2.4 PARTICLE TRANSPORT BY SEDIMENT GRAVITY FLOWS

In the preceding section, we examined sediment transport resulting from the interaction of moving fluids and sediment. During fluid-flow transport, the fluids (water, wind, ice) move in various ways under the action of gravity and the sediment is simply carried along with and by the fluid.

Sediment can also be transported essentially independently of fluid by the effect of gravity acting directly on the sediment. In this type of transport, fluids may play a role in reducing internal friction and supporting grains, but they are not primarily responsible for downslope movement of the sediment. Movement of sediment under the influence of gravity creates the flow, and flow stops when the sediment load is deposited.

Sediment can be transported by the direct action of gravity in both subaerial and subaqueous environments. Gravity transport under submarine conditions has the greater sedimentological significance. A spectrum of gravity movements exists, ranging from those in which sediment is moved en masse and fluids act mainly to reduce internal friction by lubricating the grains, to those in which transport is on a grain-by-grain basis and fluids play an important role in supporting the sediment during transport. Gravity mass movements can be grouped into rock falls, slides, and sediment gravity flows, as shown in Table 2.1. **Rock fall** involves free fall of blocks or clasts from cliffs or steep slopes. **Slides** are en-masse movements of rock or sediment owing to shear failure that take place with little accompanying internal deformation of the mass. **Sediment gravity flows** are more "fluid" types of movement in which breakdown in grain packing occurs and internal deformation of the sediment mass is intense.

Sediment gravity flows are of particular interest because they are capable of rapidly transporting large quantities of sediment, including very coarse sediment, even into very deep water in the oceans. Gravity flows that occur in subaerial environments can be considered, in a broad sense, to include snow avalanches, pyroclastic flows and base surges resulting from volcanic eruptions, grain flow of dry sand down the slip face of sand dunes, and both volcanic and nonvolcanic debris flows and mud flows, in which large particles are transported in a slurrylike matrix of finer material. Subaqueous sediment gravity flows also include grain flows and debris flows, as well as turbidity currents and liquefied sediment flows.

TABLE 2.1 Major types of mass-transport processes, their mechanical behavior, and transport and sediment support mechanisms

Mass transport processes			Mechanical behavior	Transport mechanism and sediment support
Rock fall			Elastic	Free fall and subordinate rolling of individual blocks or clasts along steep slopes
Slide		Glide		Shear failure among discrete shear planes with little internal deformation or rotation
		Slump		Shear failure accompanied by rotation along discrete shear surfaces with little internal deformation
			Plastic limit	
Sediment gravity flow	Mass flow	Debris flow Mud flow	Plastic	Shear distributed throughout sediment mass; strength principally from cohesion due to clay content; additional matrix support possibly from buoyancy
		Grain flows. Inertial Viscous	Liquid limit	Cohesionless sediment supported by dispersive pressure; flow in inertial (high-concentration) or viscous (low-concentration) regime; steep slopes usually required
	Fluidal flow	Liquefied flow	Viscous fluid	Cohesionless sediment supported by upward displacement of fluid (dilatance) as loosely packed structure collapses, settling into a more tightly packed framework; slopes in excess of 3° required
		Fluidized flow		Cohesionless sediment supported by the forced upward motion of escaping pore fluid; thin (<10 cm) and short-lived
		Turbidity current		Supported by fluid turbulence

Source: Nardin, T. R., F. J. Hein, D. S. Gorsline, and B. D. Edwards, 1979. A review of mass movement processes, sediments, and acoustical characteristics, and contrasts in slope and base-of-slope systems versus canyon-fan-basin flow systems, in L. J. Doyle and O. R. Pilkey (eds.), Geology of continental slopes: *Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 27*. Table 1, p. 64, reprinted by permission of SEPM, Tulsa, OK.

Sediment gravity flows can occur only when grains become separated and dispersed to the point that internal friction and cohesiveness are sufficiently reduced to lower the strength of the sediment mass below the critical point required for gravity to initiate movement. Four theoretical types of dispersive and support flow mechanisms that can achieve this reduction in internal strength have been identified: turbulent flow (turbulence), upward escape of intergranular fluid, grain interaction (dispersive pressure), and support by a cohesive matrix (Fig. 2.6). Four observed flow types can be identified that correspond to these theoretical support mechanisms: turbidity currents, liquefied flow, grain flow, and mud and debris flow. These four mechanisms of gravity transport, described in greater detail below, can be thought of as members of a spectrum of sediment-gravity-flow processes. One type may grade into another under some conditions, as when a submarine mud flow changes into a turbidity current downslope with additional mixing and dilution by water.

Observed Type of flow	Turbidity Current	Liquefied Flow	Grain Flow	Mud Flow; Debris Flow
Support Mechanism	Turbulent Fluid	Upward escape of intergranular fluid	Grain interaction (dispersive pressure)	Strength of matrix
Type of Fluid	Turbulent fluid	Newtonian fluid (high viscosity)	Non-Newtonian fluid	Bingham plastic

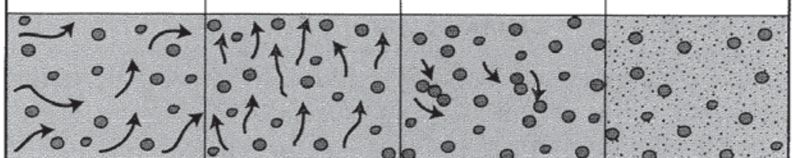

FIGURE 2.6 The principal kinds of sediment-gravity flows and the relationship of flow type to grain-support mechanisms and fluid types. [Based on Middleton, G. V., and M. A. Hampton, 1976, Subaqueous sediment transport and deposition by sediment gravity flows, in Stanley, D. H., and D. J. P. Swift (eds.), *Marine sediment transport and environmental management:* John Wiley & Sons, Inc., Fig. 1, p. 198.]

