Chapter 2. Transport and Deposition of Siliciclastic Sediments

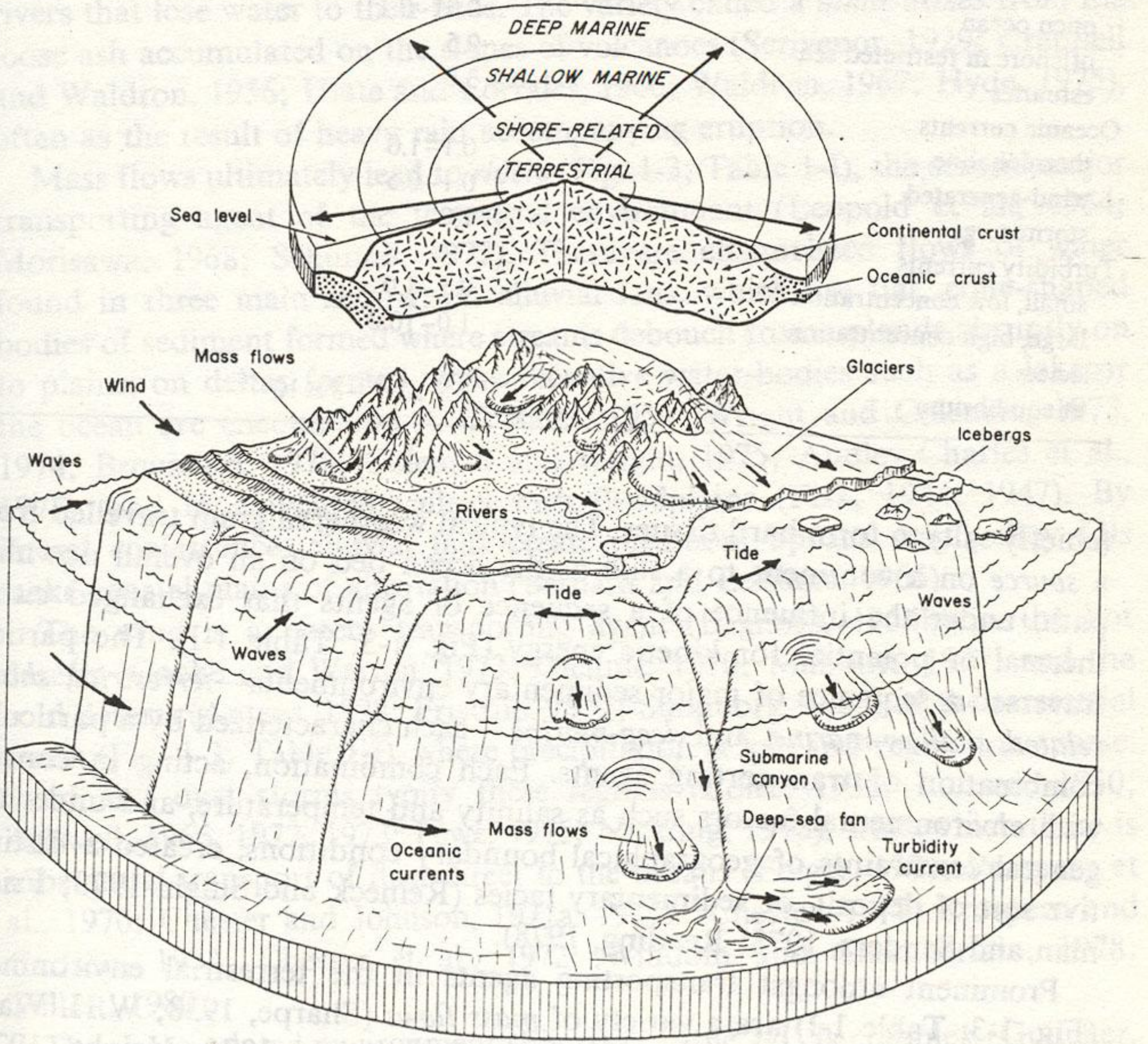


Fig. 1-3. Sedimentary environments, sedimentary agents, and sediment transport paths (source-sink relationship) for an ideal continent and bordering ocean.

Allen (1984), p.5

Sediment transport paths in continental margins

2.2 Fundanentals of fluid flow

Properties which strongly influence the flow of fluid (the ability of flow to erode and transport sediments)

(1) Density(ρ)

Determines the magnitude of forces such as stress which act within the fluid and on the bed ; the way in which waves are propagated through the fluid ; the buoyant forces acting on sedimentary particles; effective density (ρ_s -ρ_f)

(2)Dynamic viscosity (μ)

Describes the ability of the fluid to flow. It is defined as the ratio of the shear stress(τ) to the rate of deformation (du/dy) sustained by that shear across the fluid:

$$\mu = \frac{\tau}{du/dy}$$ (e.g., water at 20^{o}C, μ=0.0001kgm^{-1}s^{-1})

$$\tau = \mu \frac{du}{dy}$$ Newton's la of viscosity

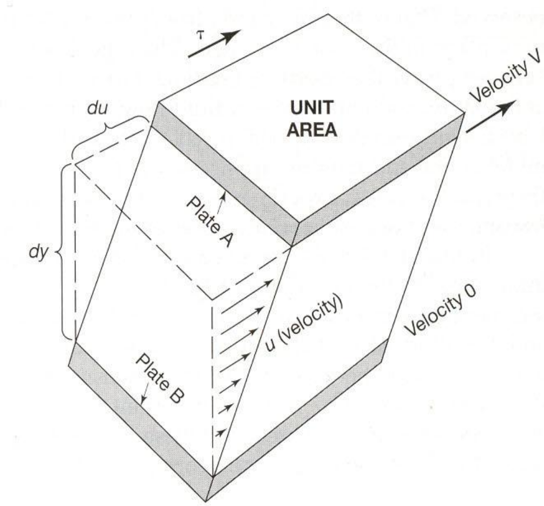
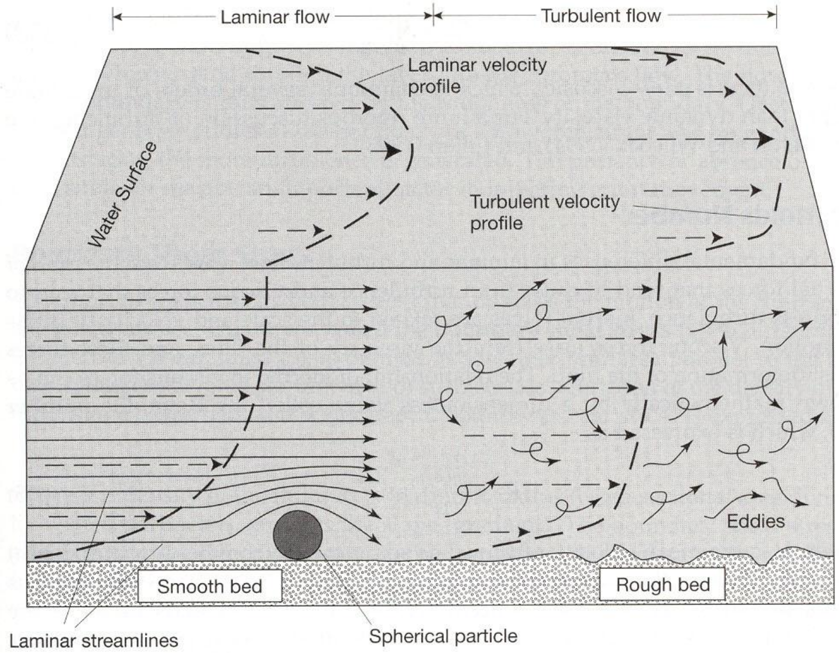


Figure 2.1
Geometric representation of the factors that determine fluid viscosity. A fluid is enclosed between two rigid plates, A and B. Plate A moves at a velocity (*V*) relative to Plate B. A shear force (τ) acting parallel to the plates creates a steady-state velocity profile, shown by the inclined line, where fluid velocity (*u*) is proportional to the length of the arrows. The shear stress may be thought of as the force that produces a change in velocity (*du*) relative to height (*dy*) as one fluid layer slides over another. The ratio of shear stress to *du*/*dy* is the viscosity (μ).

Boggs, 2001, p.26

2.2.1 Laminar vs. turbulent flow

The modes of fluid motion (laminar vs. turbulent) depend on:
(1) flow velocity;
(2) fluid viscosity; and
(3) roughness of the bed.



Boggs (2001), p.29

Figure 2.3 Schematic representation of laminar and turbulent fluid flow. Compare streamline flow under laminar flow conditions to the chaotic pattern of flow under turbulent conditions. Also, compare the shape of the laminar velocity profile to that of the turbulent velocity profile (heavy dashed lines).

Laminar flow: low velocity, high viscosity, and smooth beds (e.g., mud-supported flow, glacial ice, lava flow). Laminar flow is less easy to erode the underlying sediment bed.
Low R_e(<500-2000) and $\tau = \mu \dfrac{du}{dy}$

Turbulent flow: more rapid flow, lower fluid viscosity, rough beds. Virtually, all natural Conditions involving air or water, the flows are commonly turbulent flows. Turbulent water Masses are referred to as eddies. Eddies (1) decrease the settling velocity of suspended sediments; (2) increase the ability of flow to erode and entrain particles from a sediment bed; (3) increase apparent viscosity, called eddy viscosity, η(eta), because of the turbulent momentum transfer in the flow. So, for a turbulent flow the Newton's law of viscosity should Write:

$$\tau = (\mu + \eta)\frac{du}{dy}$$

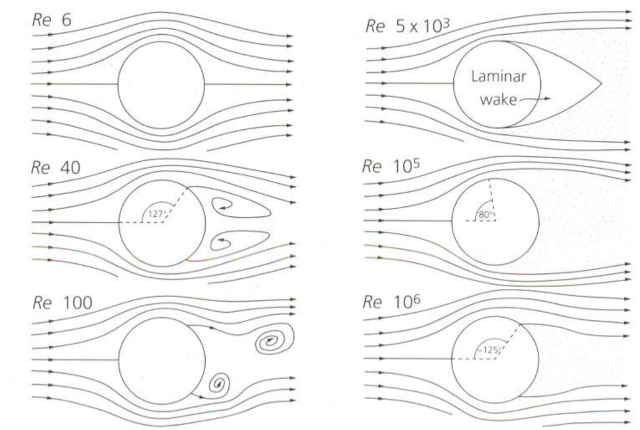
η is commonly several orders of magnitude higher than dynamic viscosity.

2.2.2 Reynolds number (R_e)

The factors that control the level of turbulence are usually combined to derive a Reynolds Number (R_e) for the flow. It is the ratio between the inertial forces related to the scale and velocity of the flow - which will tend to promote turbulence - and the viscous forces - which tend to suppress turbulence

$$R_e = \frac{Ud\rho}{\mu} \equiv \frac{Ud}{\nu}$$

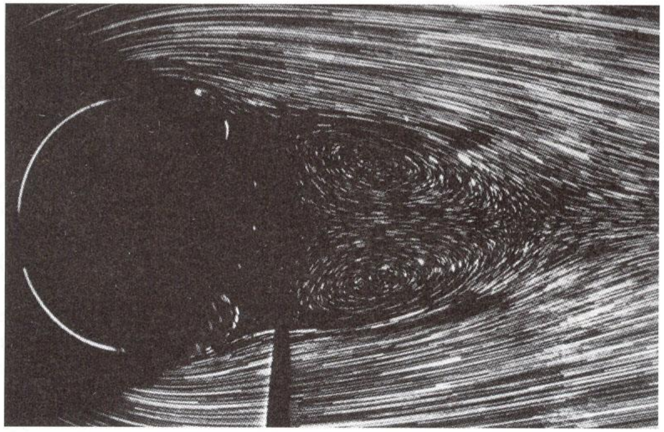
(a)



U: mean velocity of the flow; *d*: thickness of flow

v is the kinematic viscosity defined as : v $= \dfrac{\mu}{\rho}$

(b)



R_e < 500-2000 laminar flow
R_e = 500-2000 transition
R_e > 2000 turbulent flow

2.2.3 Boundary layer and velocity profile

Boundary shear stress(τ_0) is the shear stress which acting on the bed. It is a function of depth (h), bed slope (S=sinα,α is the slope angle), the nature of the fluid, and indirectly a function of velocity of flow. Boundary shear stress is important in determining the erosion and transport of sediments on the bed below a flow.

Bondary layer is the region of fluid flow next to the boundary across which the fluid velocity grades from that of the boundary (commonly zero) to that of the unaffected part of the flow.

Allen (1997), p.191

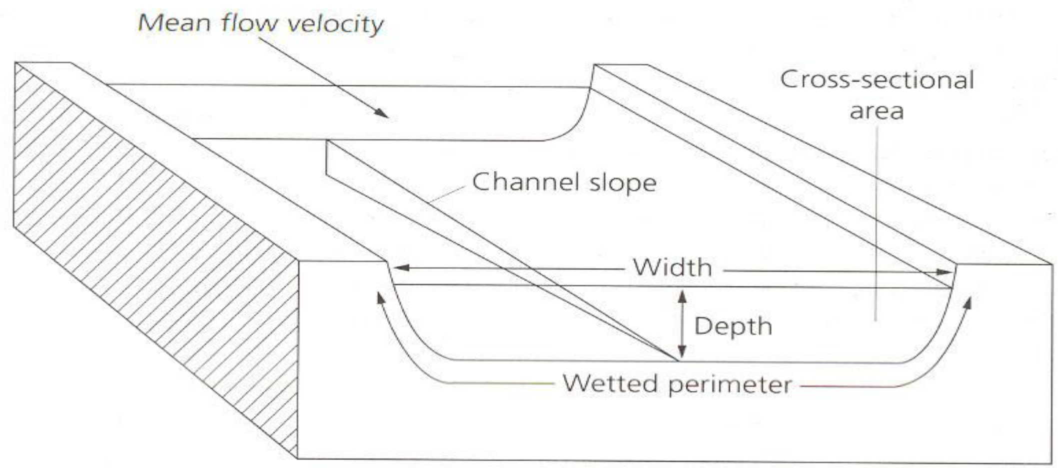


Fig. 5.8 The wetted perimeter and hydraulic radius in a river.

