Annu. Rev. Earth Planet. Sci. 1993. 21:89–114

SEDIMENT DEPOSITION FROM TURBIDITY CURRENTS

Gerard V. Middleton

Department of Geology, McMaster University, Hamilton, Ontario L8S 4M1, Canada

KEY WORDS: autosuspension, gravity flow, hydraulics, turbidite

1. INTRODUCTION

Turbidity currents are *gravity currents* (Simpson 1987; also called *density currents* by many authors, following Bell 1942) in which the excess density or unit weight is due to suspended sediment. Gravity (or density) currents are a general class of flows (also know as stratified flows) in which flow takes place because of relatively small differences in unit weight between two fluids: The gravity flow may move below, above, or between ambient fluid of different unit weight (or weights). The difference in unit weight may be due to differences in composition (for example, the flow of oil over water), in salinity, or in temperature. The flow of rivers is not normally considered a type of gravity current because the unit weight of water is several hundred times larger than that of air. The flow of cold air beneath warm air, however, is considered to be a gravity or density current. The term turbidity current is generally applied only to flows of sediment suspended in water, though strictly it applies also to flows of suspended solids in air—for example, to some atmospheric duststorms, to powder snow avalanches (Hopfinger 1983, and see *Annals of Glaciology* 13, 1989), and to phenomena such as pyroclastic flows or base surges.

Gravity currents that are driven by gravity acting on dispersed sediment in the flow were called *sediment gravity flows* by Middleton & Hampton (1973, 1976). Turbidity currents are one type of sediment gravity flow in which the sediment is held in suspension by fluid turbulence. They can be distinguished from other types (e.g. debris flows) in which sediment is

0084–6597/93/0515–0089$02.00

dispersed by other mechanisms (e.g. matrix strength, grain collisions, buoyancy, etc).

This review is concerned with the mechanics of sediment deposition from turbidity currents. Beds deposited from turbidity currents (called *turbidites*) are one of the commonest types of sedimentary rocks: They include both sands and muds, and may be of siliciclastic or other composition (e.g. carbonates, cherts). For example, the majority of sandstones in the geologic record were deposited either from rivers or from turbidity currents, and the largest sedimentary features on the modern earth are submarine fans and abyssal plains, both of which result from turbidity currents. This review will not attempt to summarize the large number of published descriptions of turbidite facies (for a recent summary see Pickering et al 1986, 1989).

The early history of the turbidity current concept has been reviewed by Walker (1973). Though Daly (1936) and Bell (1942) were pioneers, it was the experimental and field observations of Philip Kuenen that convinced geologists that many sandstones previously believed to have been deposited in shallow water were in fact deposited by turbidity currents in water hundreds or thousands of meters deep. This notion had far reaching consequences for stratigraphy, and indeed for the whole of geology. This review is dedicated to Kuenen's memory—an appreciation of his work is now available in the *Dictionary of Scientific Biography* (Bourgeois 1990).

2. HYDRAULICS

2.1 *Introduction*

Many aspects of turbidity current hydraulics are similar to those of other types of gravity flows. These have been reviewed by Turner (1979), Yih (1980), Simpson (1982, 1987), and Hopfinger (1983). The dynamics of granular flows—a class of sediment gravity flows that probably grades into turbidity currents—has been reviewed by Campbell (1990). In this section, therefore, we do not attempt to describe all aspects of gravity flow hydraulics, but discuss only the few topics that are necessary to the discussion of sediment deposition given later in the review.

2.2 *Complexity of Gravity Flows*

It must first be emphasized that gravity flows are a very complex class of flows, more complex for example than open-channel flows such as rivers (which might be considered to be merely one extreme end member of the gravity flow spectrum). No geologist would expect that rivers could deposit only a single type of sediment bed, characterized by a single suite of sedimentary structures and textures—yet many geologists have such a notion about turbidites. Turbidity currents are generally thought to be unsteady, indeed catastrophic events, generated by sediment slumping on an oversteepened subqueous slope. They flow down a subaqueous channel, erosional in its upper part, becoming depositional, with prominent subaqueous levees, near the base of slope. Further from the base of slope, the channel may disappear and the turbidity current spreads out as a flow that is unconfined, except by the topography of the sedimentary basin. As the location of the upper part of the channel (or "canyon") is fixed, and the lower depositional channels are free to migrate by avulsion, lateral migration, or braiding, such idealized flows give rise to subaqueous fans and basin-plain deposits. A single bed, whether deposited on the fan or on the plain, is expected to be size graded, and to show a sequence of structures known as the *Bouma sequence* (after the pioneering work of Bouma 1962; see Figure 1).