Turbidity Currents

FLOW MECHANISMS AND CHARACTERISTICS A turbidity current is a kind of density current that flows downslope along the bottom of an ocean or lake because of density contrasts with the surrounding (ambient) water arising from sediment that becomes suspended in the water owing to turbulence. Turbidity currents can be generated experimentally in the laboratory by the sudden release of muddy, dense water into the end of a sloping flume filled with less dense, clear water. They have been observed to occur under natural conditions in lakes where muddy river water enters the lakes, and they are believed to have occurred throughout geologic time in the marine environment on continental margins. In this setting, they originate particularly in or near the heads of submarine canyons.

Turbidity currents can be generated by a variety of mechanisms, including sediment failure, storm-triggered flow of sand and mud into canyon heads, bedload inflow from rivers and glacial meltwater, and flows during eruption of airfall ash (Normark and Piper, 1991). They may move as surges or as steady, uniform flows.

Surges, or spasmodic turbidity currents, are initiated by some short-lived catastrophic event, such as earthquake-triggered massive sediment slumping or storm waves acting on a continental shelf. Such an event creates intense turbulence in the water overlying the seafloor, resulting in extensive erosion and entrainment of sediment, which is rapidly thrown into suspension. The sediment then remains suspended, supported in the water column by turbulence. This process generates a dense, turbid cloud that moves downslope, eroding and picking up more sediment as it increases in speed. Surge flows develop into three main parts as they move away from the source: the head, body, and tail.

Some turbidity currents are steady, uniform flows that lack a turbulent head. These flows move at velocities similar to those of the body of surge-type flows. Although the velocity is sensitive to the slope over which flow takes place, flow may occur on slopes as low as 1 degree (Kersey and Hsü, 1976). Steady, uniform flows have been observed along the sloping bottom of lakes where sediment-laden rivers run into the lakes. They may occur also on continental shelves

BOX 2.4
Flow Velocity of Turbidity Currents

The **head** of the surge is about twice as thick as the rest of the flow and is characterized by intense turbulence. The velocity, U_{head} at which the head advances into still water is:

$$U_{head} = 0.7\sqrt{\frac{\Delta\rho}{\rho}gh} \qquad \textbf{2.4.1}$$

where $\Delta\rho$ is the density contrast between the turbidity current and the ambient water, ρ is the density of the ambient water, g is gravitational acceleration, and h is the height of the head (Middleton and Hampton, 1976). The head is overhanging and is divided transversely into lobes and clefts (Fig. 2.4.1 and 2.4.2).

Flow within the **body** of a surge-type turbidity current is nearly steady and uniform and the flow is almost uniform in thickness. The body moves at a velocity, U_{body}, that is:

$$U_{body} = \sqrt{\frac{8g}{f_0 + f_1}\left(\frac{\Delta\rho}{\rho}\right)hs} \qquad \textbf{2.4.2}$$

where h is the height or thickness of the flow, s is the slope of the bottom, f_0 is the frictional resistance at the bottom of the flow, and f_1 is the frictional resistance at the upper interface of the flow in contact with the overlying ambient water layer. The body flows at a velocity that is faster in deep water than that of the head. This difference in velocity causes the forward part of the body to consume itself within the head in the process of mixing with the ambient water (Allen, 1985). The **tail** of the flow thins abruptly away from the body and becomes more dilute.

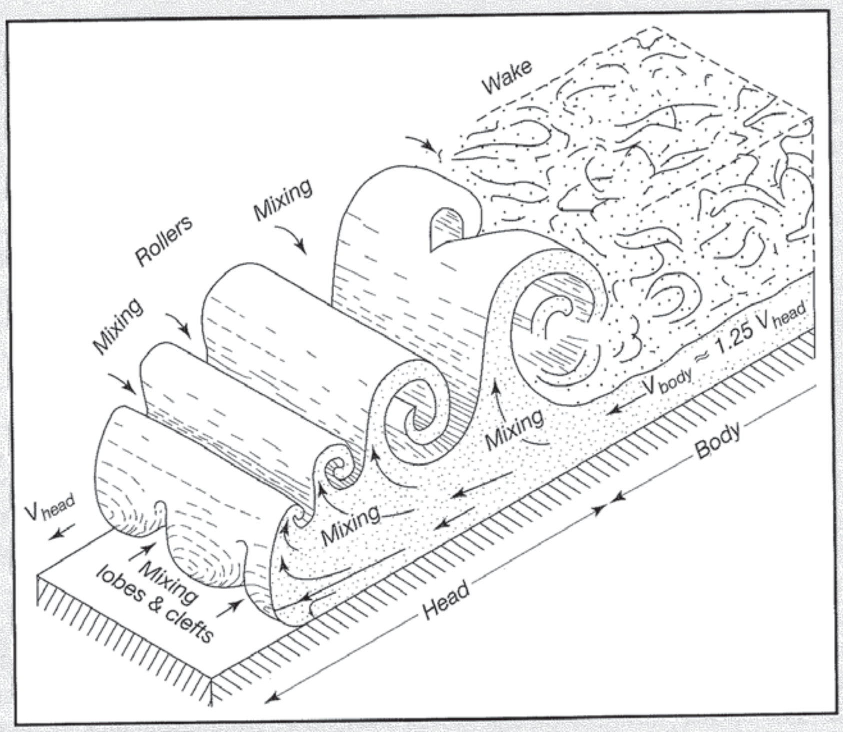

FIGURE 2.4.1 Postulated structure of the head and body of a turbidity current advancing into deep water. The tail is not shown. [From Allen, J. R. L., 1985, Loose boundary hydraulics and fluid mechanics: Selected advances since 1961, in Brenchley, P., and D. J. P. Swift (eds.), *Sedimentology: Recent developments and applied aspects.* Fig. 8, p. 20, reprinted by permission of Blackwell Scientific Publications, Ltd., Oxford.]

