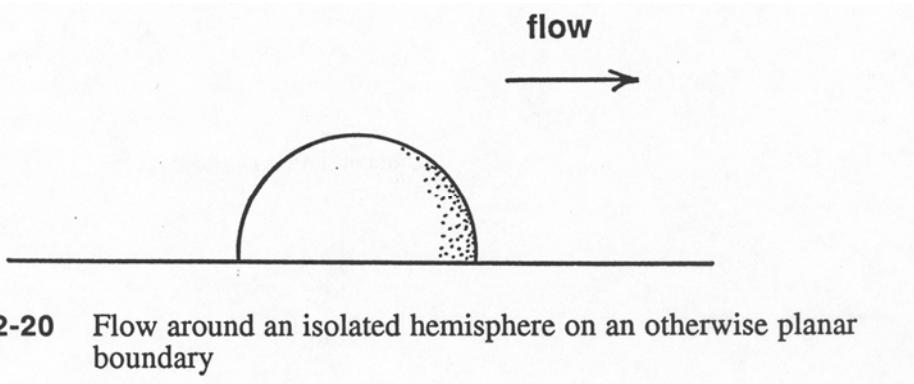
2. BASICS OF SEDIMENT TRANSPORT

2.1 Introduction

2.1.1 It seems crazy to us to start dealing with physical sedimentary structures before presenting some material on how sediment is transported and deposited in the first place, so we've included this second background chapter. It's entirely qualitative but deals with what we think are the important topics: forces on sediment particles resting on a sediment bed: the threshold of particle movement; the nature of the sediment load; and just a bit on transport rates and mixed-size sediments.
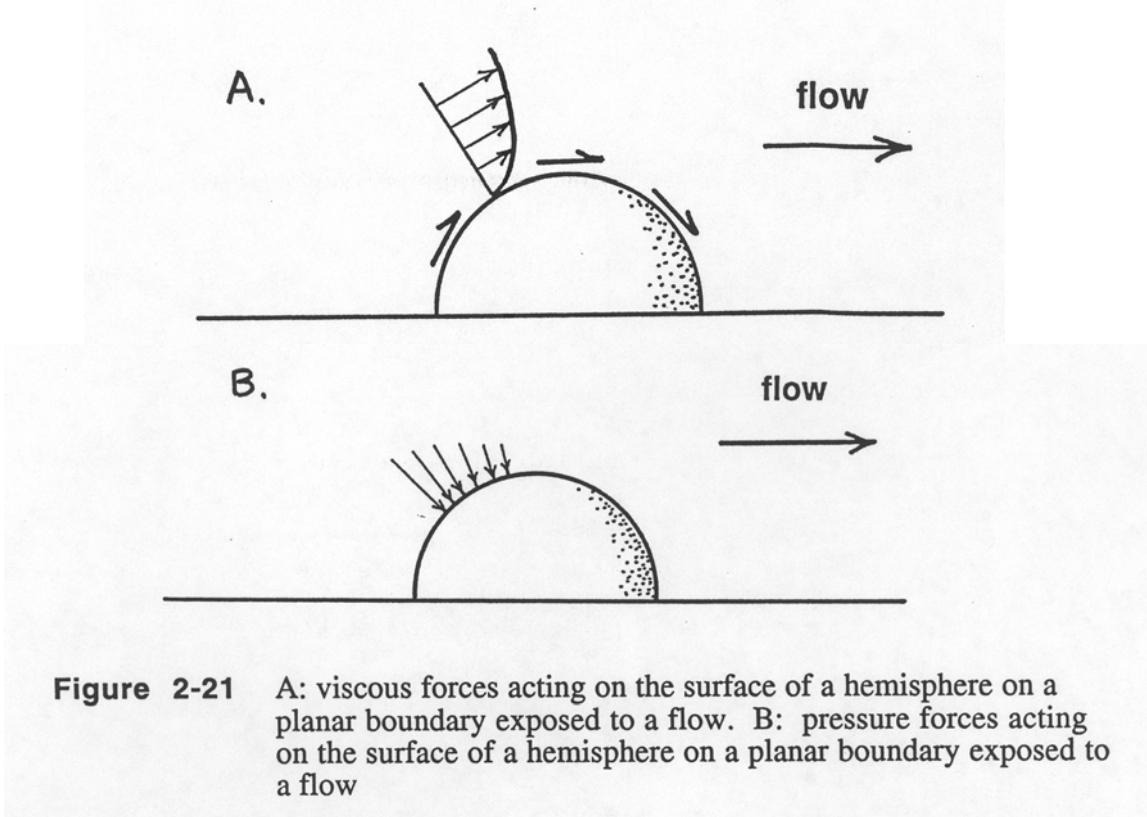

2.2 Forces on Particles

2.2.1   You saw in Chapter 2 that a fluid flow exerts forces on its solid boundary. In particular, *the flow exerts forces on all of the sediment grains exposed at the surface of a loose sediment bed*. If the force exerted by the flow on a given sediment particle is momentarily great enough, the particle is set into motion. This section presents a brief discussion of the nature of fluid forces on sediment particles resting on the sediment bed.

2.2.2 Look at the flow around an isolated hemisphere on an otherwise planar boundary (Figure 2-20).   This is a simplified but qualitatively representative model for the more complicated forces on particles of irregular shape resting on a bed of similar particles.

Figure 2-20   Flow around an isolated hemisphere on an otherwise planar boundary
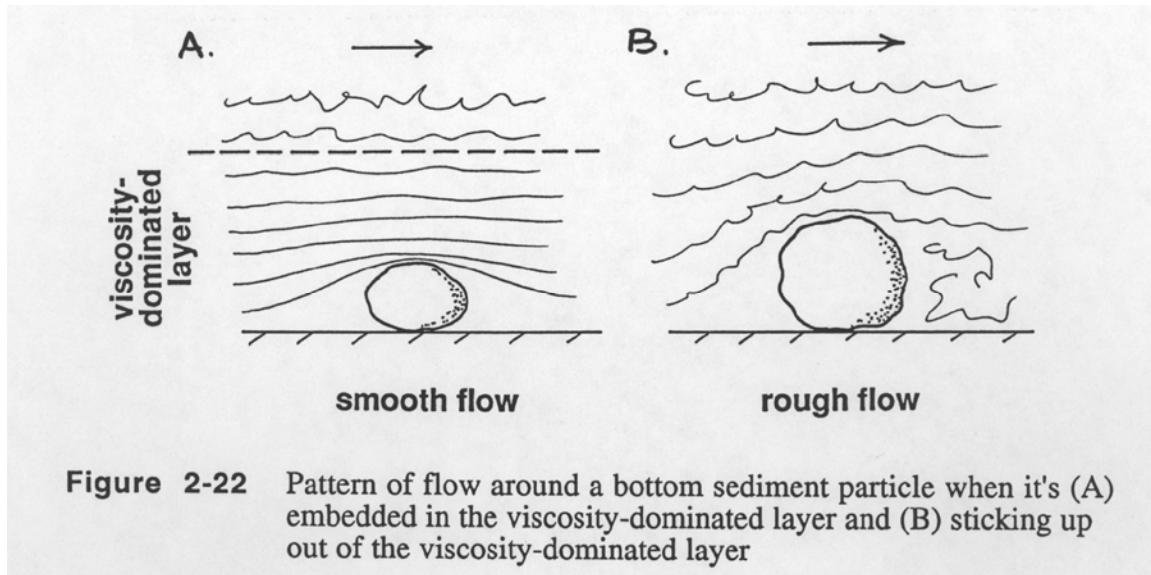
 2.2.3 The forces acting on the smooth surface of the hemisphere are of two kinds, *viscous forces* and *pressure forces*. Viscous forces are exerted tangential to the solid surface. They are present because the fluid is undergoing shear next to the solid surface (Figure 2-21A). Remember that whenever a fluid is sheared, a viscous resistance force is present. Pressure forces are easier to understand. Fluids exert pressure within themselves and also on all solid boundaries. On the solid boundaries the fluid pressure is manifested as a force per unit area acting perpendicular to the boundary (Figure 2-21B).