Such a simple model of turbidity current behavior has been surprisingly successful in accounting for many observed features of ancient turbidites, but it is now becoming clear that a better understanding is required to account for the many diverse phenomena reported from modern environments and ancient turbidite systems (Shanmugam & Moiola 1991, Hesse & Rakofsky 1992). The mechanisms for turbidity current generation are

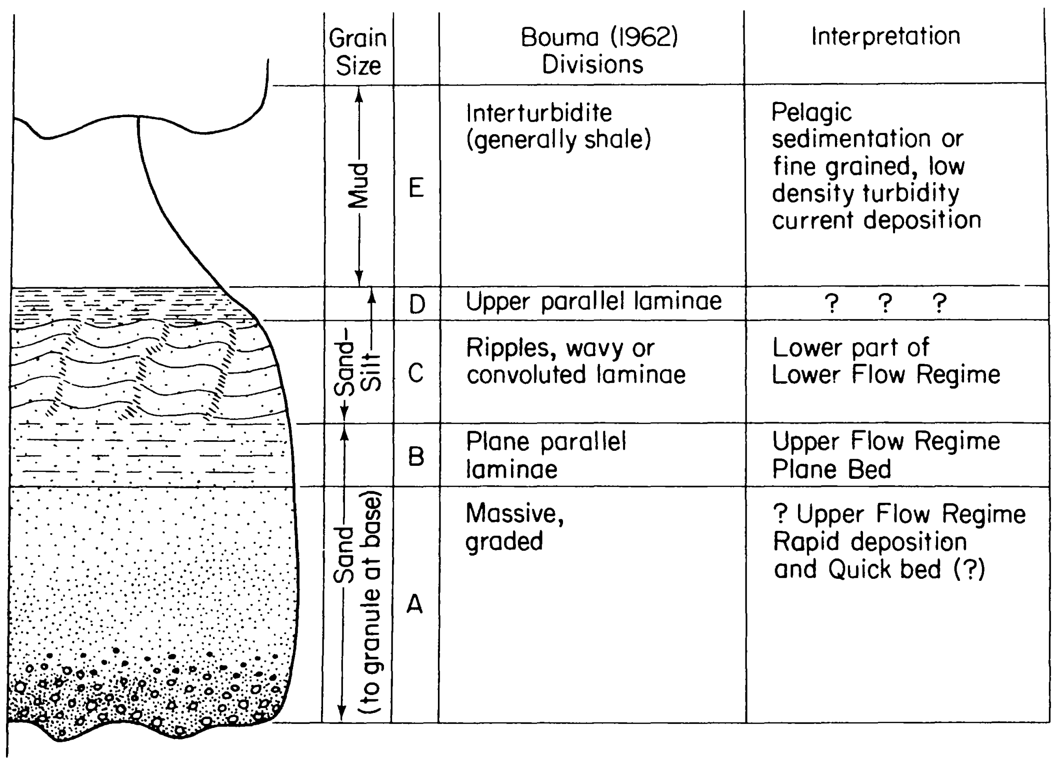

Figure 1 The Bouma sequence.

certainly diverse. Though most turbidity currents are unsteady surges, others are long-lived, more or less steady flows. Turbidity currents are capable of erosion or deposition. They span an immense range of scales from flows that are more than a hundred meters thick, carry cubic kms of sediment, and deposit the "megaturbidites" of Labaume et al (1983; Mutti et al 1984), to thin, dilute flows. They probably vary greatly in sediment concentration from "hyperconcentrations" of more than 40% by weight (Pierson & Costa 1987, Smith 1986), to "high density" turbidity currents (defined, by Kuenen 1966 and Middleton 1970, as those with more than 10% of sediment by weight) to flows with only a few parts per thousand of sediment (see below).

Like other gravity flows turbidity currents display supercritical and subcritical flow regimes, and the transition from the former to the latter can take place only through a submerged hydraulic jump (Middleton 1970, Komar 1971, Hand 1974, Weirich 1988). In rivers, such jumps take place only in small, steep tributaries: Though important in the design of hydraulic structures they have little geomorphological importance. In turbidity current systems, hydraulic jumps must take place at the base of slope, and should produce important—though presently largely unrecognized—geomorphological and sedimentological effects.