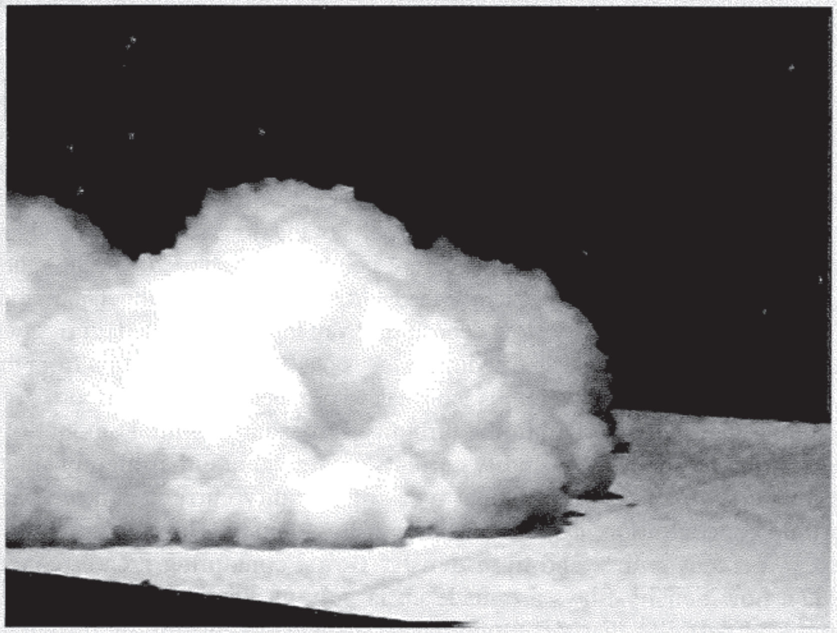

FIGURE 2.4.2 The head of an experimentally-generated turbidity current advancing across the floor of a small flume. Note the lobes and clefts created by extreme turbulence in the head. Modified from "Gravity Deposits" (video), Institut Français du Petrole. Reproduced by permission.

where muddy rivers discharge; however, they are less likely in this setting because the density contrast between muddy river water and ocean water is less than that between muddy river water and freshwater.

Once sediment is suspended in a turbidity current, the turbidity current continues to flow for some time under the action of gravity and inertia. Flow will stop when the sediment-water mixture that produces the density contrast with the ambient water is exhausted by settling of the suspended load. Rapid deposition of coarser particles from suspension appears to occur in regions near the source owing to early decay of the extremely intense turbulence generated by the initial event. As the flow continues to move forward, the remaining coarser material will be progressively concentrated in the head of the flow; denser fluid must be continuously supplied to the head to replace that lost to eddies that break off from the head and rejoin the body of the flow. Owing to differences in turbulence in the head and body, the head may be a region of erosion while deposition is taking place from the body.

Theoretically, sediment remaining in suspension after initial deposition of coarse material in the proximal area can, during further transport, be maintained in suspension for a very long time in a state of dynamic equilibrium called **autosuspension** (Bagnold, 1962; Pantin, 1979; Parker, 1982). A condition of autosuspension is presumably maintained because turbulence continues to be generated in the bottom of the flow owing to gravity-generated downslope flow of the turbidity current over the bed. Thus, loss of energy by friction of the flow with the bottom is compensated for by gravitational energy. The distance that turbidity currents can travel in the ocean is not known from unequivocal evidence. A presumed turbidity current triggered by the 1929 Grand Banks earthquake off Nova Scotia appears to have traveled south across the floor of the Atlantic for a distance of more than 300 km at velocities up to 67 km/hr (19 m/s), as timed by breaks in submarine telegraph cables (Piper et al., 1988). Transport of sediment over this distance suggests that autosuspension may actually work; nonetheless, some geologists remain skeptical of the autosuspension process (e.g., review by Middleton, 1993).

The velocity of a turbidity current eventually diminishes owing to flattening of the canyon slope, overbank flow of the current along a submarine channel, or spreading of the flow over the flat ocean floor at the base of the slope. As the flow slows, turbulence generated along the sole of the flow also diminishes, and the current gradually becomes more dilute owing to mixing with ambient water around the head and along the upper interface. The remaining sediment carried in the head eventually settles out, causing the head to sink and dissipate. The exact process by which deposition takes place from various parts of a turbidity current is still not thoroughly understood, although it seems clear from experimental results that deposition does not occur in all parts of the current at the same time. As mentioned above, for example, the head may be a region of potential erosion at the same time that the body behind the head is depositing sediment. Sediment that is deposited very rapidly from some parts of the flow, such as the head, may undergo little or no subsequent traction transport before being quickly buried. On the other hand, in more distal parts of the flow or in areas where the head overflows the channel, a period of scouring by the head may be followed by slow deposition from the body and tail, during which additional traction transport of the deposited sediment takes place. Final deposition from the tail may take place after movement of the current is too weak to produce traction transport.