Figure 2-21   A: viscous forces acting on the surface of a hemisphere on a planar boundary exposed to a flow. B: pressure forces acting on the surface of a hemisphere on a planar boundary exposed to a flow

2.2.4   Both the viscous forces and the pressure forces vary from point to point all over the surface of the hemisph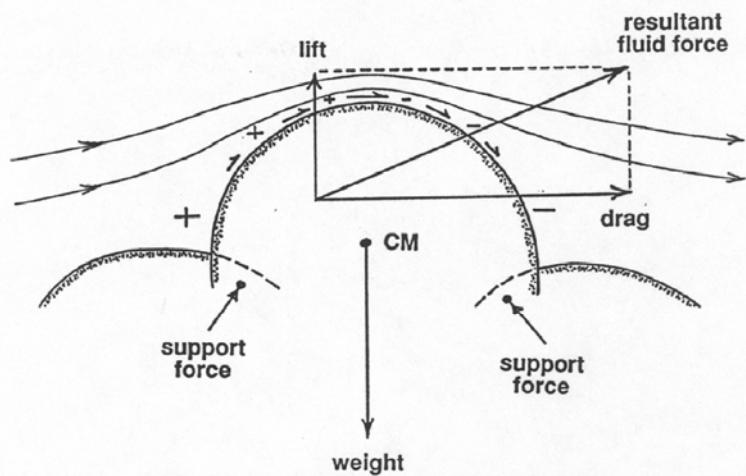ere in Figure 2-20. To find the total force or resultant force exerted by the fluid on the hemisphere, you need to *sum or integrate both kinds of forces over the surface of the hemisphere*. Although this is easy to do in one's imagination, in practice it's difficult even for very large hemispheres, and it's impossible for tiny sediment particles.

2.2.5 The nature of the forces on a bed particle depends very strongly on *the height of the hemisphere relative to the thickness of the viscosity-dominated layer*. If the height of the hemisphere is much smaller than the thickness of the viscous sublayer (Figure 2-22A), the flow around the particle is only weakly turbulent, and the forces exerted by the flow are approximately the same as in laminar flow. If the height of the hemisphere is much greater than the thickness of the viscous sublayer (Figure 2-22B), the flow around the hemisphere is strongly turbulent, and also the flow pattern is greatly different.

Figure 2-22 Pattern of flow around a bottom sediment particle when it's (A) embedded in the viscosity-dominated layer and (B) sticking up out of the viscosity-dominated layer
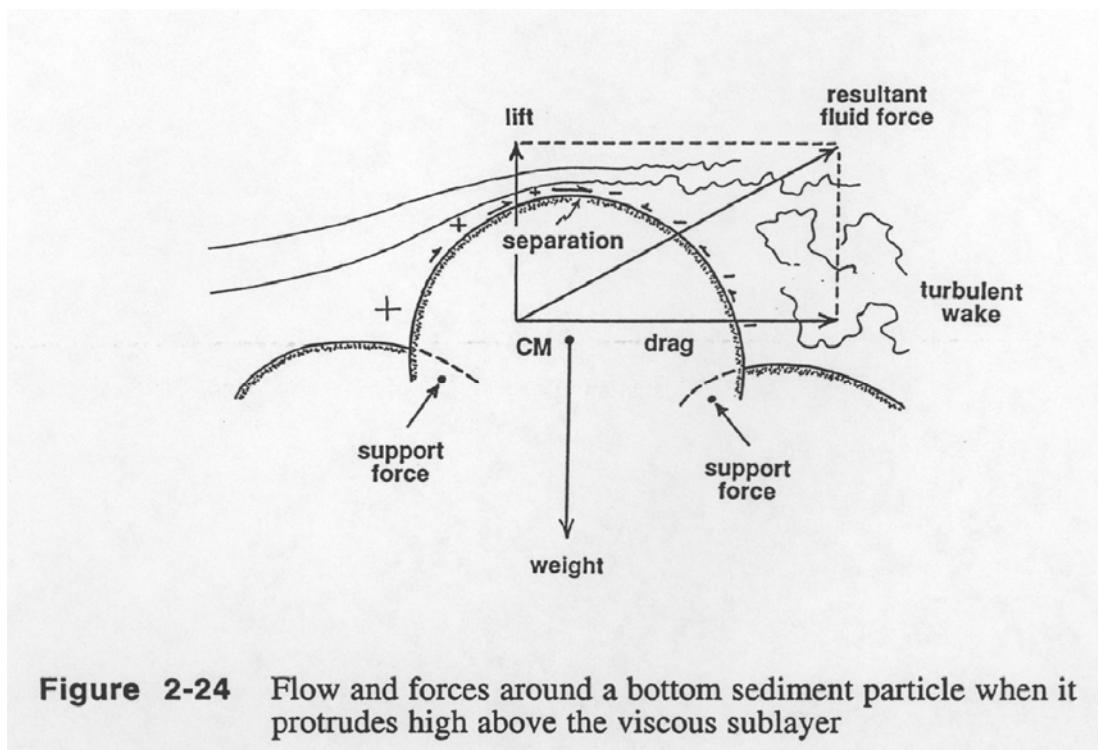
2.2.6 Figure 2-23 is a cartoon showing the flow and forces around the hemisphere when it's embedded in the viscous sublayer. The flow pattern shows streamlines diverging smoothly in front of the hemisphere, converging at the midsection, and diverging again behind the hemisphere. Turbulence in the flow is weak, and the presence of the hemisphere itself does not create any new turbulence. The fluid pressure is relatively high on the front of the hemisphere and relatively low on the back, resulting in a net downstream pressure force. It turns out that the viscous forces and the pressure forces are of about the same magnitude.

Figure 2-23 Flow and forces around a bottom sediment particle when it's embedded in the viscosity-dominated layer

2.2.7 Figure 2-24 is the corresponding cartoon showing the flow and forces around the hemisphere when it protrudes high above the viscous sublayer. The

streamlines diverge smoothly in front of the hemisphere, but *instead of converging again they separate from the surface of the hemisphere at the midsection and shoot off into the flow, leaving a region of stagnant fluid behind the hemisphere.* This phenomenon is called **flow separation**. Flow separation is inevitable around blunt bodies when the flow is fast and the body is large. Flow separation is characteristic of buildings in the wind and trucks on the road, as well as of bed forms and large sediment particles. The pressure in the stagnant fluid behind the hemisphere is very low everywhere, so the resultant pressure force on the hemisphere is very large, much larger than the viscous force.

Figure 2-24   Flow and forces around a bottom sediment particle when it protrudes high above the viscous sublayer

2.2.8 Note in Figures 2-23 and 2-24 that because the pressure is relatively low at the top of the hemisphere and relatively high around its base, there's a **lift** component to the resultant force. This is exactly the same kind of lift force that acts on airplane wings. The lift force can be almost equal to the downstream component of the resultant force (which is called the **drag**).

2.2.9 Let us remind you that the forces on the hemisphere would vary sharply with time, because of the passage of larger eddies near the boundary. *Forces can vary by factors of four or even more with time.* The time scales for change in forces on a sediment particle can be very short, fractions of a second.

2.3 Threshold of Movement

2.3.1 One of the classic problems in sediment transport is that of the ***threshold of movement***, or ***incipient movement***: *how strong a flow is needed to set the particles of a given sediment bed in motion?* This problem is obviously important for engineering purposes. An equivalent problem, which is more relevant to geology, is that of ***competence***: *how coarse a sediment can be moved by a given flow?*

2.3.2 The problem of incipient movement should be one of the simplest in sediment transport mechanics, because the sediment hasn't yet started to move, so the structure of turbulence and the pattern of fluid forces are not complicated by the presence of moving grains. Actually the problem turns out to be mostly intractable theoretically, and both engineers and sedimentologists have relied upon empirical approaches.

2.3.3 Our common sense tells us that *stronger flows are needed to move coarser sediment*. In this case, our common sense is correct! (That's not always true, especially in sediment transport mechanics.) But to know just how strong a flow is needed to move a given sediment, observation and experiment are needed.

2.3.4 Obviously the sorting and particle shape are important, because the details of the fluid forces on the particles depend both on the particle shape and on how the uppermost particles rest upon the underlying particles. In dealing with the movement threshold, it's usually assumed that the sediment is naturally well sorted, and equant and subrounded in shape. Not much work has been done on the effect of sorting and shape on the threshold of movement.