Because of the effects of buoyancy, turbidity currents are generally thicker than river flows of corresponding power: This is shown by the large size of turbidity current channels (generally over 10 m deep, even for small systems such as those in modern fjords). Channel levees are meters to tens of meters or more in thickness, and extend for hundreds of meters or more from the channels. Frequent overflow may be partly a consequence of spill from the head which forms at the front of turbidity current surges and is generally at least twice the thickness of the succeeding flow. It must also result in part from the stratification in sediment concentration and size that results from turbulent suspension of sediment, and from mixing of the finer sediment fraction from the current proper to a zone of ambient fluid that is entrained by the flow. Yet another factor is superelevation of the current surface in channel bends—transverse slopes in meandering turbidity currents are much larger than those in rivers.

Turbidity current channels seem to display the full range of channel patterns shown by fluvial channels. Meandering is well developed on some submarine fans (e.g. the Amazon fan channels described by Flood & Damuth 1987). Other large meandering, levee-bounded channels consititute a sedimentary system in themselves, which can develop independent from any submarine fan (Hesse & Rakofsky 1992). Large meandering patterns of flow are also recorded by sole marks and grain orientation in thin, apparently unchannelized turbidites (Parkash & Middleton 1970, Middleton 1970). Braiding seems to be common on the lower parts of some submarine fans, and within large channels. Though cross-bedding formed by migration of dunes is uncommon in most ancient turbidites, some modern turbidite channels are floored by large dunes (Piper et al 1988, Malinverno et al 1988) and some ancient submarine channels contain trough-cross bedded sandstones, presumably deposited by turbidity currents (Hein 1982).

Pickering & Hiscott (1985) were the first to demonstrate conclusively that some turbidite beds were deposited by a single current that reversed direction within the same basin. This phenomenon, essentially unknown in fluvial systems, has been studied experimentally by Pantin & Leeder (1987) and reported in modern basins by Muck & Underwood (1990) and Kneller et al (1991).

2.3 *Simple Hydraulic Equations*

Gravity flows, like other fluid flows, exhibit laminar and turbulent regimes. Work on laminar flows has been summarized by Harleman (1961) and Huppert (1991). The transition to the turbulent regime takes place at Reynolds numbers of less than 1000 (where the characteristic length is taken as the flow thickness). In nature, therefore, laminar flows are confined to magma chambers. An exception might be small gravity currents with very high concentrations of mud (McCave & Jones 1988). In these, the "fluid" certainly shows non-Newtonian properties, and if it is truly nonturbulent it can no longer be classified as a turbidity current. In this review, it is assumed that the flows are fully turbulent—a condition achieved in most, though not all, experimental investigations.

Based on gravity surge experiments, Middleton (1966b, 1967) divided a typical gravity current into three parts: the head, body, and tail. As first documented by Keulegan (1958) the head of a density surge has a distinct shape and hydraulics, which differ from the region behind the head. The heads of saline gravity flows were subject to very thorough experimental investigation during the 1970s and 1980s, and the results are reported by Simpson (1982, 1987). Recent numerical simulations include those by Crook & Miller (1985), Droegemeier & Wilhelmson (1985), Haase & Smith (1989a,b), and Xu et al (1992). The implications of these observations for sediment deposition from turbidity currents have been discussed by Middleton (1967), Allen (1971a), Komar (1972), and Middleton & Southard (1984).

The mass and momentum balance of the head of a gravity flow differ significantly from those of the body and tail.(See Figures 2 and 3.) In order for the head to advance it must displace the ambient fluid, which is generally at rest (for the case of a gravity flow advancing into a moving