Depending upon position within the turbidity flow and the initial amount of sediment put into suspension by the flow, turbidity currents may contain either high or relatively low concentrations of sediment. Two principal types of turbidity currents, on the basis of suspended particle concentration, can be considered: **low-density flows**, containing less than about 20 to 30 percent grains, and **high-density flows**, containing greater concentrations (Lowe, 1982). Low-density flows are made up largely of clay, silt, and fine- to medium-grained sand-size particles that are supported in suspension entirely by turbulence. High-density flows may include coarse-grained sands and pebble- to cobble-size clasts as well as fine sediment. Support of coarse particles during flow is provided by turbulence aided by hindered settling resulting from their own high sediment concentrations and the buoyant lift provided by the interstitial mixture of water and fine sediment. (High-density flows differ from debris flows in that debris flows are not turbulent and are less fluid.) Note that the heads of turbidity currents may be high-density flows, whereas the tails may be dilute, low-density flows.

Turbidity Current Deposits

The deposits of turbidity currents, commonly called **turbidites**, are of two basic types. Turbidites deposited from high-density flows with high sediment concentrations tend to form thick-bedded turbidite successions containing coarse-grained sandstones or gravels. Individual flow units typically have relatively poor grading and few internal laminations, and basal scour marks are either poorly developed or absent. Some turbidites with thick, coarse-grained basal units may grade upward to finer grained deposits that display traction structures such as laminations and small-scale cross bedding (Fig. 2.7). In the uppermost part of the flow units, the sediments may consist of very fine grained, nearly homogeneous muds deposited from the tail of the flow. The deposits of more dilute, low-concentration turbidity current flows generally form thin-bedded turbidite successions. Individual flow units are fine grained at the base and display good vertical size grading, well-developed laminations, and small-scale cross-bedding. Scour marks may be present on the soles or bottoms of the beds.

Bouma (1962) proposed an ideal turbidite sequence, now commonly called a **Bouma sequence**. This ideal sequence consists of five structural units (Fig. 2.8-1) that include the characteristics of both types of turbidites. These structural subdivisions presumably record the decay of flow strength of a turbidity current with time and the progressive development of different sedimentary structures and bedforms in adjustment to different flow regimes (upper-flow regime to lower-flow regime) as current-flow velocity wanes. Most turbidites do not contain all of these structural units. Thick, coarse-grained turbidites tend to show well-developed A and B units, but C through E units are commonly poorly developed or absent. Thin, finer grained turbidites may display well-developed C through E units and poorly developed or absent A and B units. In fact, Hsü (1989, p. 116) claims that Bouma's D unit rarely occurs and that most turbidites can be divided into only two units: a lower horizontally laminated unit (unit A + B, Fig. 2.8-2 and -3) and an upper, cross-laminated unit (unit C). Unit E is a problem because it may consist of fine material deposited slowly from the water column, and thus it may not be part of a turbidite flow unit.

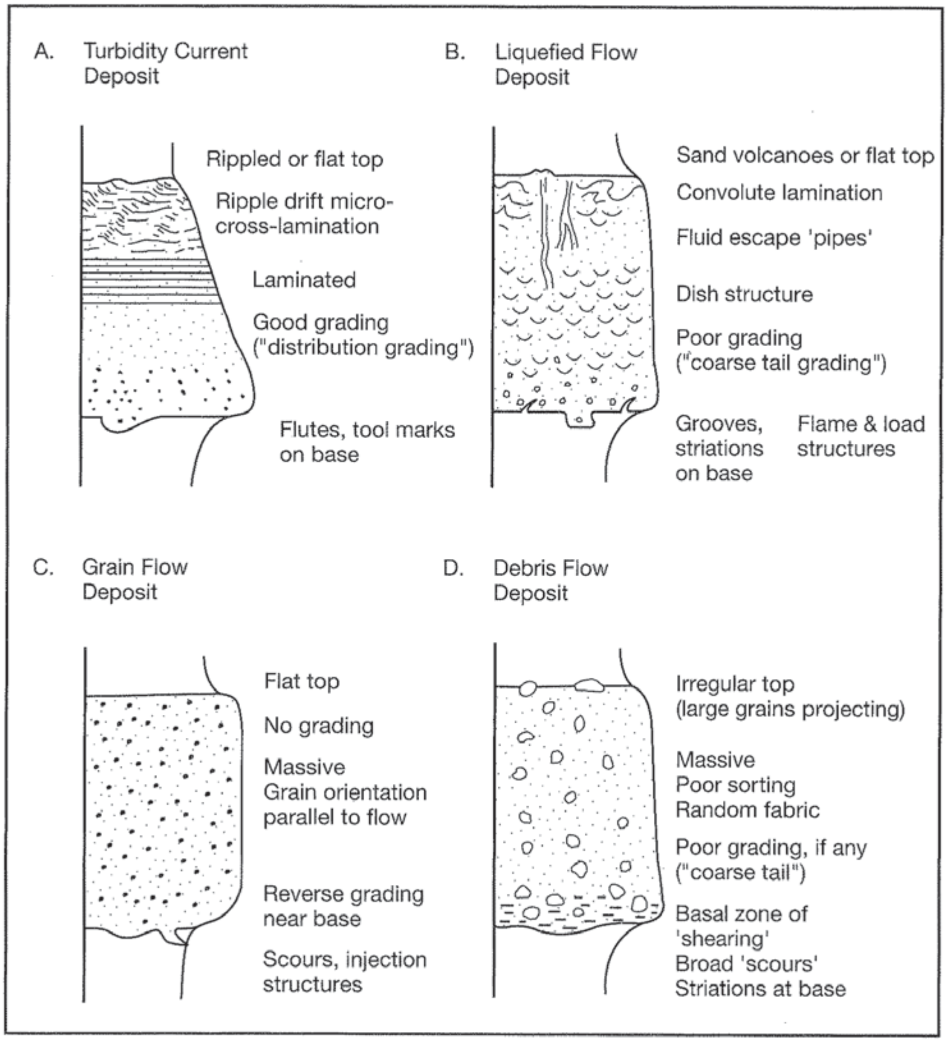

FIGURE 2.7 Comparison of sedimentary structures in different types of sediment gravity-flow deposits. [After Middleton, G. V., and M. A. Hampton, 1976, Subaqueous sediment transport and deposition by sediment gravity flows, in Stanley, D. H., and D. J. P. Swift (eds.), *Marine sediment transport and environmental management:* John Wiley & Sons, Inc., Fig. 9, p. 213. Reproduced by permission.]