Discharge = Cross-sectional area × mean flow velocity
Hydraulic radius = Cross-sectional area/wetted perimeter

A simple approximation:

$\tau_0 = \rho RhS$

Boundary Shear Stress

Where

ρ:fluid density

Rh: hydraulic radius (cross-sectional area divided by wetted perimeter)

S: slope

Velocity Profile

On the river bed:
Velocity=0(minimun)
Shear stress maximum

$\tau_0 = \rho g d S$ for a 2D case

On top of the river flow:
Velocity maximum
Shear stress=0 (minimum)

Velocity profile for a laminar flow is parabolic of the form

$$u = \frac{\rho g s}{\mu}\left\{ yd - \frac{y2}{2} \right\}$$

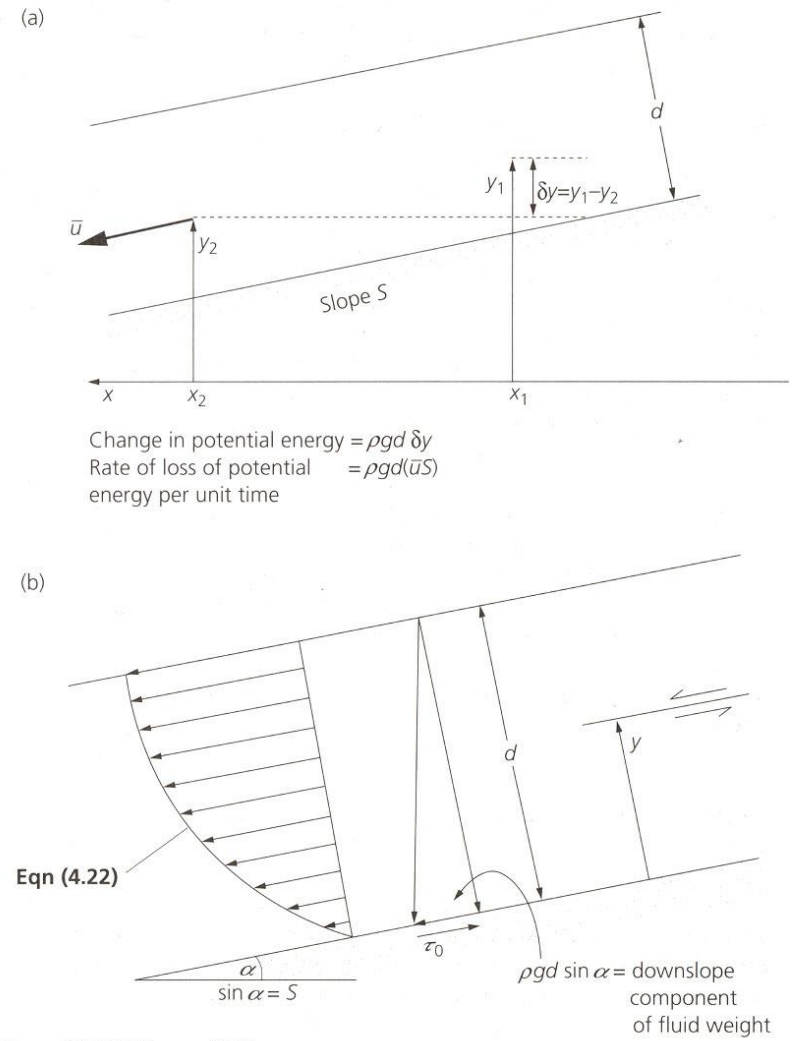


Fig. 4.12 (a) Schematic of the energy balance for a river flowing down a slope. (b) Notation used to derive the velocity profile in Practical Exercise 4.2.

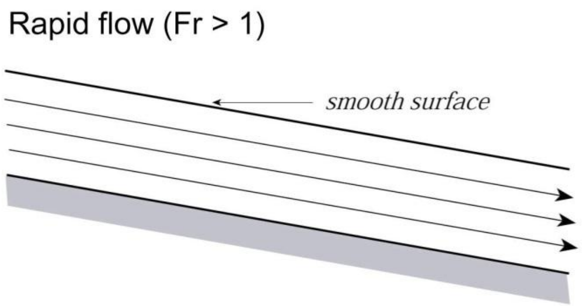
Allen (1997), p.165

2.2.4 Froude number

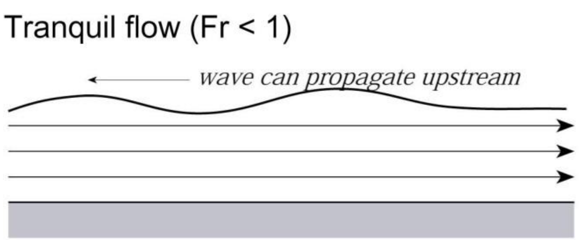
Froude number (F_r) is a dimensionless number that is proportional to the ratio of the inertial-to gravity forces within a fluid; it is equal to the average speed of a flow divided by the square root of the product of the gravitational acceleration and the depth.

$$F_r = \frac{U}{\sqrt{gd}}$$

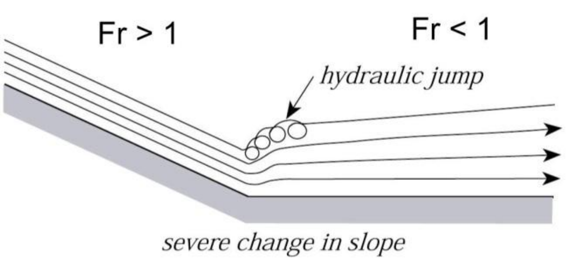
U: flow velocity
d: flow thickness



Fr>1: rapid (or shooting/supercritical) flow, turbulent suppressed, waves cannot travel upstream.



Fr<1: tranquil (or streaming/subcritical) flow, abundant turbulent, waves travel upstream (wave speed, $\sqrt{gd}$,move faster than current flow, U)



2.3 Particle transport by fluids

2.3.1 Particle entrainment by currents

Grain is moved when: fluid force (Lift force + drag force) > gravity force + grain friction

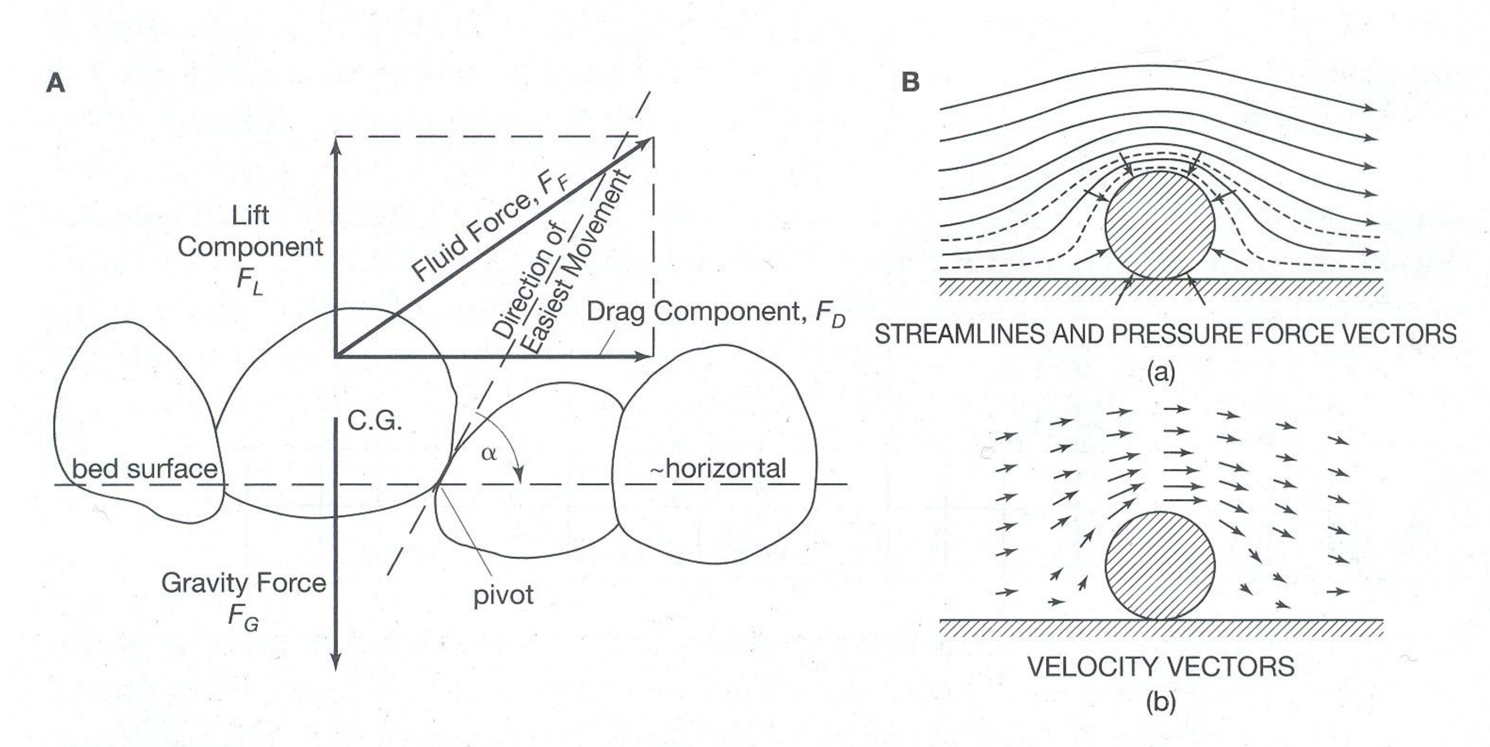


Fig. 2.2 A. Forces acting during fluid flow on a grain resting on a bed of similar grains. B. Flow pattern of fluid moving over a grain, illustrating the lift forces generated owing to the Bernoulli effect: (a) streamlines and the relative magnitude of pressure acting on the surface of the grain. (b) direction and relative velocity of velocity vectors; higher velocities occur where streamlines are closer together.

Hjulstrom diagram

The beginning of grain movement is mainly determined by: (Hjustrom's diagram)

A. Water **flow velocity**
B. **Grain size** of sediments

1. Hjulstrom diagram shows the critical velocity for movement of quartz grains on a plane bed at a water depth of 1 m.

2. for grains >0.5 mm,critical velocity increases as grain size increases (where grains are easily moved as individuals they are said to be non-cohesive).

3. for grain <0.05 mm (silt and clay-size), critical velocity increases as grain size Decreases (due to cohesion of finer-size particles—cohesive behavior).

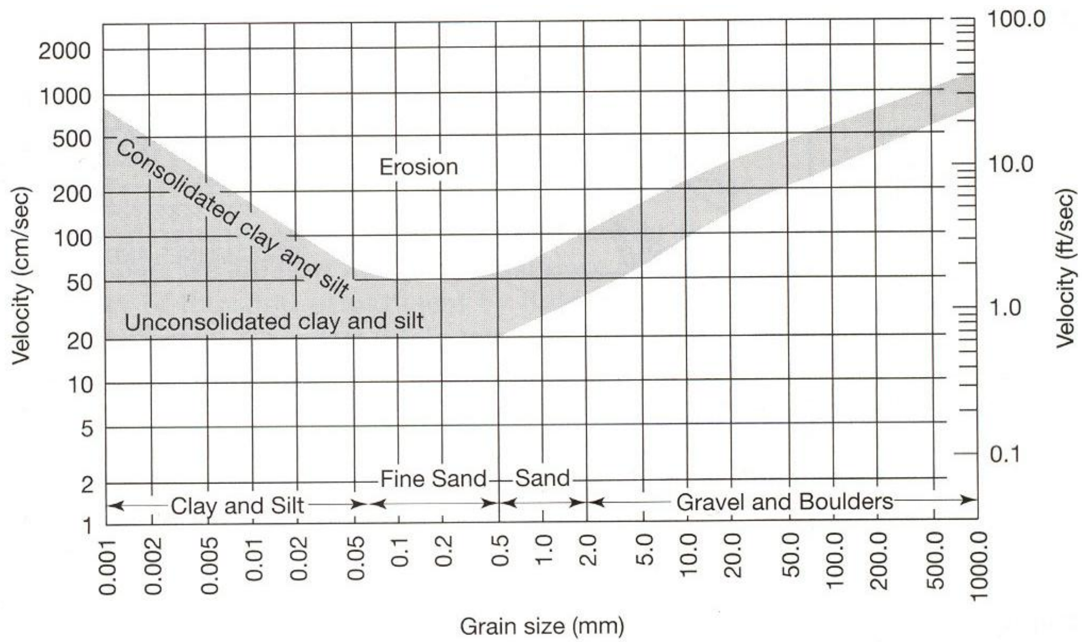


Figure 2.5
The Hjulström diagram, as modified from Sundborg, showing the critical current velocity required to move quartz grains on a plane bed at a water depth of 1 m. The shaded area indicates the scatter of exeperimental data, and the increased width of this area in the finer grain sizes shows the effect of sediment cohesion and consolidation on the critical velocity required for sediment entrainment. [After Sundborg, A., 1956, The River Klarälven, a study of fluvial processes: Geografiska Annaler, Ser A., v. 38, p. 197, reprinted by permission.]