2.3.5 The threshold conditions should depend on *independent variables that describe the flow, the fluid, and the sediment*. Either the bed shear stress or the flow velocity could be used to describe the flow. The trouble with using the velocity is that it's not clear which velocity to use: the velocity at the level of the tops of the particles, which is impossible to measure or estimate, or some other velocity like the mean velocity of flow. The problem with using the mean velocity is that then the flow depth must also be considered, because many combinations of flow depth and flow velocity are equivalent to a given boundary shear stress. But it's easier for us to attach concrete significance to the flow velocity than to the bed shear stress.
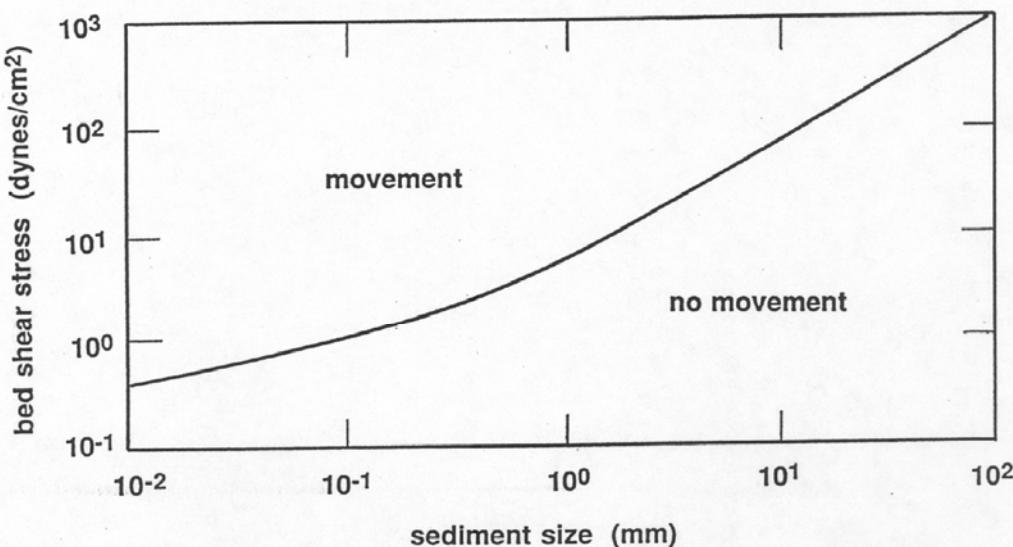
2.3.6 The best way to approach the problem of initiation of motion empirically is to decide which are the most important variables that govern whether particles are moved or not by the flow and then plot a graph that separates conditions of no movement from conditions of movement (by a curve in a two-dimensional graph or a surface in a three-dimensional graph).

2.3.7 In the general problem, lots of variables might be important: the density ρ and the viscosity μ of the fluid, the diameter D of the sediment particles (to say nothing of sorting and particle shape), and the strength of the flow, best characterized by the bed shear stress τ_o. But for flows of water-density fluid over quartz density sediment (and let's assume that sorting and shape are of secondary importance, which is probably true for "ordinary" sediments), then only D and τ_o

vary. So we can get away nicely with *a two-dimensional graph of boundary shear stress against sediment size. The curve that separates movement and no movement is called the* **threshold curve**.

2.3.8 If you read the literature on sediment movement you're likely to see mention of the *Shields diagram*, which is a similar graph but arranged in a special dimensionless form that's applicable (in theory!) to all combinations of fluids and sediments. What we're doing here is a shortcut but essentially the same thing.
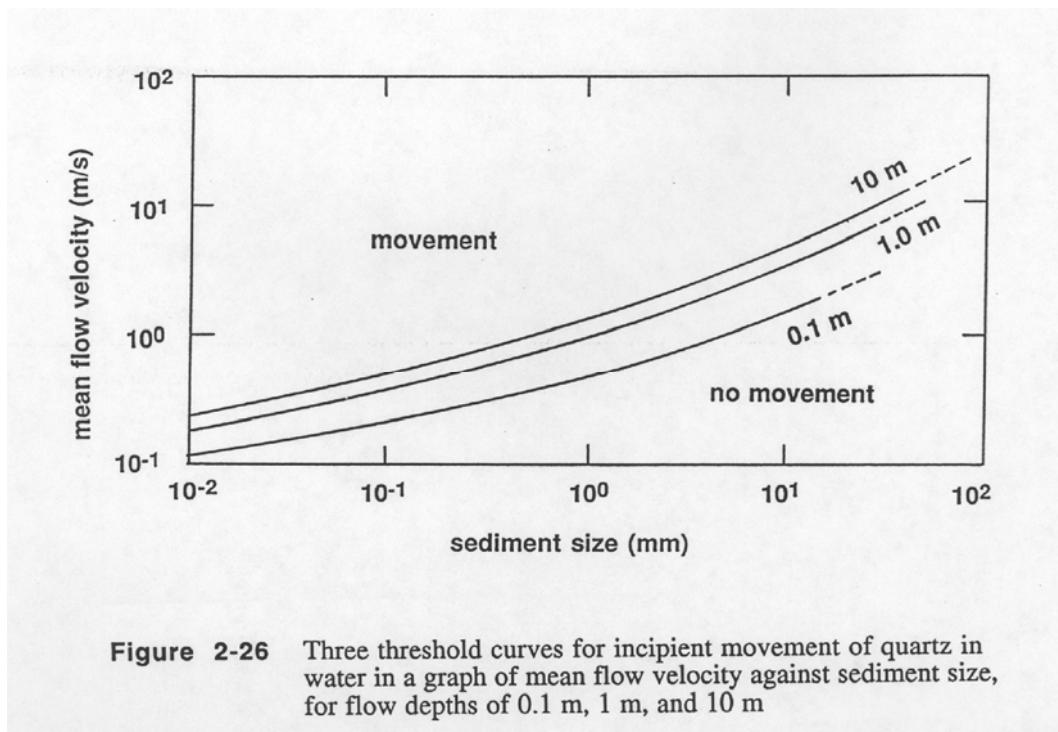
2.3.9 Figure 2-25 shows the threshold curve for quartz sediment in water. There's still the problem of concretely visualizing the boundary shear stress. Mean flow velocity would be much better for that, but then the problem is that there's a slightly different curve for each flow depth, because there's a whole range of combinations of flow depth and flow velocity that five the same boundary shear stress. Figure 2-26 shows three threshold curves for quartz in water in a graph of mean flow velocity against sediment size, for flow depths of 0.1 m, 1 m, and 10 m. Figure 2-26 should give you a concrete idea of threshold conditions in water flows.

Figure 2-25 Threshold curve for incipient movement of quartz sediment in water
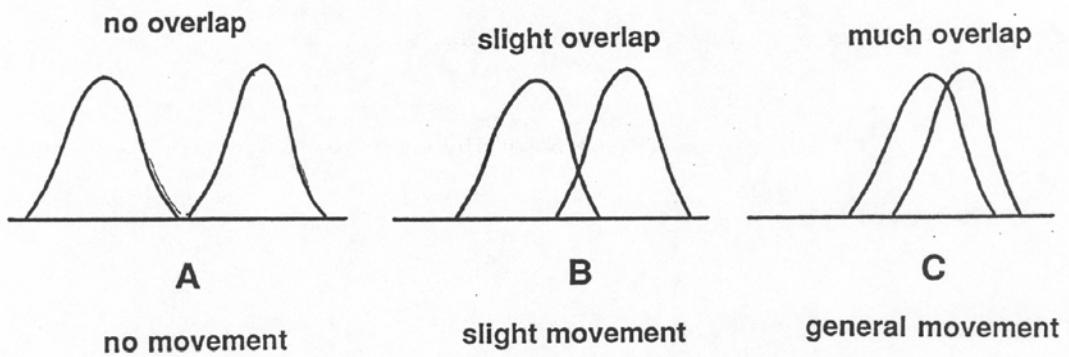
2.3.10 The data that have been used to find the curves in Figures 2-25 and 2-26 show quite a lot of scatter. This is mainly because *the range of flow conditions for which there is slight sediment movement is very wide*. In other words, *it's difficult to define the threshold of movement*. You could observe this for yourself by preparing a planar sediment bed in a laboratory channel and then passing a flow of water over it. As you gradually increase the flow strength,

eventually there's slight grain movement, in the form of pulses or gusts at random points on the bed where one or more grains are moved by a locally strong eddy. As you continue to increase the flow strength, the frequency of the pulses or gusts increases, and the number of grains moved by each pulse or gust increases as well. Eventually there's abundant grain movement.