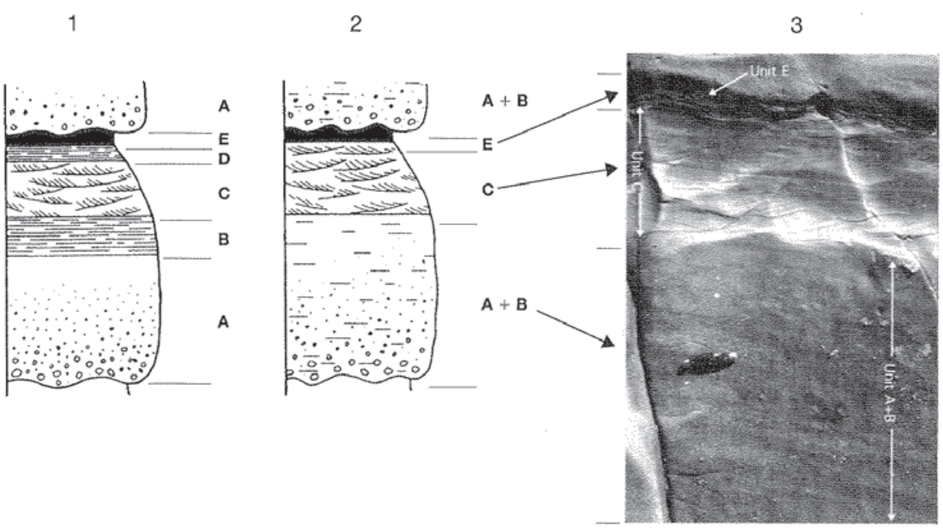

FIGURE 2.8 Ideal sequence of sedimentary structures in graded-bed units as proposed by Bouma (1) and Hsü (2). Note that in Hsü's model, Bouma units A and B are combined and unit D is omitted. (3) Photograph of a Bouma unit that is very similar to Hsü's model (Cretaceous, southern Oregon coast). [1 and 2 after Hsü, K.J., 1989, *Physical principles of sedimentology*, Fig. 7.8, p. 116, reprinted by permission of Springer-Verlag, Berlin.]

Turbidites laid down near the source, particularly within the main transport channel where suspended sediment concentrations are high, are generally the coarse-grained, massive, or poorly laminated type. Some very high concentration flows may also deposit coarse-grained turbidites within the main channel at considerable distances from the source. On the other hand, thin, fine-grained turbidites can also be deposited near the source where turbidity currents overflow the banks of a channel and become more dilute as they spread out over the seafloor, as well as in areas more distant from the source. The deposits of a single turbidity current flow typically display horizontal size grading in addition to vertical size grading. That is, thick, coarse-grained

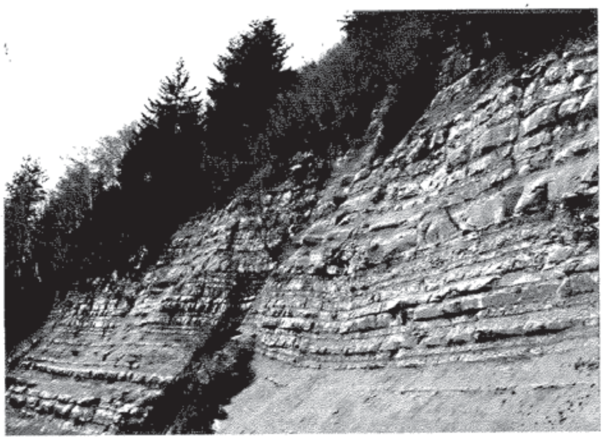

FIGURE 2.9 Rhythmically bedded turbidites, Tyee Formation (Eocene), southern Oregon coast.

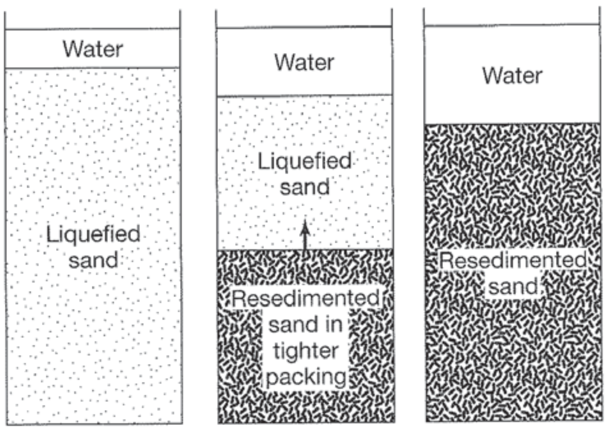

FIGURE 2.10 Schematic representation of grain settling and water expulsion during deposition of sand from a liquefied flow. [After Allen, J. R. L., and N. L. Banks, 1972, An interpretation and analysis of recumbent-folded deformed cross bedding: *Sedimentology*, v. 19, Fig. 3, p. 267, reprinted by permission of Elsevier Science Publishers, Amsterdam.]

deposits generally grade laterally to thinner and finer grained sediments. Repeated occurrence of turbidity-current flows produces a rythmic succession of turbidite beds (Bouma sequences) that may be hundreds of meters thick (e.g., Fig. 2.9). For a more comprehensive treatment of turbidites, see Mutti (1992) and Bouma and Stone (2000).