Boggs (2001), p.34

2.3.2 Role of particle settling velocity in transport

Drag force exerted by the fluid on a falling grain

$$C_D \pi \frac{D^2}{4} \frac{\rho_f V^2}{2}$$

where C_D is a drag coefficient theat depends upon the

grain Reynolds number ($R_{eg} = \frac{U^* * D \rho_f}{\mu}$,U* is the shear velocity)

and the particle shape.

Downward force owing to gravity $\frac{4}{3}\pi\left(\frac{D}{2}\right)^3 \rho_s g$

Upward force due to buoyancy $\frac{4}{3}\pi\left(\frac{D}{2}\right)^3 \rho_f g$

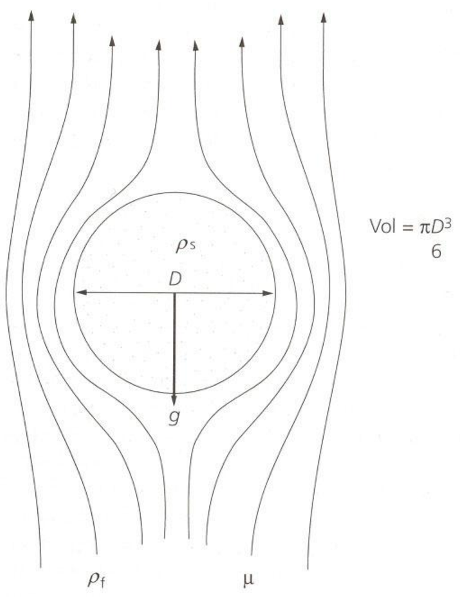


Fig. 4.6 Notation for a solitary spherical particle falling through a still fluid.

Allen (1997), p.158

As the grain falls down in a constant velocity →Drag force=gravity force-buoyancy force

Rearranging terms and for slow laminar flow at low concentrations of particles and low

grain Reynolds number, $C_D = \frac{24}{R_{eg}} = \frac{24\mu}{U^* D \rho_f}$

Substituting C_D,we have

$$V = \frac{1}{18} \frac{(\rho_s - \rho_f) g D^2}{\mu}$$

which is **Stokes' law** of settling.
V: settling velocity
D: grain diameter

2.3.3 Sediment loads and transport paths

Three modes of sediment transports:

Bed load: The sediment grains are in contact with the sediment beds and move by traction.

Suspended load: The sediment grains move above the sediment bed, but can be intermittently changed with the bed load (or called ***intermittent suspension load***).

Wash load: They are very fine-grained particles, and once taken into suspension, remain in suspension until deposited by decelerating flows.

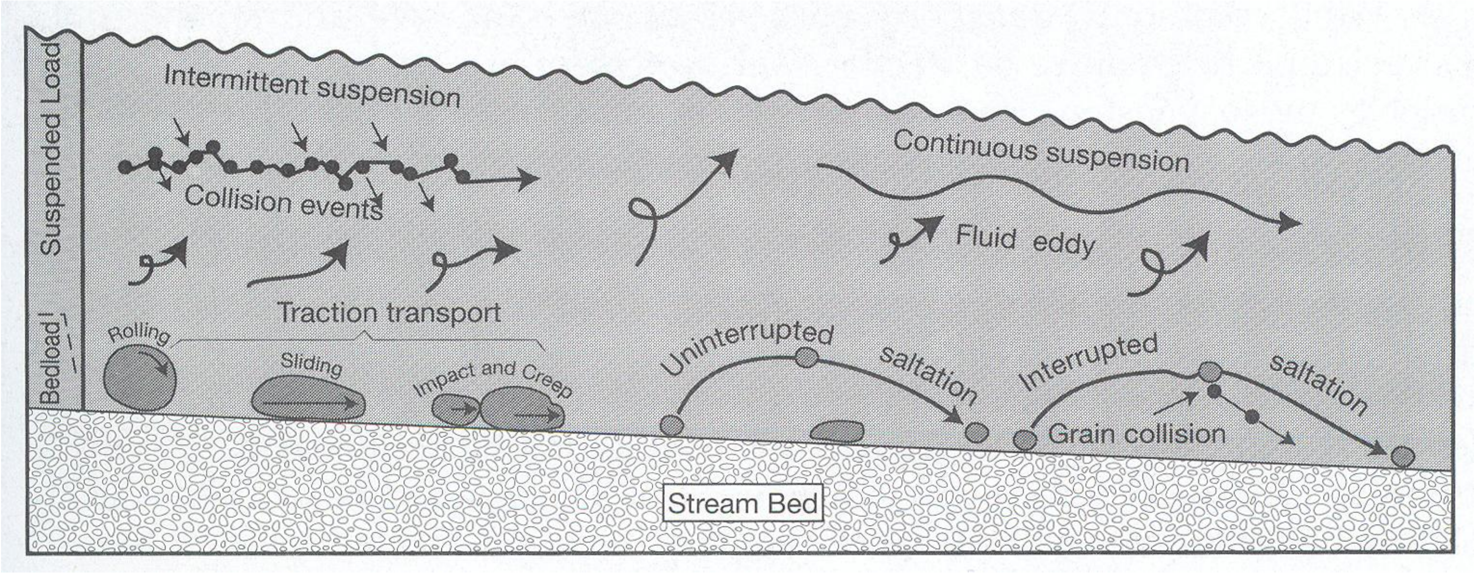
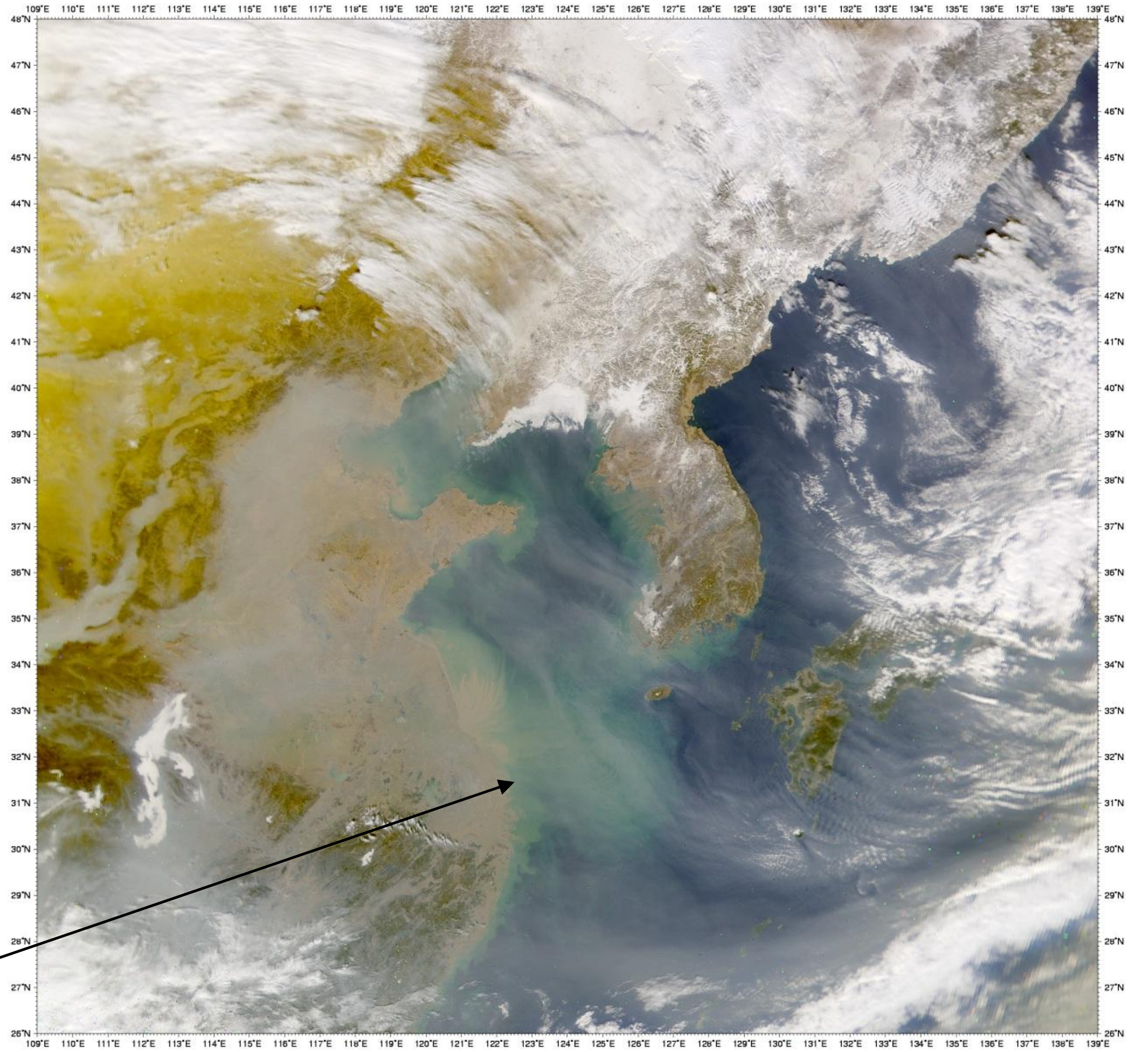


Fig. 2.4 Schematic illustration of grain paths during bedload, suspension, and saltation transport.

2.3.4 Transport by wind

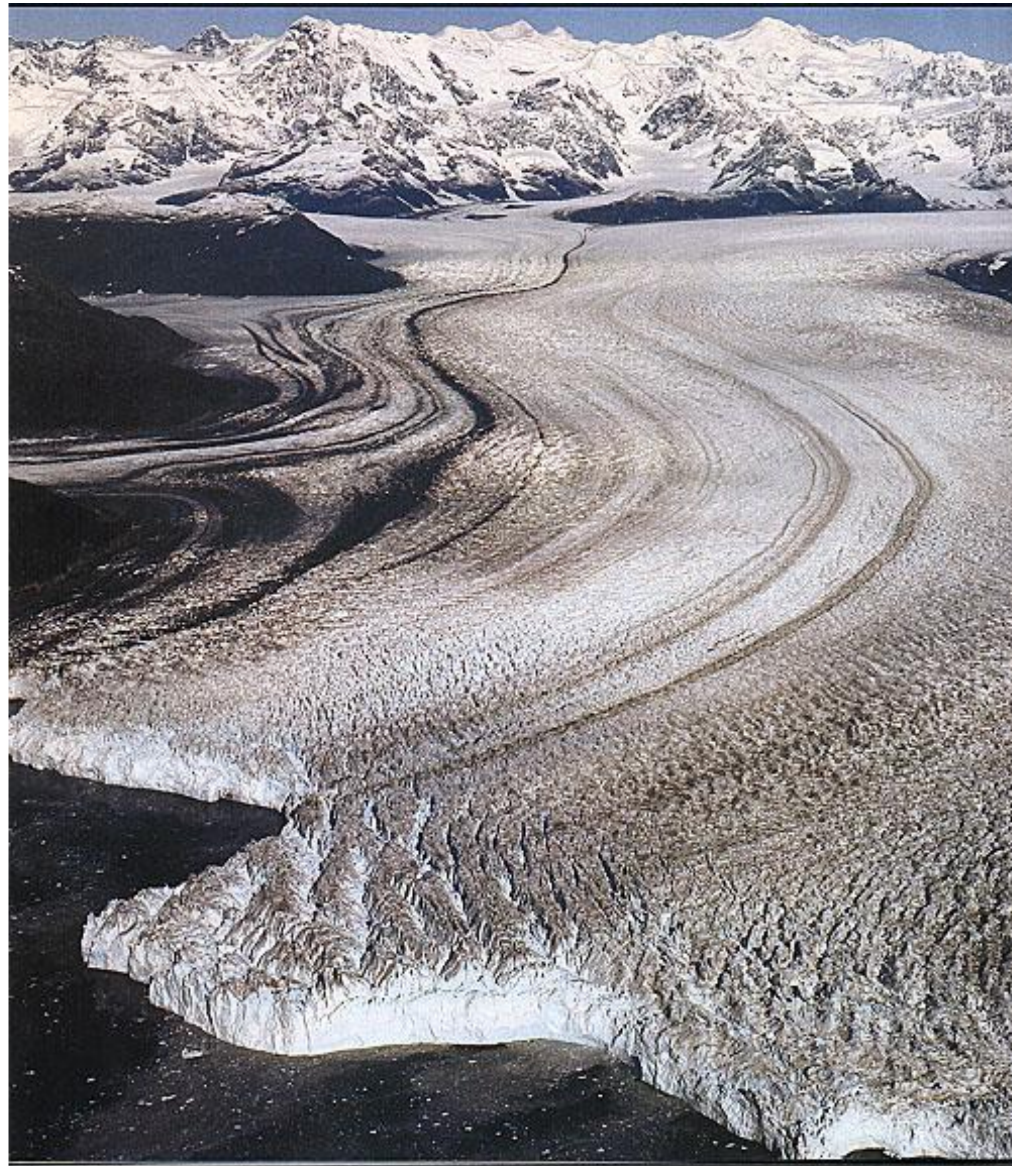
Wind is competent to transport and deposit particles in the size range of sand to dust (clay) only because of its low density and viscosity. Sand-sized particles move by traction and saltation; dust-sized particles move by suspension (e.g. dust clouds). The very fine-grained component of deep-sea pelagic sediments is believed to be largely of windblown origin.