Figure 2-26 Three threshold curves for incipient movement of quartz in water in a graph of mean flow velocity against sediment size, for flow depths of 0.1 m, 1 m, and 10 m

2.3.11 One big problem with all of this development is that *it applies only to incipient movement on a perfectly planar preexisting sediment surface.* When was the last time you saw a perfectly planar sediment surface in nature? (They do exist, but they're certainly not common.) *Sediment movement on bed forms starts at much lower flow strengths*, just because the flow is concentrated at the crests of the bed forms.

2.3.12 Incipient movement is thus *probabilistic* or *stochastic*. A good way of thinking about this is in terms of the relationship between two different probability distributions (Figure 2-27): the distribution of instantaneous local bed shear stress needed to move the set of particles in some area of the bed, and the distribution of local bed shear stress which acts on any small area of the bed through time. When the two distributions don't overlap, the flow is never strong enough to move any of the grains (Figure 2-27A). When the two distributions overlap slightly, a small subset of the bed particles are moved by the flow (Figure 2-27B). With increasing flow strength the distributions eventually overlap entirely, and all the bed particles are susceptible to movement (Figure 2-27C).

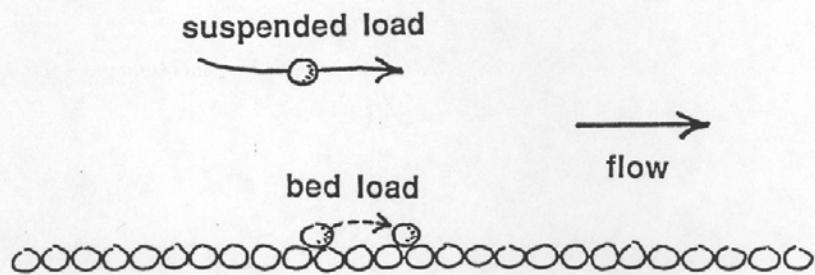| no overlap	slight overlap	much overlap
A	B	C
no movement	slight movement	general movement

Figure 2-27 Threshold condition viewed in terms of two probability distributions. In each sketch, the curve on the left shows the distribution of instantaneous local fluid forces on bottom sediment particles, and the curve on the right shows the distribution of instantaneous fluid forces needed to set the bed particles in motion.
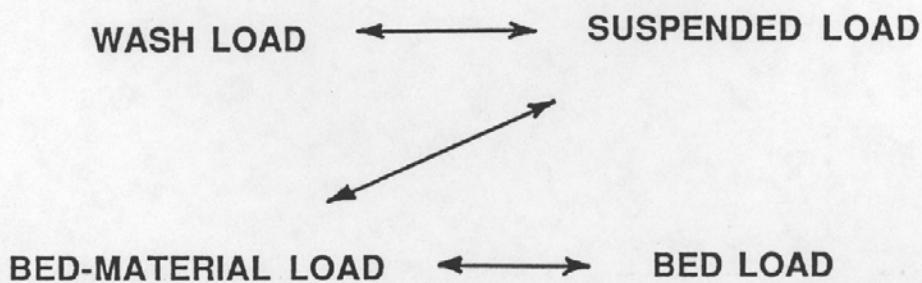
2.4 The Sediment Load

2.4.1 Classification of the Sediment Load

2.4.1.1 *The aggregate of sediment particles that are transported by a flow at a given time* is called the **load**. 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 in nonnegligible percentages*, 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 is always the finest fraction of the load, is carried through a long segment of the flow without any exchange of sediment between the bed and the flow.

2.4.1.2 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 upwardly moving turbulent eddies* (Figure 2-28). We hope it's 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*. This may sound confusing, but it makes sense. Figure 2-29 may or may not be of help.
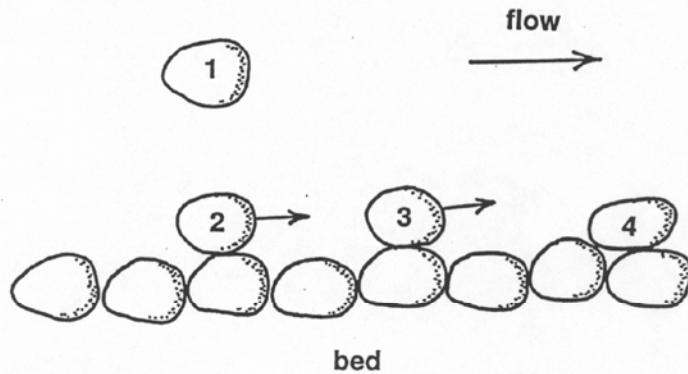
Figure 2-28 Bed load and suspended load

Figure 2-29 Relationships among the different kinds of sediment load
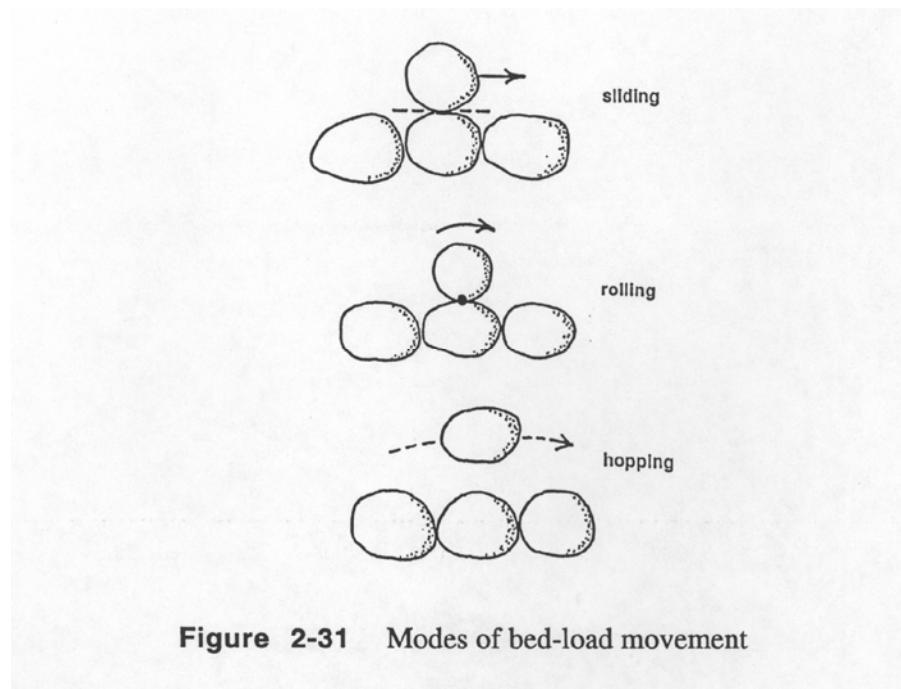
2.4.1.3 Keep in mind that *the set of particles that form the load keeps changing from time to time*, because particles are continually coming to rest and being set into motion again. The best way to make the distinction between the bed and the load is to consider that a particle in motion is part of the load, even if it is still in contact with the bed (Figure 2-30). When a particle starts to move, it stops being a bed particle and starts being part of the load. Conversely, when it comes to rest again it tops being part of the load and becomes a bed particle again.

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Figure 2-30 Particles that belong to the bed and that belong to the load. Particle 1 is part of the flow. Particle 2 has just come into contact with the bed but it's still moving so it's part of the flow. Particle 3 has just been set into motion, so it's part of the flow. Particle 4 is exposed to the flow but is at rest so it's part of the bed.