Liquefied Flows

FLOW MECHANISMS Liquefied flows are concentrated dispersions of grains in which the sediment is supported either by the upward flow of pore water escaping from between the grains as they settle downward by gravity or by pore water that is forced upward by injection from below. Loosely packed, cohesionless sediment such as sand can become temporarily liquefied owing to a sudden shock, or series of shocks, that causes the grains to momentarily lose contact with each other and become suspended in their own pore fluid. Grain contact may also be lost if a fluid is introduced into the base of a mass or column of cohesionless sediment and injection is continued until the grains are pushed apart, with their weight being supported by the rising fluid. This process is called fluidization. Once the cohesionless sediment has become liquefied (or fluidized), it loses its strength and behaves like a high-viscosity fluid that can, nonetheless, flow quite rapidly down slopes as low as 3°.

Liquefied flow can occur only as long as grain dispersion is maintained. As soon as the grains settle out of the fluid and reestablish grain-to-grain contact, the flowing layer will "freeze up" and stop moving. "Freezing" begins at the base of the flow; a surface of settled grains rises up through the dispersion at a rate determined by the settling velocity of the particles (Fig. 2.10). The time required for settling to occur is on the order of hours for thick, fine-grained flows (Lowe, 1976); therefore, liquefied flows may travel short, though potentially important, distances before deposition. The upward movement of pore waters through the settling grains as deposition occurs leads to formation of a number of fluid escape structures, such as dish structures. Some liquefied flows may become turbulent as the flowing sediment mass is accelerated downslope and thus change into turbidity currents.

Liquidized-Flow Deposits The deposits of liquefied flows are typically thick, poorly sorted sand units. They are characterized particularly by fluid escape structures (Chapter 4), such as the dish structures, pipes, and sand volcanoes shown in Figure 2.7B.

Grain Flows

FLOW MECHANISM Grain flows are dispersions of cohesionless sediment in which the sediment is supported in air by dispersive pressures owing to direct grain-to-grain collisions and in water by collisions and close approaches. Sediment can flow rapidly under both subaerial and subaqueous conditions, especially on steep slopes that approach the angle of repose for the sediment. Grain flow results in the movement of cohesionless sediments down steep slopes owing to sudden loss of internal shear strength of the sediment. Grain flow begins when traction processes cause cohesionless sediment, commonly sand, to be piled up beyond the critical angle of repose. This angle is a function of grain packing and grain shape and tends to be greatest in deposits with angular grains of low sphericity. When the angle of repose for a particular sediment is exceeded, avalanching occurs; flow quickly begins when the internal shear stresses owing to gravity exceed the internal shear strength of the sediment. The **dispersive pressures** needed to force the grains apart and keep them suspended during flow are provided not by fluid but by grain-to-grain collisions in air and grain collisions and close encounters in water as the failed mass of sediments moves down a slope. During the interaction of grains, dispersive pressure is the force normal to the plane of shearing which tends to expand or disperse the grains in that direction. Bagnold (1956) suggested that the relation between the shear stress (T) acting on grains and the dispersive pressure (P) is

$$T/P = \tan a \qquad \textbf{2.11}$$

where a is the angle of internal friction. This formula suggests that the minimum slope on which sustained grain flow in air is possible is about 30°; under water, greater slopes may be required for flow to occur. Although dispersion or dilation of sand grains is achieved and maintained during flow primarily by grain collisions, dispersion may be aided under some conditions by upward flow of pore fluids as grains settle or possibly by buoyancy of a dense mud matrix. Grain flow is similar to liquefied flow in many respects and may, in fact, grade into these flows. In contrast to liquefied flows, grain flow can occur under subaerial conditions as well as subaqueous conditions.

Grain flow is common on the lee slopes of sand dunes. Flows of cohesionless sand have also been observed and photographed in the ocean as they moved down steep slopes in submarine canyons (Shepard, 1961; Dill, 1966; Shepard and Dill, 1966). Grain flows over the floors of Norwegian fjords are reported to be responsible for breaking submarine telephone cables. Grain flows may be of limited geological significance because of the steep slopes required to initiate flow, although it has been suggested that grain flow may accompany turbidity currents on less steep slopes, moving beneath but independently of the turbidity currents. Deposition of grain-flow sediment occurs quickly and en masse by sudden "freezing," primarily as a result of reduction of slope angle.

GRAIN-FLOW DEPOSITS Grain-flow origin has been suggested by some workers for very thick, almost massive sandstone beds; however, Lowe (1976) concluded that the deposits of a single grain flow in any environment cannot be thicker than a few centimeters for sand-size grains. Reverse grading (that is, grading from fine size to coarse size upward, which occurs in some sandstones) has been attributed to grain-flow processes. Reverse grading is assumed to form during grain flow as a result of smaller particles filtering down through larger particles while they are in the dispersed state, a process called **kinetic sieving**.

Grain-flow deposits are massively bedded with little or no internal laminations and grading except possible reverse grading in the base (Fig. 2.7C). Deposits of a single grain flow are commonly less than about 5 cm thick.

Debris Flows and Mud Flows

Subaerial debris flows occur under many climatic conditions but are particularly common in arid and semiarid regions where they are usually initiated after heavy rainfalls. They are also common in volcanic regions where volcanic debris become water saturated during heavy rains that accompany eruptions or from melting ice and snow that accumulate on volcanic cones between eruptions. Debris flows are slurrylike flows composed of highly concentrated, poorly sorted mixtures of sediment and water that behave in a different manner than fluid flows.