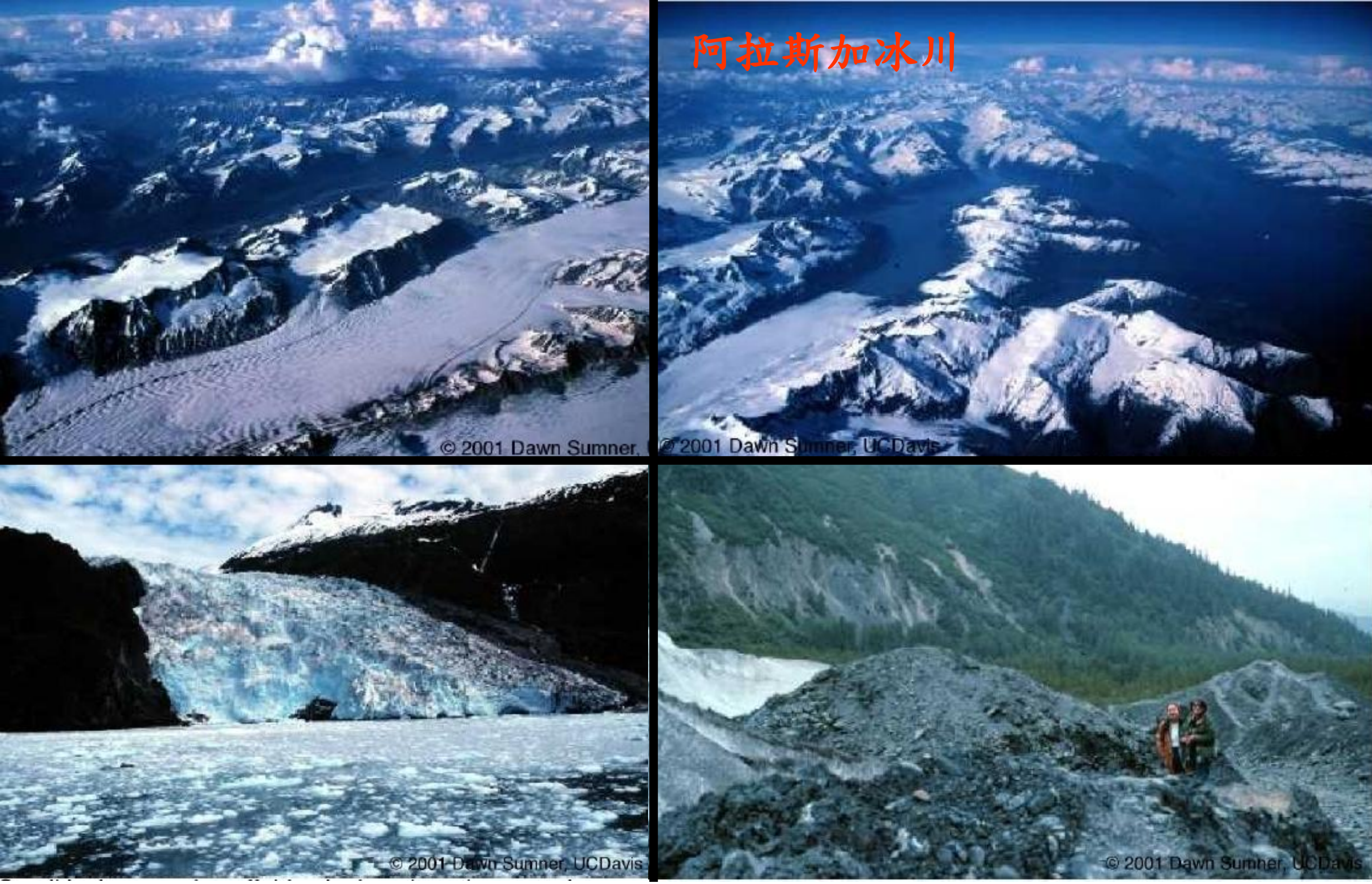


Dust cloud from China

2.3.5 Transport by glacial ice

Glaciers flow as a non-Newtonian pseudoplastic fluid. Glacial transport does not generate bedforms. Glacier is able to transport particles of enormous size as well as particles of the smallest sizes because of its high viscosity. When melting occurs at the front of a glacier, the sediment load is dumped as unsorted, poorly layered glacial moraine.

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2.3.6 Deposits of fluid flows

Sediments **deposited from traction current** commonly preserved sedimentary structures such as cross-beds, ripple marks, and pebble imbrication that display directional features from which the direction of the ancient fluid flow can be determined.

Sediments **deposited from suspension** lack these flow structures and are commonly characterized instead by fine laminations.

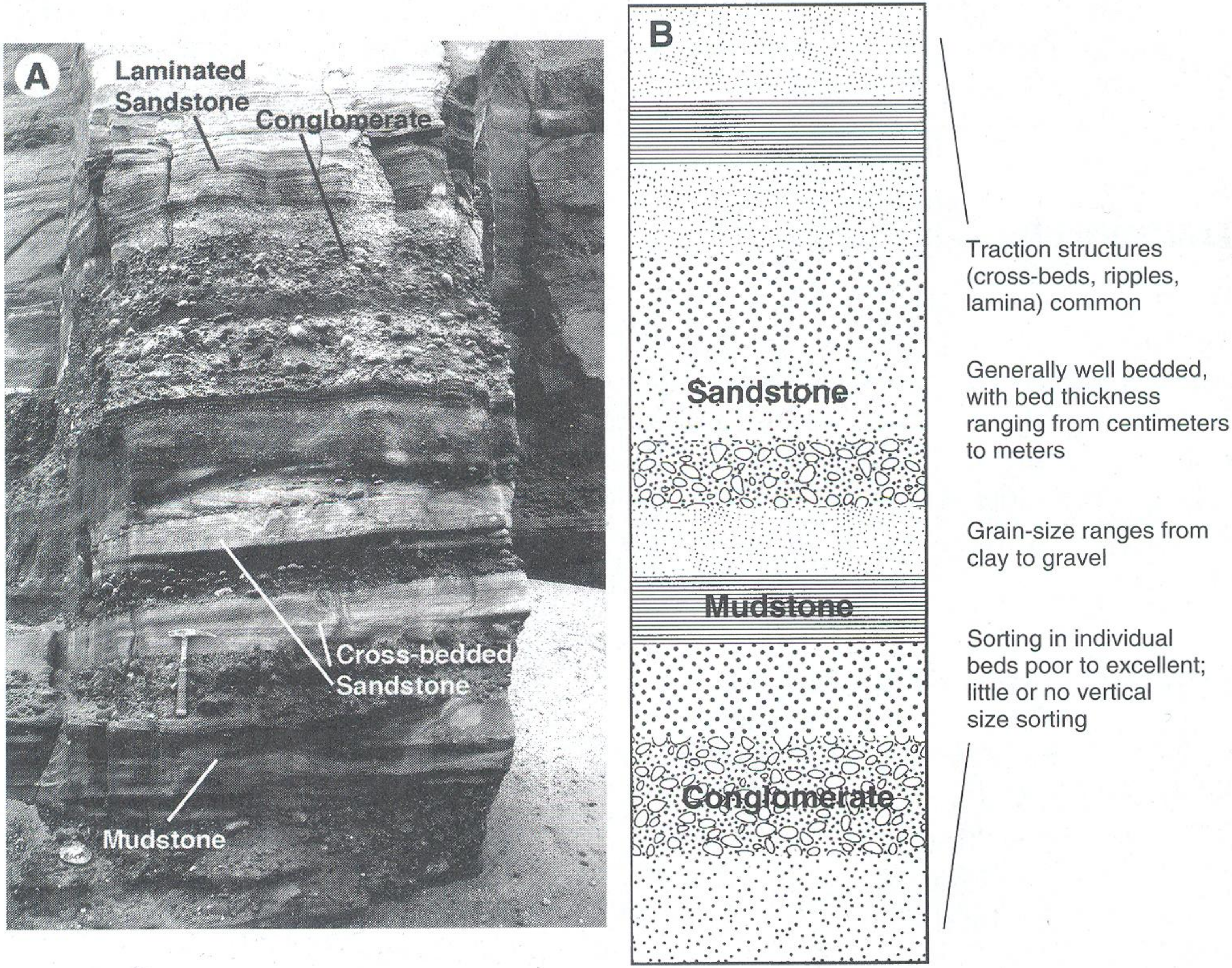
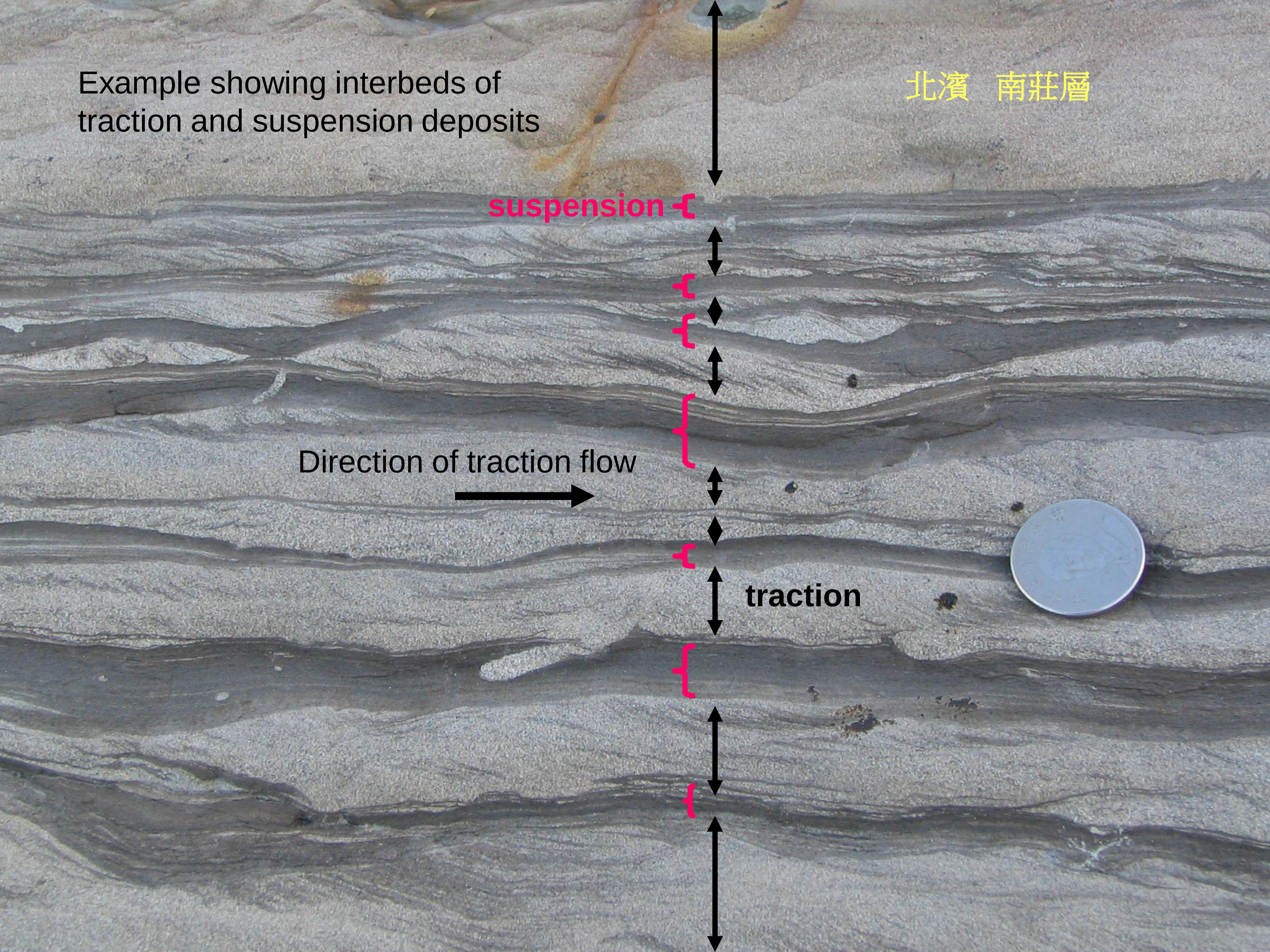
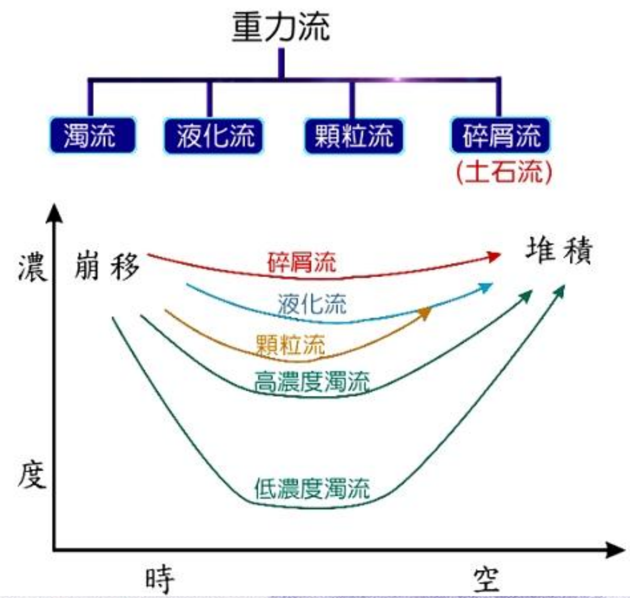


Fig. 2.5 A. Photograph of well-bedded fluid-flow deposits, Miocene, Blacklock point, southern Oregon coast. B. Schematic representation of typical characteristics of fluid-flow deposits.