2.4.2 Bed Load
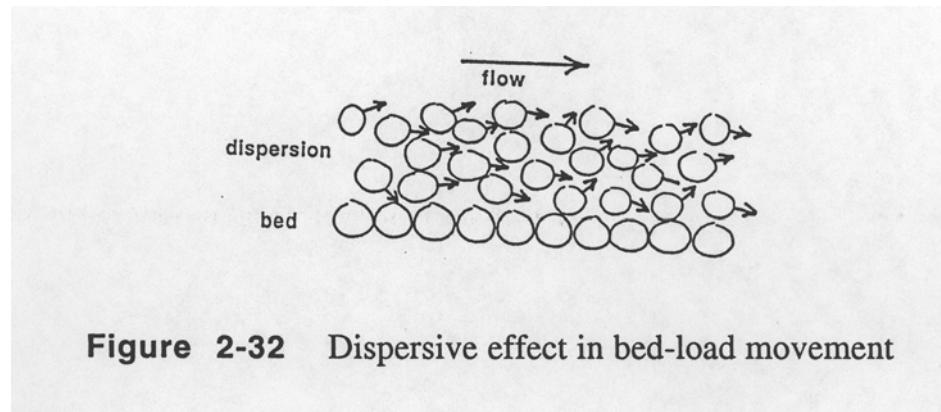
2.4.2.1 The movement of bed load is sometimes called ***traction***. Bed-load movement can be by *rolling, sliding, or hopping* (Figure 2-31). It's not easy to observe bed-load movement in great detail, but when you have the chance to watch a carefully made high-speed closeup motion picture of bed load (in flows where the load is not yet so abundant as to obscure the view) you see that the particles characteristically take occasional excursions downstream, by rolling and hopping irregularly with brief stops along the way, and then come to rest for some time before being moved again. The explanation is apparently that once a grain is dislodged from a place of rest it's susceptible to continued movement by the flow until it finally finds a rather sheltered position, and then it's not dislodged again until it's affected by an especially strong shear stress or until one or more of the bed particles sheltering it are themselves put into motion.

Figure 2-31   Modes of bed-load movement

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2.4.2.2   The problem in discussing bed load is that *we know very little about the characteristic modes of bed-load movement except in weak flows in which not much sediment is moving*. This is an area in which there has been much speculation supported by very little observational data. And yet such conditions are very important sedimentologically. One popular theory, which is intuitively attractive, is that when there's a high concentration of moving bed-load particles, the particles farther away from the bed are maintained partly or largely in their motion by collisions with particles moving closer to the bed (Figure 2-32). This involves a kind of *dispersive* effect, and the dispersion of grains maintained above the bed in this way has been called a **traction carpet**.   Unfortunately the importance of this dispersive effect is still not known.

Figure 2-32   Dispersive effect in bed-load movement

2.4.3 Suspended Load

2.4.3.1   There's clearly a problem in distinguishing between bed load and suspended load: how far can a grain move up into the flow and still be considered bed load?   The standard criterion is *whether or not fluid turbulence has a sub-stantial effect on the time and distance involved in the excursion*. Although the distinction between bed load and suspended load is a convenient one, there's no sharp break between bed load and suspended load.
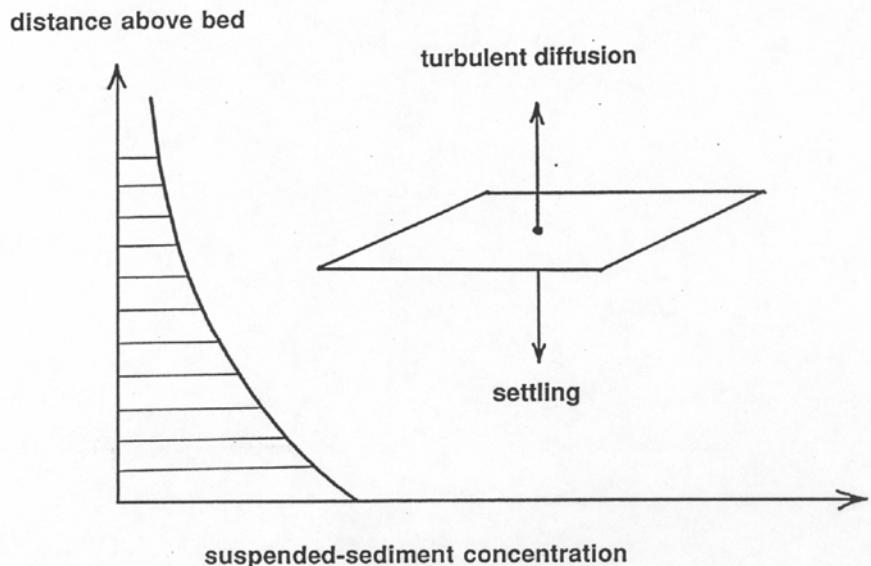
2.4.3.2 Keep in mind also that *a given particle can be part of the bed load at one moment and part of the suspended load at another moment*, depending upon the time history of fluid forces and motions to which it is subjected. (And of course at still other times the same particle might not be moving at all.) Therefore, at any given time *there's an appreciable overlap in the size distributions of the bed load and the suspended load*, although clearly the suspended load tends always to be finer on the average than the bed load.

2.4.3.3 *Particles moving as bed load are susceptible to being carried up into suspension when the maximum vertical turbulent velocity fluctuations are greater in magnitude than the settling velocities of the particles*. If the conditions of the flow and the settling velocities of the particles fulfill that condition, then some of

the moving bed-load particles will occasionally find themselves caught in a strong upward-moving eddy, and the particle will be carried for some distance above the bed. The particle will be affected by a series of eddies as it moves downstream; depending on the motions of the individual eddies, the particle may rise only a short distance from the bed and travel only a short distance downstream before it settles back to the bed, or it may rise high above the bed, even almost to the water surface, and travel far downstream. Obviously, *the smaller the settling velocity and the stronger the turbulence, the greater the average height above the bed and the greater the distance of travel downstream the particle will attain*.
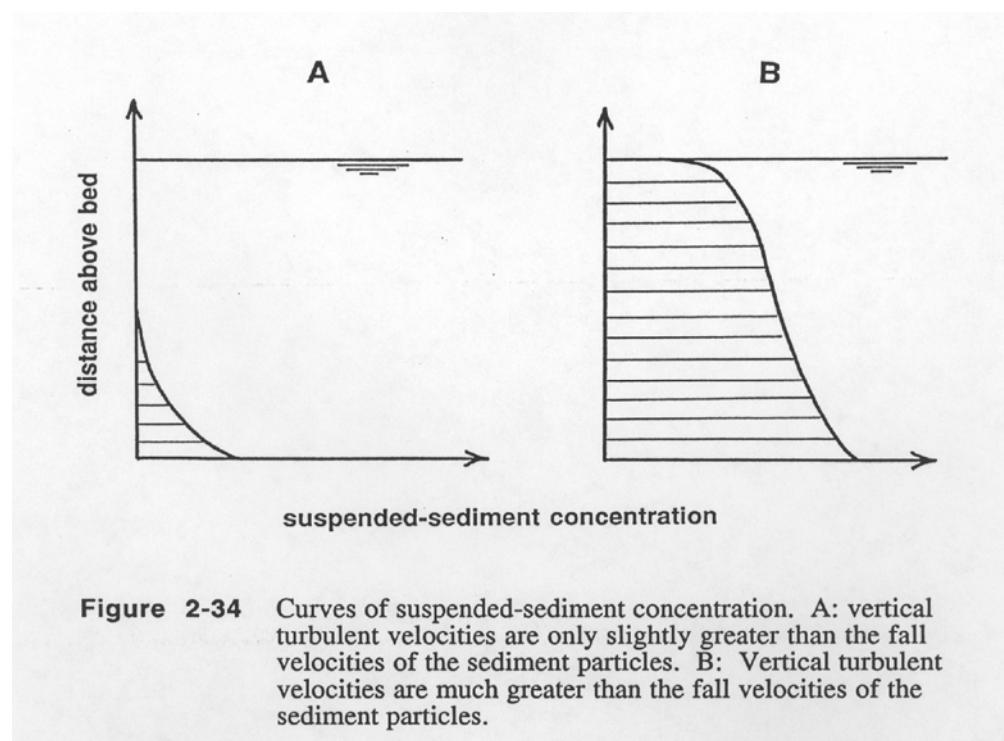
2.4.3.4 The sediment particles are not *really* suspended above the bed, in the way that a painting is suspended on a nail in the wall: they are always settling back toward the bed and will eventually return to the bed. Only particles of colloidal size, much finer than a micrometer, can be truly suspended. Such particles have such small mass that Brownian motions caused by the random collisions of molecules against the particle keep the particle in permanent suspension.