DEBRIS-FLOW MECHANISMS Debris flows behave as Bingham plastics; that is, they have a yield strength that must be overcome before flow begins. The cohesive mud matrix in debris flows has enough strength to support large grains, but cohesiveness is not great enough to prevent flow on an adequate slope. Debris flows are generally initiated on steep slopes $\left(>10^\circ\right)$ but they can flow considerable distances on gentle slopes of 5° or less; they occur in both subaerial and subaqueous environments. They consist of poorly sorted mixtures of particles, which may range to boulder-size, in a fine gravel, sand, or mud matrix. Those composed predominantly of mud-size grains are **mud flows** and those with a lower but substantial mud fraction (>5 percent by volume) are **muddy debris flows** (Middleton, 1991). The grains in these mud-bearing debris flows are supported in a matrix of mud and interstitial water that has enough cohesive strength to prevent larger particles from settling but not enough strength to prevent flow. Debris flows that have a matrix composed predominantly of cohesionless sand and gravel are **mud-free**

BOX 2.5
Types of Fluids

Depending upon the extent to which dynamic viscosity (μ) changes with shear or strain (deformation) rate, three general types of fluids can be defined. **Newtonian fluids** have no strength and do not undergo a change in viscosity as the shear rate increases (Fig. 2.5.1). Thus, ordinary water, which does not change viscosity as it is stirred or agitated, is a Newtonian fluid. **Non-Newtonian fluids** have no strength but show variable viscosity (μ) with change in shear or strain rate. Water that contains dispersions of sand in concentrations greater than about 30 percent by volume, or even lower concentrations of cohesive clay, behaves as a non-Newtonian fluid. Therefore, highly water-saturated, noncompacted muds may display non-Newtonian behavior. Such muds may flow sluggishly at low flow velocities but display less viscous flow at higher velocities.

Some extremely concentrated dispersions of sediment may behave as plastic substances, which have an initial strength that must be overcome before they yield. If the plastic material behaves as a substance with constant viscosity after the yield strength is exceeded, it is called a **Bingham plastic**. Debris flows in which large cobbles or boulders are supported in a matrix of interstitial fluid and fine sediment are examples of natural substances that behave as Bingham plastics. Water with dispersed sediment, and other plastic materials (such as ice), which behave as substances with variable viscosity after yield strength is exceeded and they start to flow, are called **pseudoplastics**. **Thixotropic substances**, a special type of pseudoplastic, have strength until sheared. Shearing destroys their strength; the substances then behave like a fluid (commonly non-Newtonian) until allowed to rest a short while, after which strength is regained. Freshly deposited muds commonly display thixotropic behavior. Shearing resulting from earthquake tremors, for example, can cause liquefaction and failure of such muds. Such momentary liquefaction may result in downslope movement of sediment that otherwise would not undergo transport. Differences in behavior of Newtonian fluids, non-Newtonian fluids, and plastic substances in response to shear stress are illustrated schematically in Figure 2.5.1.

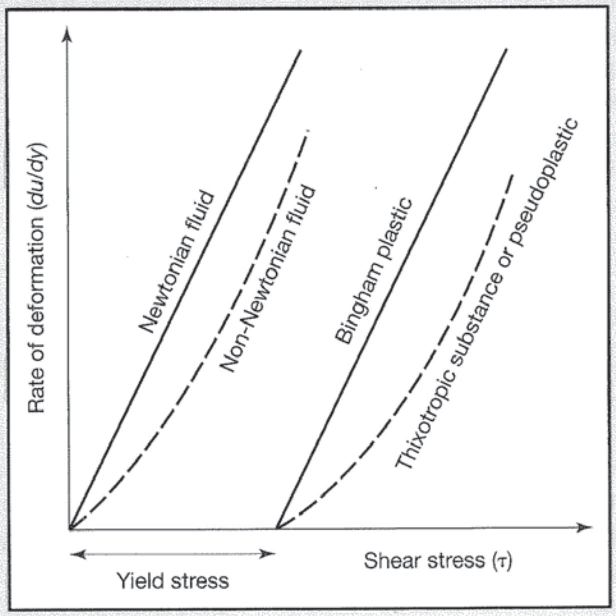

FIGURE 2.5.1 Types of fluids: Rates of deformation vs. shear stress for fluids and plastics. [After Blatt, H., G. V. Middleton, and R. Murray, 1980, Origin of sedimentary rocks, 2nd ed., Fig. 5.26, p. 187. Reprinted by permission of Prentice-Hall, Englewood Cliffs, N.J.]

debris flows (Middleton, 1991). The support mechanism for these mud-free debris flows is poorly understood.

After the yield strength of a debris flow has been overcome owing to water saturation, and movement begins, the flow may continue to move over slopes as low as 1° or 2° (Curray, 1966). Debris flows are believed to occur also in subaqueous environments, possibly as a result of mixing at the downslope ends of subaqueous slumps. As subaqueous debris flows move rapidly downslope and are diluted by mixing with more water, their strength is reduced, and they may pass into turbidity currents. Deposition of the entire mass of debris flows and mud flows occurs quickly. When the shear stress owing to gravity no longer exceeds the yield strength of the base of the flow, the mass "freezes" and stops moving.

DEBRIS-FLOW DEPOSITS Debris-flow deposits are thick, poorly sorted units that lack internal layering (Fig. 2.7D; Fig. 2.11). They typically consist of chaotic mixtures of particles that may range in size from clay to boulders. The large particles commonly show no preferred orientation. They are generally poorly graded, but if grading is present, it may be either normal or reverse.