Example showing interbeds of
traction and suspension deposits
北濱　南莊層
suspension
Direction of traction flow
traction

2.4 Particle transport by sediment gravity flows



Examples of sediment gravity flows

Snow avalanches;

Pyroclastic flows and base surges resulting from volcanic eruptions;

Subaerial grain flows of dry sand down the slip face of sand dunes, subaqueous grain flows;

Debris flows and mud flows of nonvolcanic or volcanic origin;

Turbidity currents


Mudflow


Pyroclastic flow

Major types of mass-transport processes

Table 2.2 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	Elastic	Shear failure among discrete shear planes with little internal deformation or rotation
Slide		Slump	Elastic	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
Sediment gravity flow	Mass flow	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
Sediment gravity flow	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
Sediment gravity flow	Fluidal flow	Fluidized flow	Viscous fluid	Cohesionless sediment supported by the forced upward motion of escaping pore fluid; thin (<10 cm) and short-lived
Sediment gravity flow	Fluidal flow	Turbidity current	Viscous fluid	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, Okla.

An example of glide (古亭坑層, 台南, 林殿順, 1991)

An example of slumped facies of complete disruption, mixing and brecciation of the strata

An example of slump (牡丹層, 恆春)

FIG. 12—Slumped facies, Cretaceous at Point Fermin, California. Laminated shales are slumped and contorted, and sandstone blocks (presumably torn from originally interbedded turbidites) are incorporated into slump.

Sediment-support mechanism for sediment gravity flows

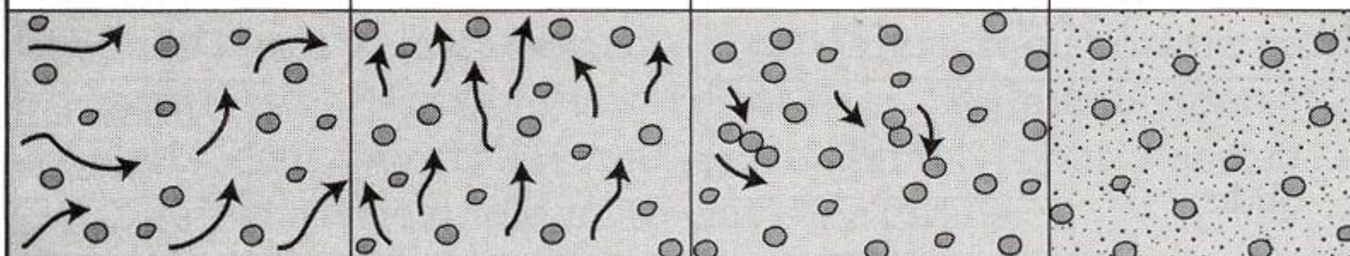
Observed Type of flow	Turbidity Current	Liquefied Flow	Grain Flow	Mud Flow; Debris Flow
Support Mechanism	Turbulent Fluid	Upward escape of intergranular fluid	Grain interaction (dispersive pressure)	Strength of matrix
				
Type of Fluid	Turbulent fluid	?Newtonian fluid (high viscosity)	Non-Newtonian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

Figure 2.14
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. Reprinted by permission of John Wiley & Sons Limited.]

Boggs (2001), p.48

2.4.1 Turbidity currents

A turbidity current is a kind of density current that flows downhill along the bottom of an ocean or lake because of density contrasts with the surrounding water arising from sediment suspended in the water owing to turbulence.

Trigger mechanisms:

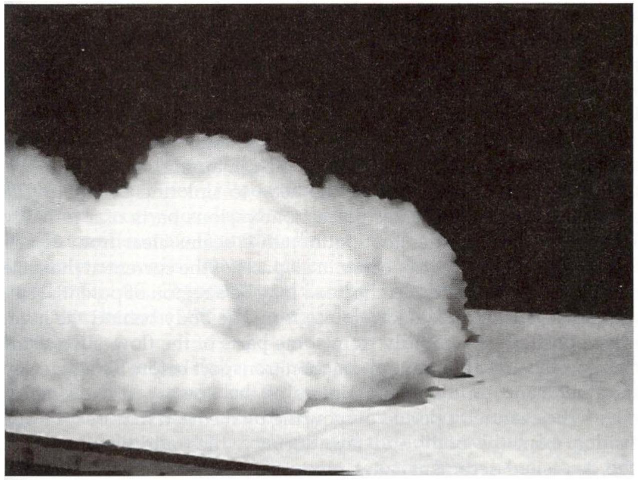
Sediment failure (e.g., earthquake-triggered);

Flow of sand triggered by storms into canyon heads;

Bedload inflow from rivers and glacial meltwater into FRESH water body;

(turbidity current triggered by this mechanism may not often occur on continental shelves where density contrast between muddy river water and saline ocean water is less than that between muddy river water and fresh water);

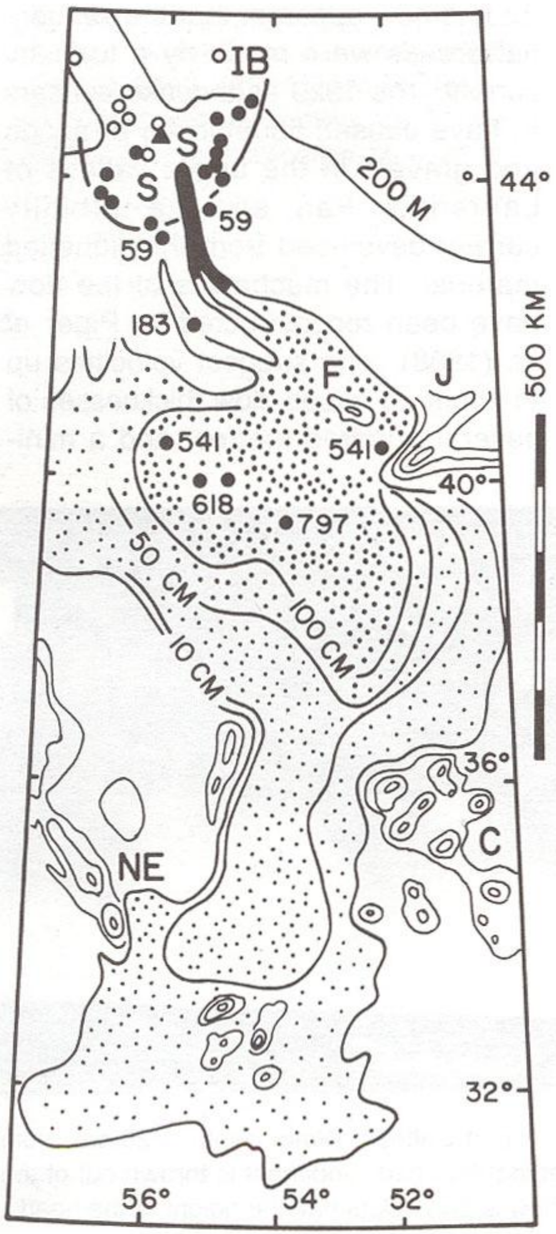
Flows during eruption of airfall ash.



Boggs (2001), p.49

Figure 2.16
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.

Example: 1929 Canadian Grand Banks turbidity current triggered by earthquake, U @ 20 m/s, distance: > 300 km, thickness: several hundred meters, turbidite: over 1 m thick



Walker (1992), p.240

Figure 2 (left) Map of the Laurentian Fan – Sohm Abyssal Plain, off the Grand Banks of Newfoundland. Shelf edge marked by the 200 m isobath, with Laurentian Channel entering map area from the northwest. Black dots show cable breaks that followed the 1929 earthquake (epicentre shown by triangle, with post 1929 quake epicentres shown with open circles). North of the line IB, the breaks were instantaneous (IB = instantaneous breaks, S = zone of slumping). South of the line, cables were broken in sequence, 59 to 797 minutes from the quake. Isopachs show the resulting turbidite (stippled), and the canyon fill (black) indicates debris flow deposits. Seafloor topography locally deflected the turbidity current, particularly the J Anomaly Ridge (J), the Fogo (F) and Corner (C) Seamounts, and part of the New England Seamount Chain (NE). Diagram at top shows detail of the upper fan channels and levees. Redrawn from Piper *et al.* (1984).

Structure of turbidity currents

Velocity of the head

$$U_{head} = 0.7\sqrt{(\frac{\Delta\rho}{\rho})gh}$$

where Δρ (ave. 0.1 g cm^{-3}) is the density contras
between the turbidity current and the ambient
water, ρ is the density of the ambient water, and
is the height of the head.

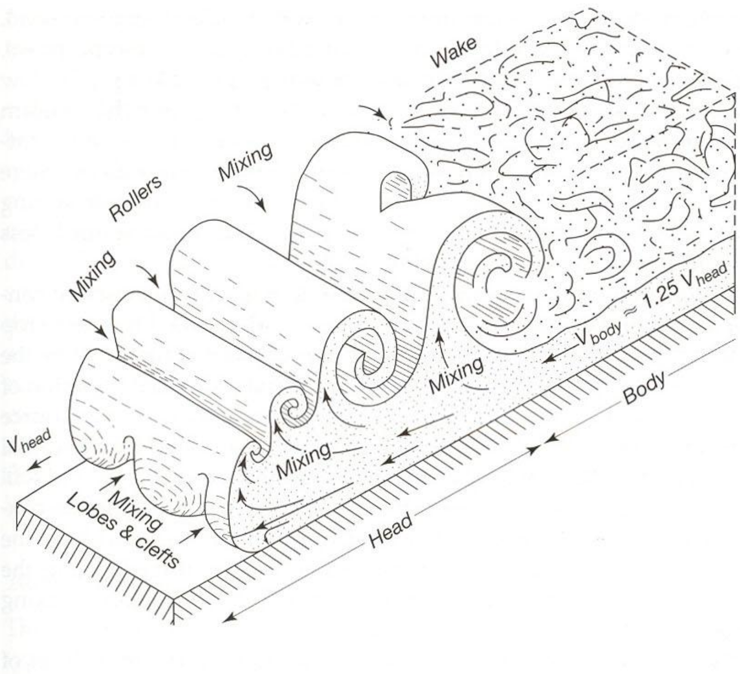


Figure 2.15
Postulated structure of the head and body of a turbidity current advancing into deep water. The tail is not shown. [From Allen, J. R. L., 1985, Loose boundary hydraulics and fluid mechanics: Selected advances since 1961, *in* Brenchley, P. J. and D. J. P. Swift (eds.), Sedimentology: Recent developments and applied aspects. Fig. 8, p. 20, reprinted by permission of Blackwell Scientific Publications, Ltd., Oxford.]

Boggs (2001), p.49

Velocity of the body

$$U_{body} = \sqrt{\frac{8g}{f_o + f_1}(\frac{\Delta\rho}{\rho})hs}$$

where *h* is the thickness of the flow, *s* is the slope of the bottom, f_0 and f_1 are the frictional resistance at the bottom and top of the flow respectively.

The head maybe a region of erosion while deposition is taking place from the body owing to differences in turbulence in the head and body.

Deposits

Types of deposits depend on:
(1) Size of currents (up to a few hundred meters thick);
(2) Density of the current
(3) Grain size of the source materials

Figure 4 Classical turbidites; monotonous interbeds of sandstone and shale, with no evidence of topography on the seafloor. Stratigraphic top to right; Devonian, Cape Liptrap, near Melbourne, Australia.
Walker (1992), p.241

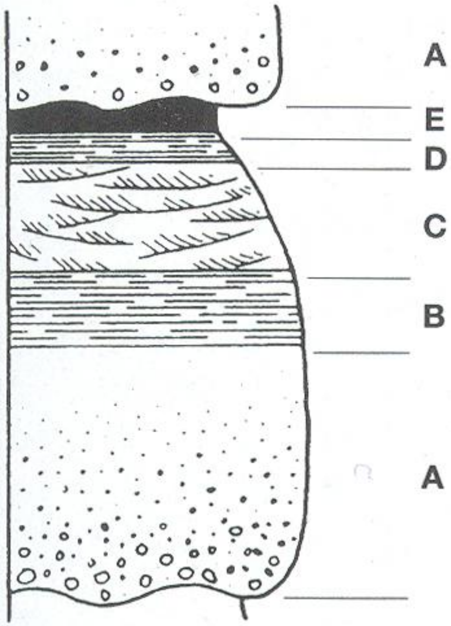
Two types of turbidity currents (depending on sediment concentration)

Low-density turbidity currents **(**<20-30% sediments**)**

These currents transport sediment up to medium sands, kept in suspension entirely by the fluid turbulence. As the flow decelerates sediment is moved as bed load in a traction carpet. The deposits of these currents are commonly classical turbidites.

Deposits
Thin- to medium bedded turbidite (few cm to @30 cm thick);
Consist of up to medium-grained sands;
Scour marks common at the bottom;
Bouma sequence.