2.4.3.5 The controls on the vertical distribution of suspended sediment in steady and uniform flows are well understood. The basic idea is that *if one considers a small horizontal unit area somewhere in the flow, there must be a time-average balance between the rate at which sediment settles down through that unit area and the rate at which sediment is transported up through the unit area by turbulent diffusion* (Figure 2-33). (The idea behind turbulent diffusion is that because there is an upward decrease in suspended sediment concentration, the random motions of eddies across the unit area result in a net upward movement of suspended sediment even though the volumes of fluid exchanged cross the unit area are the same.)

Figure 2-33 Vertical balance between downward settling and upward turbulent diffusion in suspended-load transport

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2.4.3.6 Prediction of the vertical distribution of suspended load in a steady uniform channel flow on the basis of the foregoing idea is one of the few theoretical successes in the mechanics of sediment transport. The theory predicts *an exponential decrease in suspended-sediment concentration with height above the bed*. When the vertical turbulent velocities are only slightly greater than the settling velocities of the particles, the suspended-sediment concentration is not very great even near the bed and drops rapidly to essentially zero a short distance above the bed (Figure 2-34A). When the turbulent velocities are much greater than the settling velocities of the particles, however, the suspended-sediment concentration near the bed is high and falls off only slightly with height above the bed (Figure 2-34B). Laboratory measurements of vertical profiles of suspended sediment match the theoretical predictions very well.

Figure 2-34 Curves of suspended-sediment concentration. A: vertical turbulent velocities are only slightly greater than the fall velocities of the sediment particles. B: Vertical turbulent velocities are much greater than the fall velocities of the sediment particles.

2.4.3.7 The theory is only a qualified success, however, because the it provides only *the distribution relative to some assumed reference concentration at some reference elevation above the bed*. You can think of the bed-load layer as forming the bottom boundary condition for the suspended-sediment distribution. The key element—how the flow establishes the reference suspended-sediment concentration just outside the bed-load layer—is still not well known.
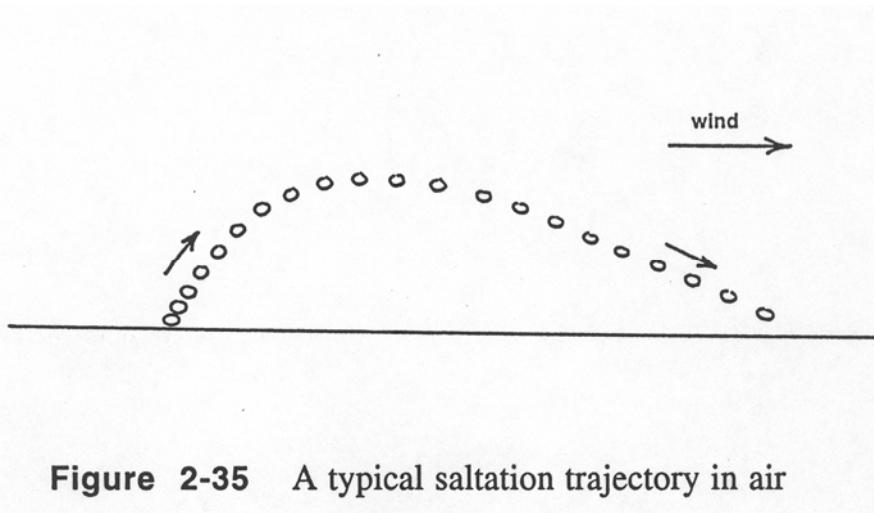
2.4.3.8 Although no measurements have been made, it seems reasonable to suppose that *in most sediment-transporting flow systems, like rivers or storm-driven currents in the shallow ocean, most of the sediment transported past a given cross section of the flow over a long period of time is in suspension*. Even though bed-load transport may be occurring most of the time, the total volume

moved as bed load is relatively small. On the other hand, *the imprint or signature of bed-load transport on the sedimentary record is very great.*

2.4.4 Saltation

2.4.4.1 All the material discussed so far in this section on the sediment load is relevant to the transport of sediment in water. The picture for transport of sediment by *wind* is rather different. The reason is that the ratio of sediment density to fluid density is far greater in air than in water, so *the sediment grains have great relative inertia and can travel through the air without their trajectories being much affected by the local fluid turbulence.*

2.4.4.2 The characteristic mode of sediment movement in air is saltation. **Saltation** is *a mode of particle movement in which particles are launched from the bed at moderate to steep angles, follow a regular arching trajectory, and descend back to the bed at small angles.* Figure 2-35 shows a typical saltation trajectory taken by a saltating grain of sand in air. Because of the great relative density of the sediment particles, saltation trajectories are rather regular, and show little of the sinuosity that might be expected from passage through turbulent eddies.
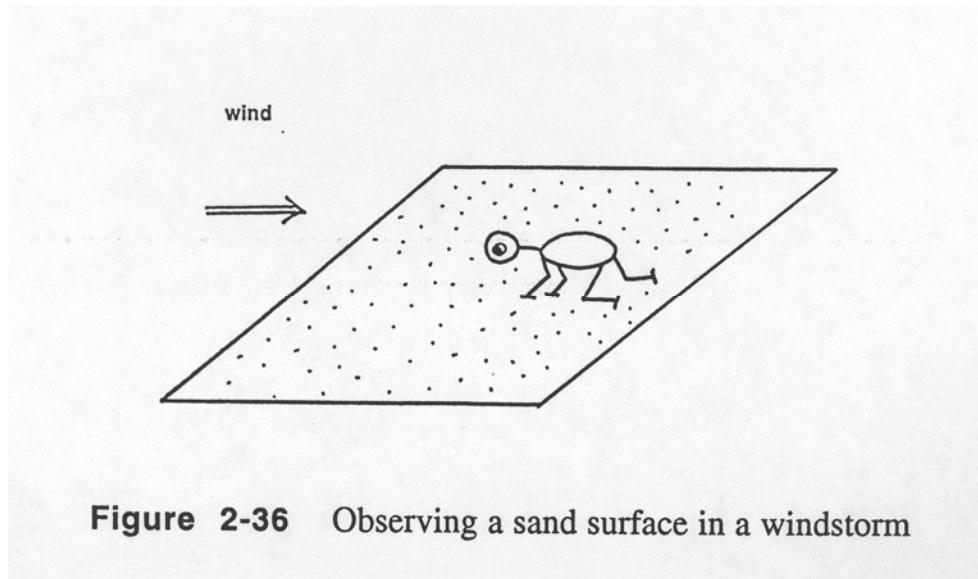

Figure 2-35   A typical saltation trajectory in air

2.4.4.3 Saltation is the dominant mode of particle movement when a strong wind blows over a sand surface. Except in the very strongest winds, the *saltation heights* attained by the saltating grains seldom exceeds a meter, and the *saltation lengths* are mostly less than a few meters. There is, of course, a continuous distribution of jump heights and jump lengths, from zero to the maximum. There's also wide variability in take-off angles: they range from just a few tens of degrees to vertical. After colliding with some particularly immovable bed grain, a saltating grain can even take off with an upstream component to its motion.

2.4.4.4 If you're lucky enough to be out on a dry sand surface during a strong wind, take the risk of getting some sand in your eyes and nose and mouth, and get down on the sand surface for a horizontal view at an eye level of a few tens of centimeters above the surface (Figure 2-36). You'd see a hazy layer of saltating
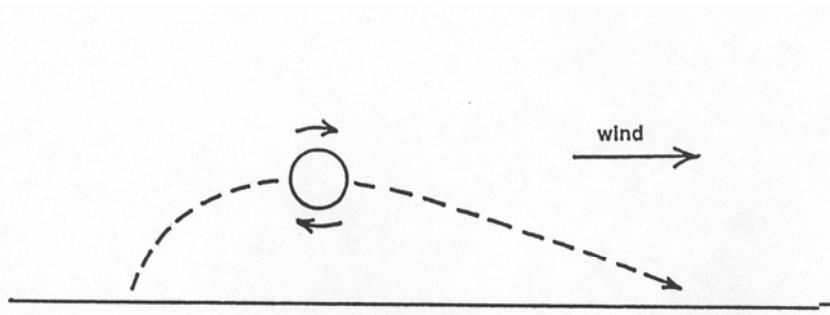
grains, which tails off imperceptibly upward. If you then looked downward at the sand surface, you'd see an abundance of surface grains being pushed along for short distances, just one or a few grain diameters at a time, presumably by being struck by saltating grains. That mode of movement is called **surface creep**—but keep in mind that there's really no sharp break between surface creep and saltation.