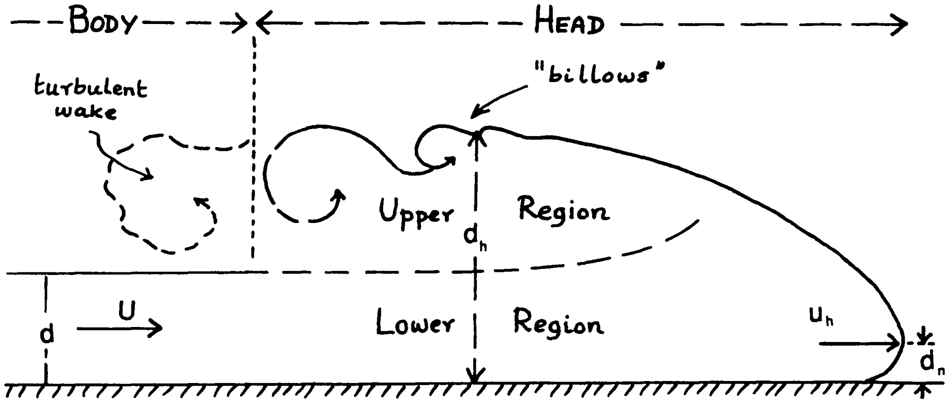

Figure 2 The structure of the head of a turbidity current.

fluid see Bühler et al 1991; for the effect of turbulence in the ambient fluid, see Linden & Simpson 1986; Noh & Fernando 1991b, 1992). Accelerating the ambient fluid produces a resistance to the flow which is larger than friction at the bed or the upper interface. Therefore the head of the current must be thicker (have more gravitational potential energy) than the current behind the head, where only frictional resistance is important. Not that friction is unimportant at the head: The bed friction produces an overhanging "nose," so that the head overrides lighter ambient fluid. This is one way that ambient fluid is mixed into the flow, and results in a gravitational instability that in turn produces the three-dimensional, lobate structure seen at the front of the flow. Though this effect should be less important for large-scale natural flows than it is in the laboratory, Simpson & Britter (1980) showed that a 10% overhang was present even in very large scale atmospheric fronts. Viscous shear at the upper surface of the head leads to the formation of Kelvin-Helmholz "billows" and ultimately to large-scale turbulent mixing at the back of the head.

The back of the head is marked by a sharp discontinuity in flow thickness. The flow pattern is similar to the flow separation and turbulent wake observed in the lee of a blunt (solid) body moving through a stationary fluid. In this case, however, the moving body is fluid, not solid, and mass is constantly lost from it to large eddies that break away from the back of the head. To achieve a constant rate of advance, this fluid must be replaced from the body behind the head, and this implies that the average speed of the flow in the body must be larger than the speed of advance of the head. How much larger depends on the rate of loss of fluid from the head, which is determined mainly by the densiometric Froude number and therefore by the slope (Middleton 1967; and see discussion below).