1



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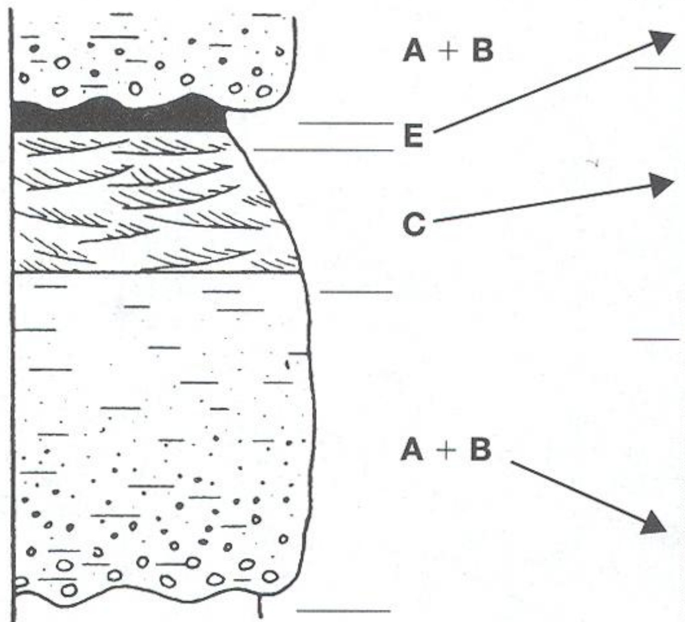

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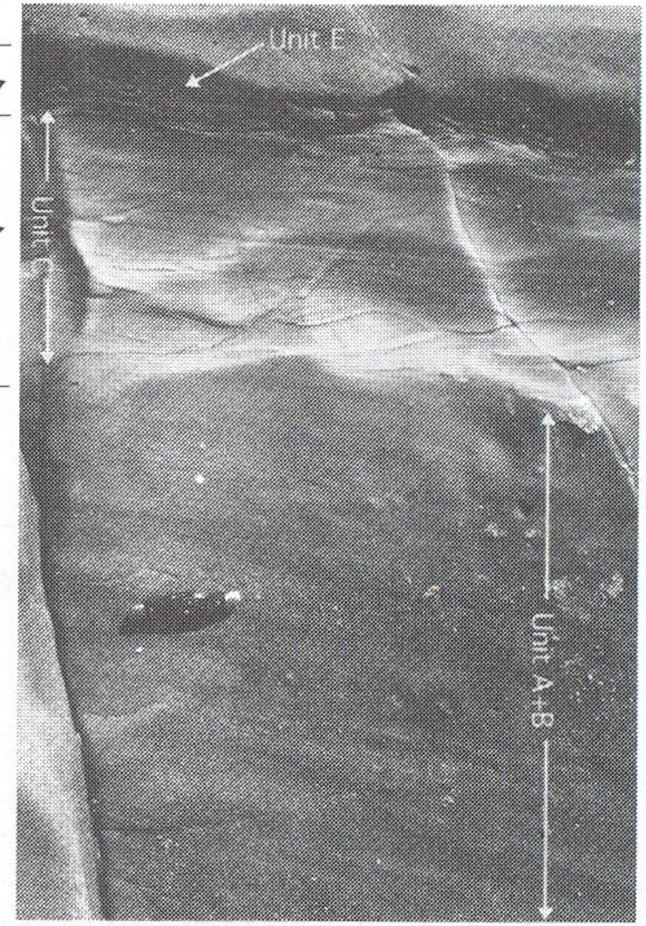


Bouma sequence

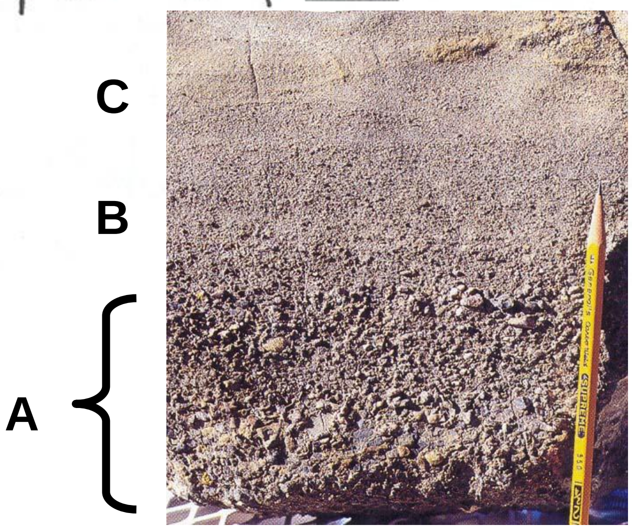


Fig. 2.8 Ideal sequence of sedimentary structures in graded-bed units as proposed by Bouma (1) and Hsu (2). Note that in Hus's model, Bouma units A and B are combined and unit D is omitted. (3) Photograph of a Bouma unit that is very similar to Hsu's model (Cretaceous, southern Oregon coast).

High-density turbidity currents (>20-30% sediments, >1.1g cm^{-3})

These currents can carry gravels and coarse sands, mostly in the form of a traction carpet at the base of the flow and in suspension just above. Fluid turbulence, dispersive pressure from grain collisions, and finer sediment exerting a matrix buoyancy lift, keep the gravels and sands moving until the flow decelerates through increasing slope or dilution.

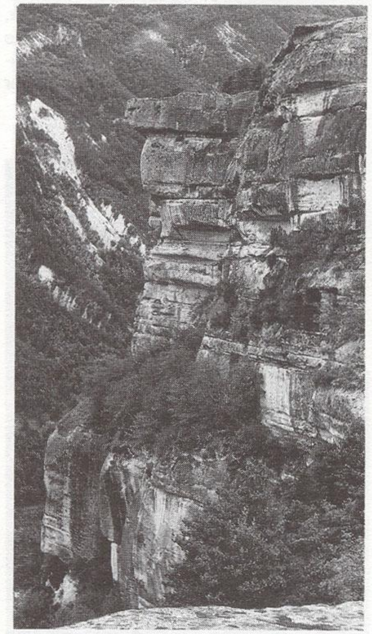


Figure 6 Massive sandstone facies in the Upper Eocene Annot Sandstone, southern France. About 180 m of section can be seen; note thickness of individual sandstone beds and absence of mudstone interbeds.

Walker (1992), p.241

Deposits

Thick-bedded turbidites containing coarse-grained sands or gravels;
Relatively poor vertical size grading and few internal laminations;
Poor developed basal scour marks

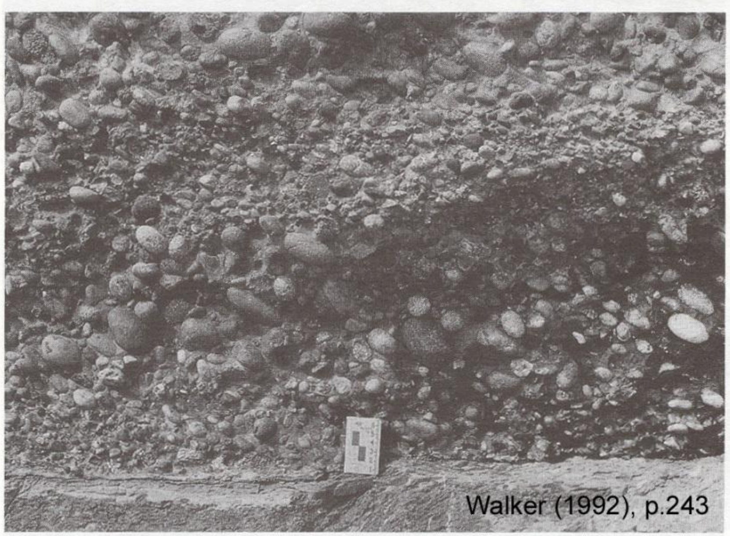


Figure 10 Inverse to normal grading in conglomerates of the Cretaceous La Jolla Formation, Tourmaline State Surfing Beach, near San Diego, California. Note also the well-developed imbrication, with clasts dipping upstream (to the right).

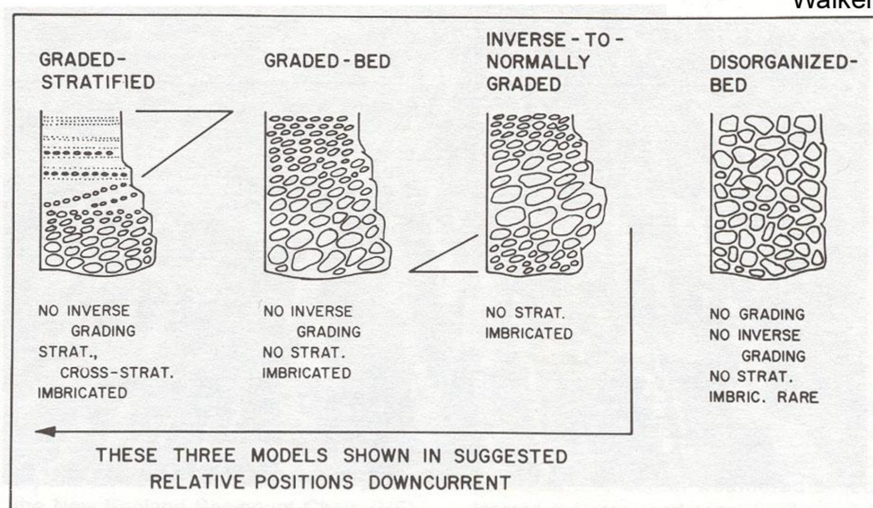


Figure 9 Four models for resedimented conglomerates, shown in their inferred downcurrent relative positions. From Walker (1975a).

Walker (1992), p.242

Comparison among different sediment gravity-flow deposits

Boggs (2001), p.52

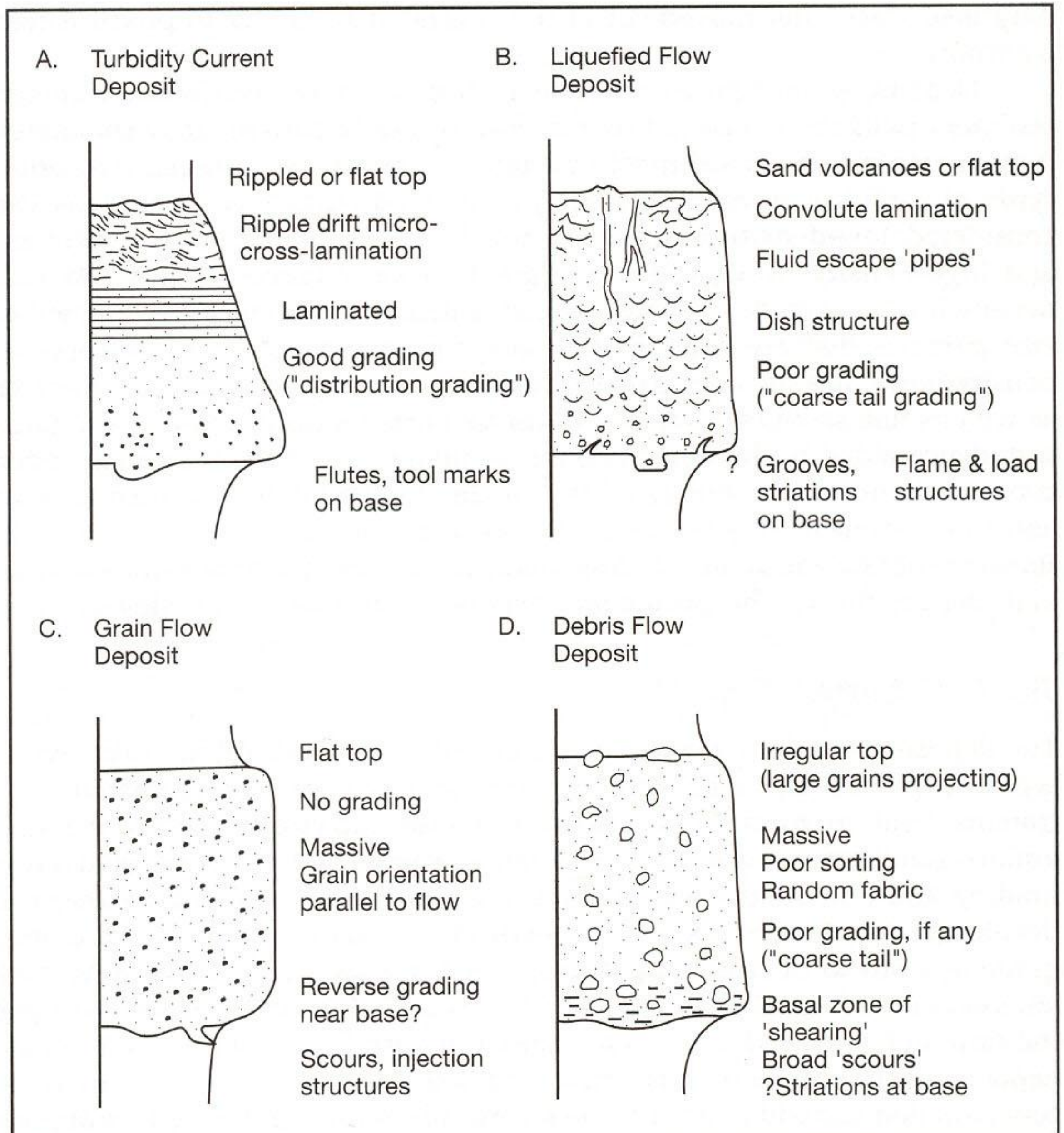
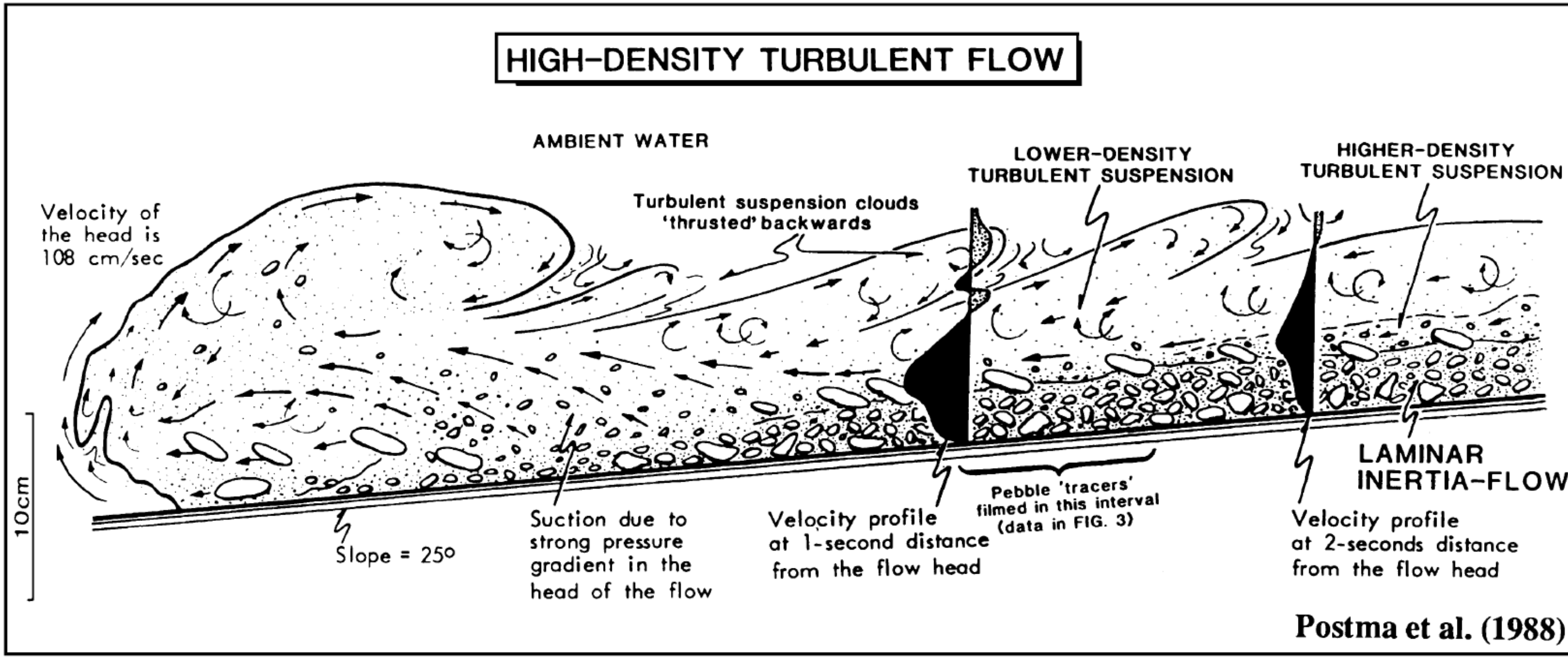