Figure 2-36   Observing a sand surface in a windstorm

2.4.4.5   Many aspects of saltation are not well understood. Among these is *the mechanism that causes the initial rise of the grains*. Two different effects come to mind: *lift forces*, as was discussed in an earlier section, and *the geometry of collisions between already saltating grains and stationary grains on the bed*. Another problem is that of the importance of saltation in water. Several researchers have described trajectories of sand grains moving in trajectories close to the bed in water flows as a kind of saltation. Such trajectories are not much like the characteristic saltation trajectories of dense grains in air, and whether what is described as saltation in water is related to true saltation in air is an unresolved question.

2.4.4.6   An interesting feature of saltation is the rotation rates of the saltating particles. Ultra-high-speed motion pictures have revealed *strikingly high spin rates of hundreds of revolutions per second!* The sense of spin is such that that the top of the particle moves faster in the downstream direction than the bottom of the particle (Figure 2-37). The origin of such high rates of spin is still a mystery.

Figure 2-37 Spin of a particle saltating in air

2.5 Transport Rates

2.5.1 The rate at which sediment is moved past a cross section of the flow, called either the **sediment transport rate** or the **sediment discharge**, usually denoted by Q_S, is one of the most important concepts on sediment transport mechanics. The sediment transport rate can be measured in *mass per unit time*, or in *weight per unit time*, or in *volume per unit time*. Often the sediment transport rate is considered per unit width of flow; then it's called the **unit sediment transport rate**, and is usually written q_S.

2.5.2 The concept of sediment transport rate is most naturally applied to unidirectional flows, but *even the most nearly symmetrical oscillatory flows usually have at least slight net transport in one direction or another*; this latter transport is the difference between the transport in one direction and the transport in the opposite direction during one oscillation, averaged over a large number of oscillations.

2.5.3 There are two different ways of approaching the characterization of sediment transport rates. On the one hand, you could take a purely empirical approach: make a list of all the variables that should affect the transport rate, and then make a dimensional analysis to provide a rational framework for handling or organizing measurement data. If enough good data are already at hand, you'd then have some predictive capability.

2.5.4 What's usually done, however, is something different: think about the physics of sediment transport in a way that allows you to develop the form of some rational equation for transport rates, which contains within it one or more "adjustable parameters" whose values are assigned by analysis of selected data sets already at hand.

2.5.5 Our common sense tells us, correctly again, that the stronger the flow the greater the sediment transport rate. And an important first-order fact of observation is that *the sediment transport rate is a very steeply increasing function of the flow strength*. Perhaps the simplest approach to quantifying q_S is to write an expression like

$$q_S = A\,\tau_o{}^n \tag{2.1}$$

where A is a coefficient and n is an exponent much larger than one. Better yet, τ_o might be replaced with $\tau_o - \tau_{oc}$, where τ_{oc} is the threshold boundary shear stress for sediment movement. Such an approach has no particular physical basis, but with A and n adjusted by use of observational data, Equation (2.1) can serve for very crude estimates of q_S. But the trouble with Equation (2.1) is that it has no strong physical basis. The true situation must be much more complicated than Equation (2.1).

2.5.6 It's almost no exaggeration to say that *dozens of sediment discharge formulas have found their way into the hydraulic-engineering literature*. And it's an understatement to say that *the physical basis behind these formulas ranges widely*. You might think that the ease with which data can be fitted to such a formula should give an idea of the validity of the underlying assumptions about the physics of sediment transport. But a skeptic might point out that an equation of almost any form can accommodate a given data set fairly well, so long as there are two or three adjustable parameters in that equation—as is usually the case with sediment discharge formulas. In the context of this course, there's no point in pursuing formulas for sediment transport rate any further, despite the fact that they're so important in engineering approaches to sediment transport.

2.5.7 *Measuring the sediment transport rate is a notoriously difficult task*. Ideally one would like to be able to keep track of every single particle that passes across a given cross section of the flow in some representative time interval, and measure the mass of each of those particles (somehow without disrupting the flow), and then divide the total mass by the time interval during which it was measured. Unfortunately there's no good way of doing that.

2.5.8 In the laboratory it's often possible (although difficult) to arrange the flow so that all the sediment is extracted at some point, for example by means of a slot trap in the bottom of the channel. Then the catch can be weighed and put right back into the flow immediately. In a few cases this been done even on small rivers, for research purposes.

2.5.9 Usually bed load is measured with devices called *bed-load traps* or *bed-load samplers*. They come in all sizes and geometries, but their purpose is to catch all the bed load that comes upon the trap from upstream—without catching too much or too little. Unfortunately there are many practical problems connected with bed-load traps. Suspended load is easier to measure, by taking water samples at a number of points along the vertical and making a numerical integration.

2.6 Mixed-Size Sediments
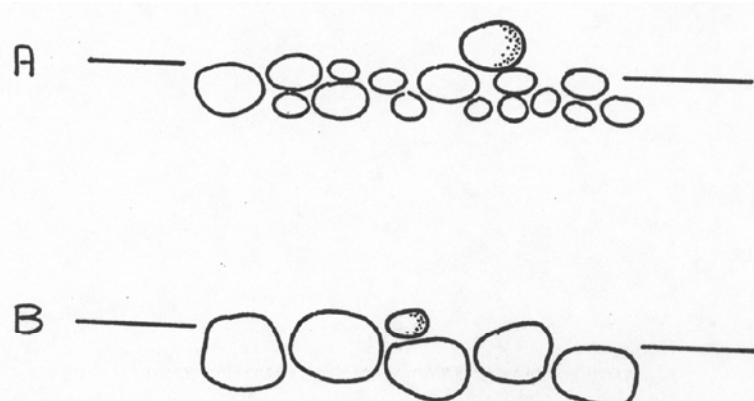
2.6.1 Until recently, almost all research work and development of theory on sediment transport has ignored the effect of sorting (the spread of sizes around the mean size). For the purposes of much engineering work, the assumption that sorting is unimportant has not caused great difficulties, partly because the particle size itself is not an important consideration and partly because the uncertainties in theory and measurement overshadow the errors arising from neglect of sorting. But in sedimentology one of the most important aspects of sediment transport is *the tendency for fractionation or spatial segregation of different particle sizes*: the

particle-size distribution of a clastic sediment, which is one of the few available reflections of the paleoflow environment, is in large part the outcome of fractionation arising from differing rates of transport of different particle sizes.
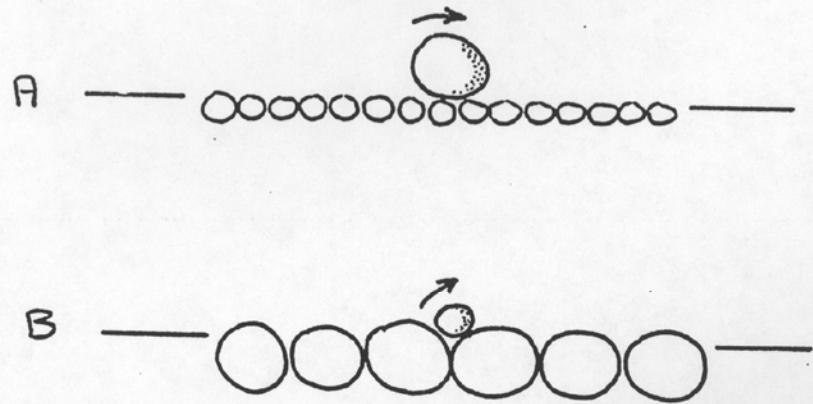
2.6.2 Understanding of the phenomena of mixed-size sediment transport is still rudimentary. Much more work needs to be done before that understanding can provide a good basis for interpreting the information contained in the size distributions of clastic sediments. In this section we'll only point out some of the important aspects of the problem.