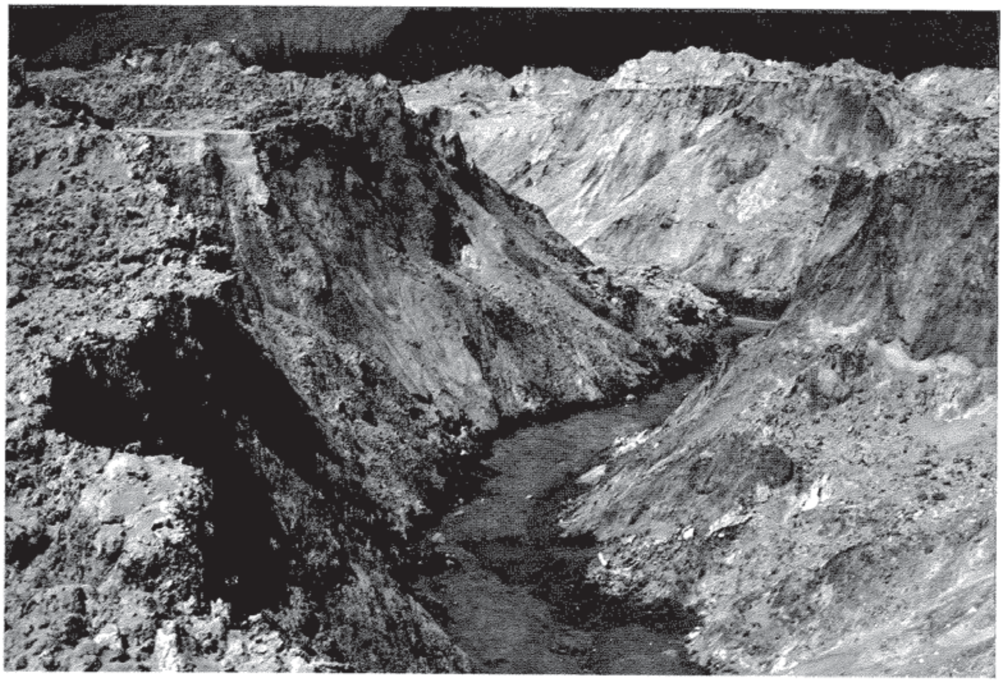

FIGURE 2.11 Photograph of recent (1980) debris flow on Mt. St. Helens, Washington. (Photograph courtesy of R. L. Schuster, online, U.S. Geological Survey Photographic Library.)

Further Reading

Balson, P. S., and M. B. Collins (eds.). 2007. Coastal and shelf sediment transport. *Geological Society Spec. Pub. 274*, London: The Geological Society.

Bouma, A. H., and C. G. Stone (eds.). 2000. *Fine-grained turbidite systems*. Tulsa, OK: American Association of Petroleum Geologists (AAPG) Memoir 72 and the Society for Sedimentary Geology (SEPM) Special Pub. 68.

Gyr, A., and K. Hoyer. 2006. *Sediment transport: A geophysical phenomenon*. Dordrecht: Springer.

Robert, A.. 2003. *An introduction to fluvial dynamics*. Arnold, London, 214 p.

Rowan, J. S., R. W. Duck, and A. Werritty (eds.), 2006, Sediment dynamics and the hydromorphology of fluvial systems. *International Association of Hydrological Sciences Pub. 306*. Oxfordshire, UK: Wallingford.

Seminara G., and P. Blondeaux (eds.). 2001. *River, coastal, and estuarine morphodynamics*. Berlin: Springer.

References Cited

Allen, J. R. L. 1985. Loose-boundary hydraulics and fluid mechanics: Selected advances since 1961. *in* Brenchley, P. J., and B. P. J. Williams (eds.). *Sedimentology: Recent developments and applied aspects*. Oxford: Geol. Soc., Blackwell. 7–30.

Allen, J. R. L. 1994. Fundamental properties of fluids and their relation to sediment transport processes. *in* Pye, K. (ed.). *Sediment transport and depositional processes*. Oxford: Blackwell Scientific Publ. 25-60.

Bagnold, R. A. 1956. *The flow of cohesionless grains in fluids*. London: Royal Soc., Philos. Trans., Ser. A. 249:235–297.

Bagnold, R. A. 1962. Auto-suspension of transported sediment. Turbidity currents. Royal Soc. London Proc. (A). 265:315–319.

Bitzer, K., and R. Pflug. 1990. A three-dimensional model for simulating clastic sedimentation and isostatic compensation in sedimentary basins. *in* Cross, T. A. (ed.). *Quantitative dynamic stratigraphy*. Englewood Cliffs, NJ: Prentice Hall. 335–348.

Blatt, H., G. V. Middleton, and R. Murray. 1980. *Origin of sedimentary rocks*. 2nd ed. Englewood Cliffs, NJ: Prentice-Hall.

Bouma, A. 1962. *Sedimentology of some flysch deposits*. Amsterdam: Elsevier.

Bouma, A. H., and C. G. Stone. 2000. *Fine-grained turbidite systems*. AAPG Memoir 72. Tulsa, OK: Amer. Assoc. Pet. Geol.

Curray, R. R. 1966. Observations of alpine mudflows in the Tenmile Range, central Colorado. *Geol. Soc. America Bull.* 77:771–776.

Dill, D. F. 1966. Sand flows and sand falls. *in* Fairbridge, R. W. (ed.). *Encyclopedia of oceanography*. New York: Reinhold. 763–765.

Galehouse, J. S. 1971. Sedimentation analysis. in Carver, R. E. (ed.). *Procedures in sedimentary petrology*. New York: John Wiley & Sons. 69–94.