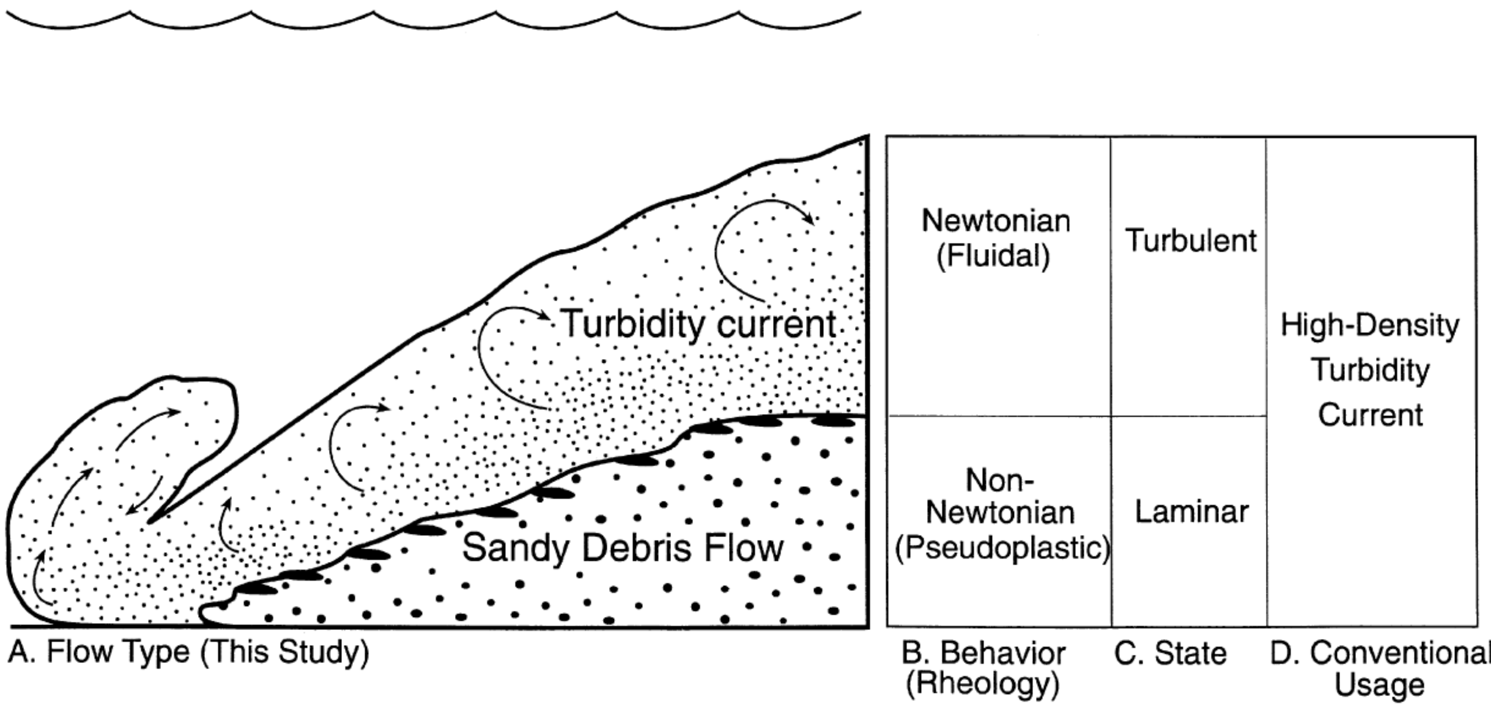
Figure 2.17
Comparison of sedimentary structures in different types of sediment gravity-flow deposits. [After Middleton, G. V., and M. A. Hampton, 1976, Subaqueous sediment transport and deposition by sediment gravity flows, *in* Stanley, D. H. and D. J. P. Swift (eds.), Marine sediment transport and environmental management: John Wiley & Sons, Inc., Fig. 9, p. 213. Reproduced by permission of John Wiley & Sons Limited.]



4. A profile of experimental "high-density turbidity current" showing density stratification. Note a lower laminar inertia-flow and an upper turbulent flow. From Postma et al. 3).

High-density turbidity currents

(or termed sandy debris flow for lower part and turbidity current for upper part)



(Postma et al, 1988)

Fig. 6. (A) An experimental view of 'high-density turbidity currents' by Postma et al. (1988) who suggested that the basal high-concentration layer (labeled as sandy debris flow) was driven by the upper low-concentration layer (labeled as turbidity current). According to Postma et al. (1988), both upper and lower layers comprise the 'high-density turbidity currents'. I consider upper and lower layers as rheologically different entities and therefore separate flow processes (see Shanmugam, 1996a). (B) According to Postma et al. (1988), lower and upper layers represent non-Newtonian and Newtonian rheology, respectively. (C) According to Postma et al. (1988), lower and upper layers represent laminar and turbulent states, respectively. The basal laminar layer (i.e., sandy debris flow) is variously termed as inertia-flow layer, traction carpet, flowing-grain layer etc. by various authors (see Shanmugam, 1996a). The basal layer with high sediment concentration promotes hindered settling, and allows development of floating mudstone clasts and quartz granules. (D) Interpretation of Postma et al. (1988). Because sediment-gravity flows are classified on the basis of rheology and sediment-support mechanism (Lowe, 1982), a single flow (i.e., high-density turbidity current) cannot be both Newtonian and non-Newtonian in rheology, and laminar and turbulent in state at the same time. This type represents gravity flow transformation of Fisher (1983). From Shanmugam (1997a).

Debrites in a submarine setting (Pyrenees, Spain)

2.4.2 Liquefied flows

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.

Liquefaction: movement of grains with saturated pore water under increased pressure. The increased pressure maybe caused by a sudden shock (e.g., earthquakes) or increasing overburden.

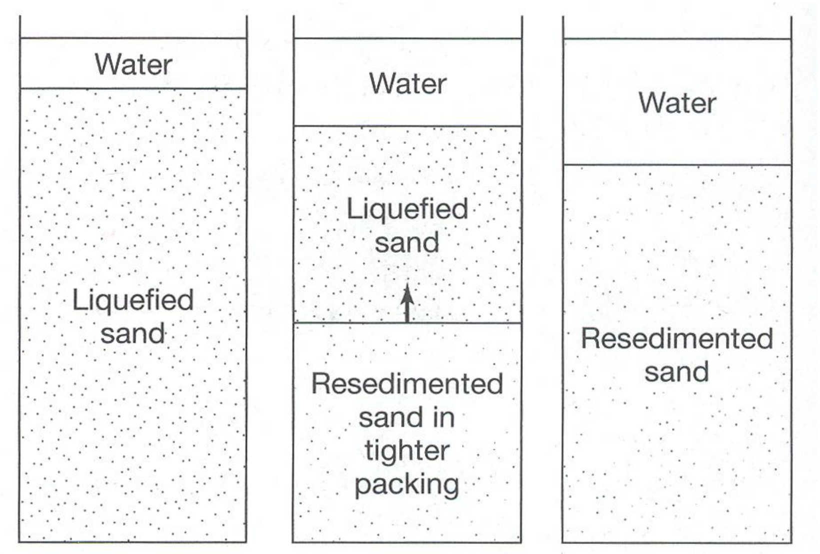


Fig. 2.9 Schematic representation of grain settling and water expulsion during deposition of sand from a liquefied flow.

Liquefied flows may travel short and stop moving once it “freezes up” because of the reestablishment of grain-to-grain contact. Liquefied flows may evolve into turbidity currents.

Deposits

Thick, poorly sorted sand unit with fluid escape structures (dish structures, pipes, and sand volcanoes).

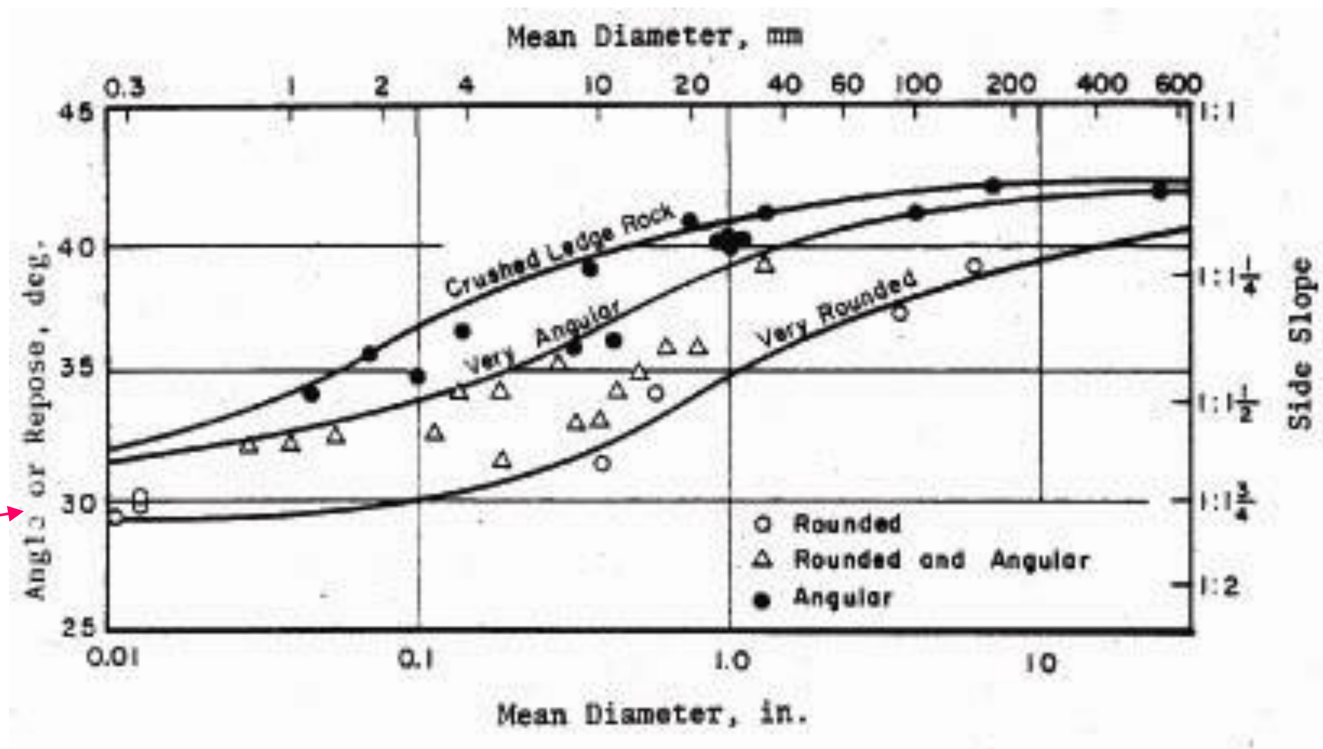
2.4.3 Grain flows

Grain flows are dispersions of cohesionless sediment in which the sediment is supported in air by dispersive pressures through direct grain-to-grain collisions and in water by collisions and close approaches. Grain flows may grade into liquefied flows if there is water present in the processes.