2.6.3 Let us pose an important question for you. Think carefully before answering. Consider a planar bed of poorly sorted sediment, with sizes ranging perhaps from sand size to large pebble size. A strong current is passed over the sediment bed, causing bed-load transport. Think about the ***fractional transport rates***—that is, *the transport rates of the individual size fractions in the sediment mixture*. The question is: if you normalize those fractional transport rates by dividing the transport rate of each fraction by the percentage of that fraction in the total mixture (to take into account the differing abundances of the fractions), which fractions would have the highest normalized transport rates: the *finest sediment*, the *coarsest sediment*, or *some fraction or fractions in between*?

2.6.4 This is a situation where you have to be careful about what your intuition or common sense tells you. It's easy to adduce three important effects here: *particle weight*, *exposure/shielding*, and *rollability*. Even though the large particles are much heavier than the small particles, and therefore are more difficult to move, they project much higher into the flow, and therefore the fluid forces on them are much greater (Figure 2-38). But it's easy for a large particle to roll over a bed of smaller particles (Figure 2-39). On the other hand, the fine particles are much lighter but they tend to be sheltered from the force of the flow by the larger particles. But a smaller particle has a hard time rolling up and over a bed of larger particles to be set into motion.

Figure 2-38 Exposure/shielding effect in mixed-size sediment transport. A: Larger particles tend to be exposed to the force of the flow. B. Smaller particles tend to be shielded from the force of the flow.

Figure 2-39 Rollability effect in mixed-size sediment transport. A: Larger particles roll easily on a bed of small particles. B: Smaller particles roll with difficulty on a bed of larger particles.

2.6.5 Interestingly, available laboratory data show that *for bed-load transport all of these effects tend quite closely to cancel or offset one another, so that the normalized transport rates of all size fractions are not greatly different*. However, the small differences that do exist may be important when transport takes place over long time scales and for long distances. When some of the sediment is carried in suspension, it can easily outrun the fractions traveling as bed load, and then there can be large differences in normalized fractional transport rates.

2.7 More Concentrated Flows

2.7.1 Dispersed-Sediment Flows and Concentrateed-Sediment Flows

2.7.1.1 When viewed from a dynamical standpoint, *there's a continuous spectrum of concentrations of the sediment load, from vanishingly small, in flows which are only a little stronger than the threshold condition, to extremely high*. The upper limit of sediment concentration is the condition that there's enough pore space between the sediment particles so that the sediment-fluid mixture can be deformed without the interlocking of grains preventing the deformation. Near this upper limit, sediment concentrations can be as much as 60% by volume.

2.7.1.2 But from the standpoint of practical sedimentology, we've found it useful to *think in terms of two different ranges of sediment concentrations*. On the one hand, we commonly deal with *deposits that were made by flows that carried relatively small concentrations of sediment*. Concentrations in these flows are usually very small, but even though the concentrations might be of the order of ten percent, the flows can still be viewed as relatively "thin" and certainly fully turbulent.   All pressure-gradient-driven marine currents, like tidal currents or wind-driven currents, as well as most river flows, are in this category. we'll call such flows **dispersed-sediment flows**. You'll see that these dispersed-sediment flows are responsible for distinctive kinds of deposits which show generally good

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stratification and evidence of tractional transport. They tend to be relatively well sorted.

2.7.1.3 On the other hand, *many flows that are of great sedimentological interest have sediment concentrations near the upper end of the possible range*. Concentrated sediment gravity flows like debris flows are the main kinds of flows with sediment concentrations this high, although a few rivers (like the Yellow River in China) with what's called *hyperconcentrated flow* also have extremely high sediment concentrations. we'll call flows of this kind **concentrated-sediment flows**. The deposits produced by these flows tend to lack clear stratification, and they're usually much less well sorted than dispersed-sediment flows. Obviously there are intermediate cases, not only in rivers but also in sediment gravity flows.

2.7.1.4 These two terms are *very unofficial*, and you will not find them in the sedimentological literature. But we think they're a useful "crutch" for thinking about the origin of clastic deposits when you're on the outcrop. When I'm looking at a particular bed, we usually ask ourselves whether that bed was deposited by a dispersed-sediment flow or a concentrated-sediment flow, or perhaps by a flow with intermediate characteristics.

2.7.1.5 To our mind, the main reason why concentrated-sediment flows are more difficult to come to grips with than dispersed-sediment flows has to do with the related problems of *overloading* and *past history*. Suppose you start with a flow of water, with a given discharge, over a sediment bed. If the flow is strong enough, some of the sediment will be entrained and transported, either as bed load or as a combination of bed load and suspended load. (If the sediment is entirely in the clay to fine silt range, the flow must be so strong to even begin to erode the bed, and the resulting entrained sediment has such a small settling velocity, that all of the load might be suspended load.) But in any case an equilibrium becomes established whereby the flow transports just as much sediment as it is able to, and there's a balance between entrainment and redeposition. You could look at an established flow like that and not be able to say anything about how or when it got started.

2.7.1.6 But the range of such flows by no means exhausts all of the possible sediment-transporting flows. That's because *it's possible for a preexisting non-flowing mass of sediment and water to undergo mass mobilization of some kind, by liquefaction, for example, and then flow as a high-concentration two-phase flow with a concentration much higher than could have become established by starting with a clear-water flow over a bed of the same sediment*. A flow like this has an important aspect of past history. And it's likely to be overloaded—although, depending upon the particular conditions of flow, it may be self-sustaining until it eventually reaches a gentler slope down which to flow. Of course, any flow, not just mass flows of this kind, will deposit its sediment as it becomes overloaded by decelerating. It's just that *the degree of overloading, or the rate at which the flow loses sediment as it decelerates, is typically much greater and more spectacular with concentrated-sediment flows than with dispersed-sediment flows*.

2.7.2 Classification and Discussion of Concentrated-Sediment Flows

2.7.2.1 At the risk of oversimplifying complex phenomena, here's a very crude and very generalized classification of concentrated-sediment flows (Table 2-1). The orientation here is practical rather than theoretical.

	concentration	vertical differentiation	vertical mixing	turbulence	deposition
slumps/ slides	high	slight- moderate	slight- moderate	(none)	sudden
debris flows	high	slight	slight- moderate	none- some	sudden to incremental
hyperconcentrated flows	moderate- high	slight- moderate	moderate	weak- strong	incremental
turbidity currents	low- moderate	moderate- strong	strong	moderate- strong	incremental
grain flows	high	slight- moderate	slight- moderate	(none)	sudden to incremental

Table 2-1   Crude and generalized classification of concentrated sediment flows

slumps/slides: large masses of solid material move downslope under the pull of gravity; presence of interstitial fluid may be important but is not essential for the existence of the flow; not usually viewed as "sediment transport". Vertical mixing ranges from slight to substantial; deposition is probably usually sudden, by "freezing" of the moving mass.

debris flows: highly concentrated mixtures of sediment and water with little or no differentiation that flow as thick two-phase fluids; little or no vertical differentiation in the flowing mass; vertical mixing is usually substantial; nonturbulent to turbulent; deposition ranges from sudden to incremental.

hyperconcentrated flows: moderately concentrated mixtures of sediment and water, with little to perhaps moderate differentiation; vertical mixing is substantial; weakly to strongly turbulent; deposition is incremental.

turbidity currents: slightly to moderately concentrated mixtures of sediment and water; moderate to great vertical differentiation in the flowing mass; vertical mixing substantial; moderately to strongly turbulent; deposition incremental.

grain flows: downslope flows of solid particles under gravity; importance of interstitial fluid negligible to slight; deposition ranges from sudden to incremental.

2.7.2.2 With the exception of the last item, grain flows, the foregoing list is approximately in order of decreasing solids concentration, increasing turbulence and mixing, and a shift from sudden to incremental deposition. We emphasize that *there is great variability within each of the categories*, and certainly *gradations in both physics and features among categories*. In a way, grain flows are like slumps/slides (or should we make that vice versa?), but in scale and origin they seem to us rather different.

2.7.2.3 Debris flows and turbidity currents are of undoubtedly great sedimentological importance. The importance of hyperconcentrated flows is less clear but is certainly nonnegligible, and maybe much greater than most workers suspect. In regard to grain flows, the only grain flows we've ever been convinced are sedimentologically important are on very small and very local scales, mainly those that transport sand down the lee faces of dunes in both subaerial and subaqueous environments.