Example: sand avalanche down the steep side of dunes because the angle of repose is exceeded.

Angle of repose
maximum angle at which a slope of loose material will lie without cascading downward

Around 30° in sands



Deposits

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 are of limited geological significance because of the steep slopes required to initiate flow.

Characteristics: A few cm thick of sand, inverse grading


sand avalanche down the steep side of dunes
台南七股海灘風成沙丘

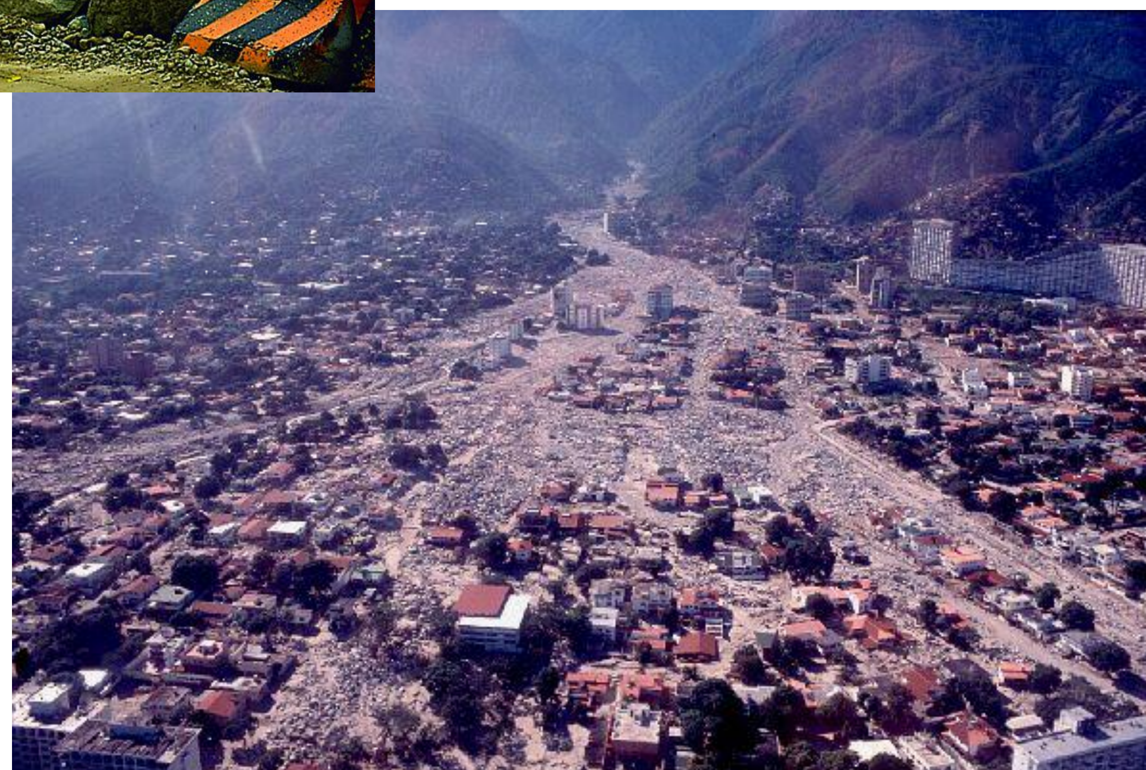
2.4.4 Debris flows and mud flows

Debris flows and mud flows are slurry like flows composed of highly concentrated, poorly sorted mixtures of sediment and water that behave as Bingham plastics; that is, they have a yield strength that must be overcome before flow begins.

Mud flows: predominantly of mud-sized grains
Muddy debris flow: matrix > 5% mud
Mud-free debris flow: matrix composed predominantly of cohesionless sand and gravel
Lahar: composed largely of volcanic materials.

雲南蔣家溝mudflow

Debris-flow deposits along a road side at 溪頭 emplaced during the 2001桃芝颱風



A series of debris flows and mudflows occurred in December 1999 at a coastal village north of Caracas, Venezuela. These sediment gravity flows claim the lives of anywhere from 10,000 to 50,000 people. The picture was taken a few days after the largest of the catastrophes. (from http://www.passcal.nmt.edu/~bob/passcal/venezuela/ven002.htm)

Deposits

Thick, poorly sorted, lack internal layering, matrix-supported (in most cases), clast-supported (uncommon), rare grading (inverse grading may present), a-axis of gravels parallel to flow direction, a-axis imbricated

(Inverse grading maybe caused by the dispersive pressures of a grain flow tend to push the larger particles to the top of the flow where they encounter less friction. The finer grain sizes, on the other hand, can move more easily in the base of the flow, where the shear stress against the bottom is greater.)

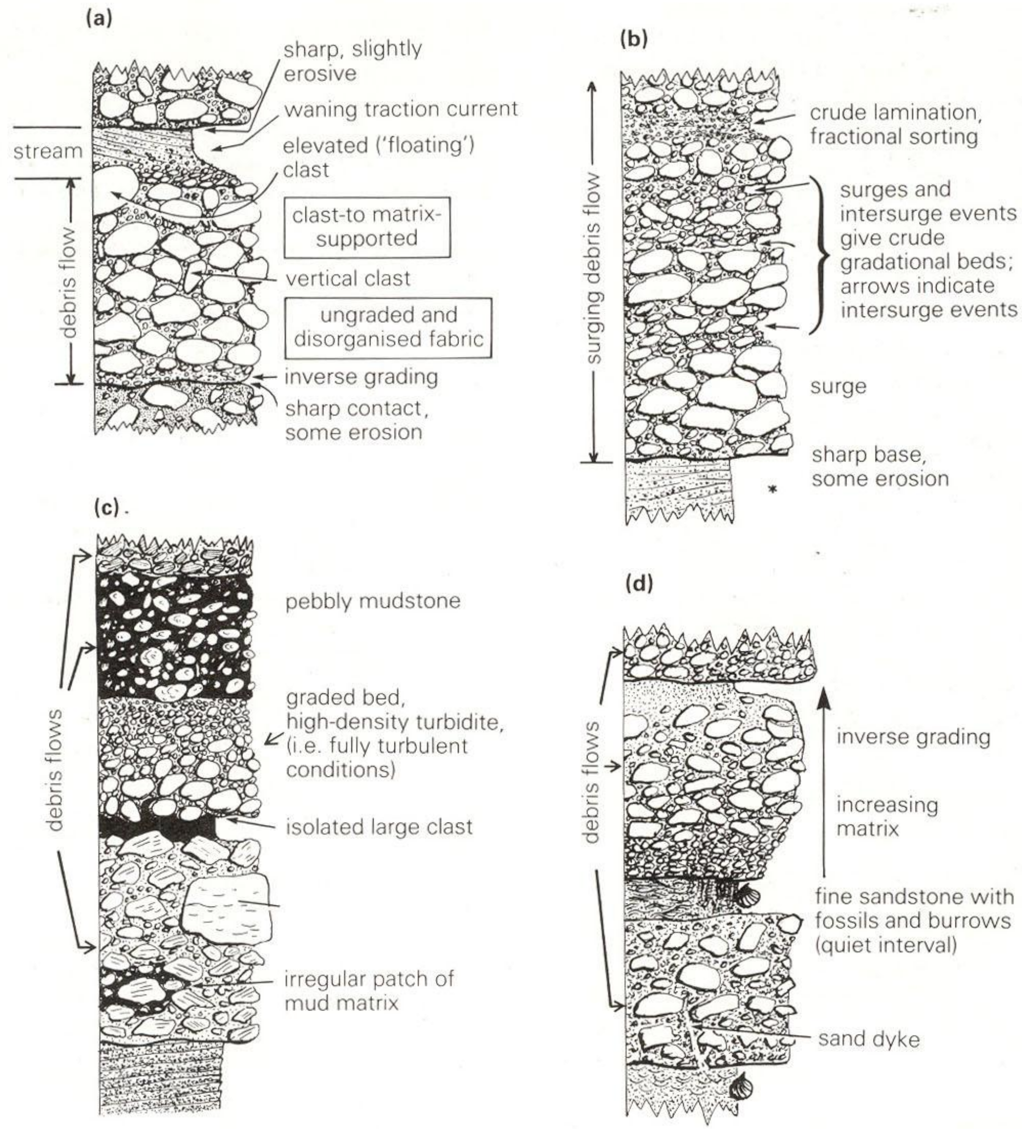


Figure 7.4 Examples of different styles of grading in conglomerates due to different depositional processes acting in a variety of typical sub-aerial (a,b) and sub-aqueous settings (c,d). Bed thickness ranges from a few decimetres to metres (after Nemec & Steel 1984, Figs 15 & 16).

Collinson & Thompson (1989), p.117

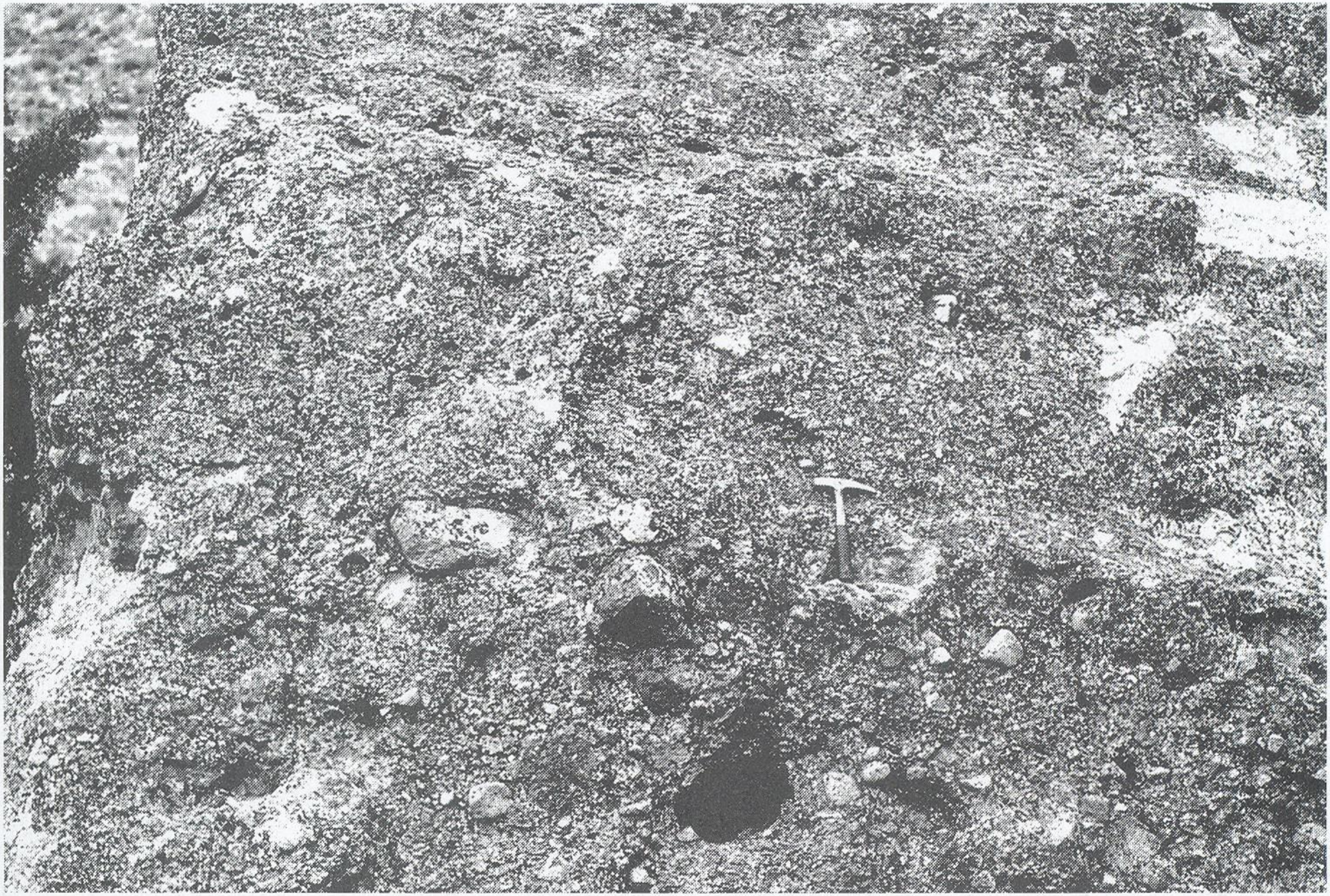


Fig. 2.10 Poorly sorted debris-flow deposits (Eocene), north-central Oregon.