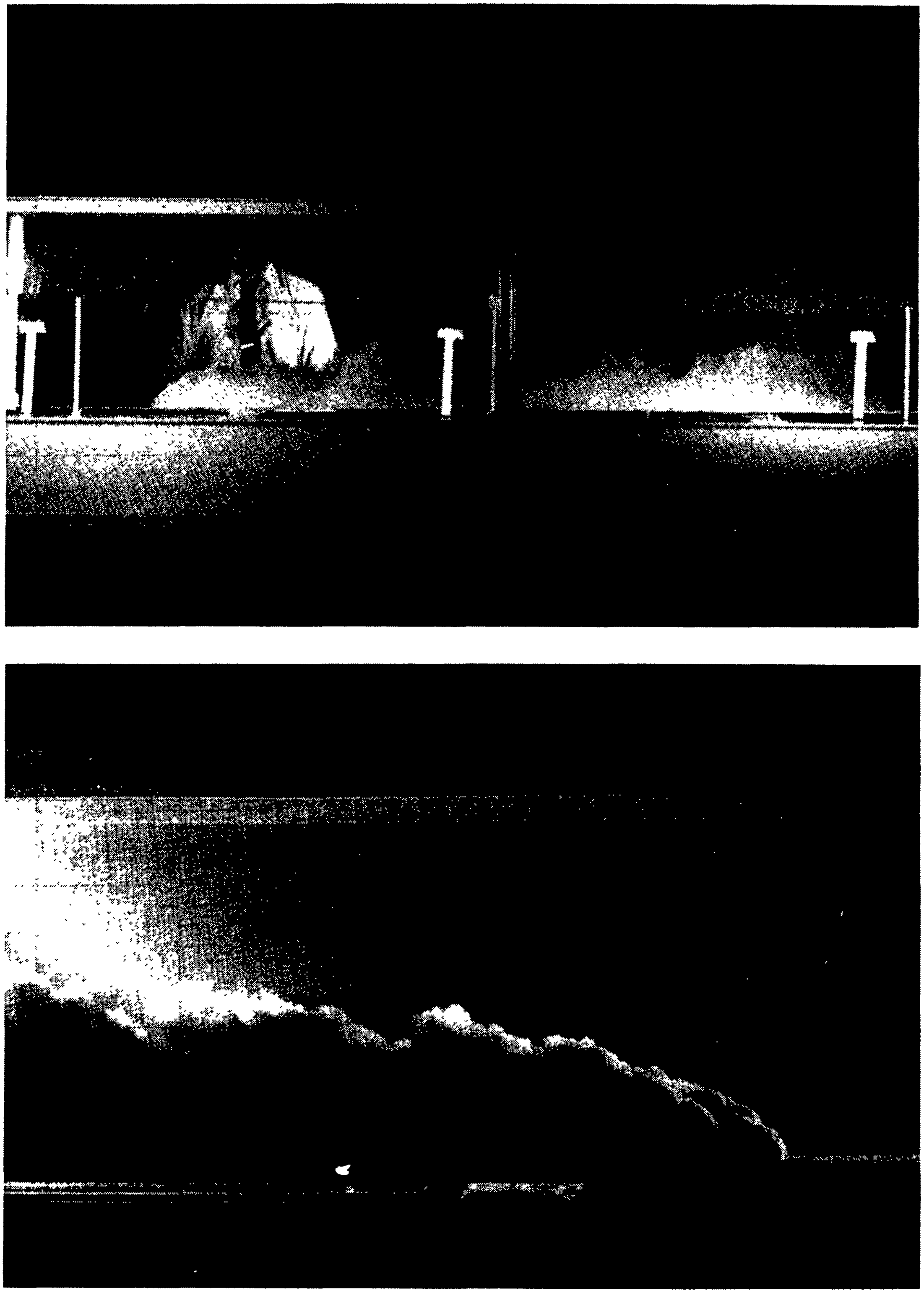

Figure 3 Experimental turbidity currents. The top photo, taken by Roger Walker in 1964, shows the author admiring one of the flows he produced in the W. M. Keck laboratory at the California Institute of Technology. The bottom photo (courtesy of Roger Walker) shows details of the head of another turbidity current.

The *densiometric Froude number* for a flow of thickness d and depth-averaged velocity U is defined as

$$Fr = \frac{U}{\sqrt{g'd}}, \tag{1}$$

where g' is the buoyancy-reduced gravitational acceleration,

$$g' = \frac{(\rho_f - \rho)}{\rho} g, \tag{2}$$

ρ_f is the bulk density of the flow, and ρ is the density of the ambient fluid.

The densiometric Froude number of the head itself can be defined as

$$Fr_h = \frac{u_h}{\sqrt{g'd_h}}, \tag{3}$$

where u_h is the speed of the head, and d_h is its thickness. Keulegan showed that, for two-dimensional saline heads advancing into a horizontal channel filled with freshwater, this number was a constant with a value of about 0.7–0.8. Subsequent experiments have shown (e.g. Simpson 1982, Hay 1983) that this value is only slightly altered by large increases in scale (Reynolds number), several degrees of bed slope, or by an increase in the depth of ambient fluid, or by replacing the saline solution with a sediment suspension (even a suspension with as much as 40% sediment by volume).

Provided that there is a sufficient volume of suspension released by the initiating event, a region will develop behind the head of a gravity surge in which there is a fairly close approximation to steady, uniform flow. Such flows are also expected if there is a steady discharge of denser fluid supplied from the source (e.g. from a river entering a lake). For the two-dimensional case, a simple force balance applied to a control volume leads to the Chézy-type equation:

$$U = \sqrt{\frac{8g'}{f_o + f_i}} \sqrt{dS}, \tag{4}$$

where U is the velocity averaged over the depth d, and f_o and f_i are the friction factors for the bottom and upper interface respectively, defined by the equations

$$\tau_o = \frac{f_o}{4} \frac{\rho_f U^2}{2}, \tag{5}$$

$$\tau_i = \frac{f_i}{4} \frac{\rho_f U^2}{2}, \tag{6}$$

where τ_o and τ_i are the shear stresses at the bottom and upper interface respectively. Notice that the ratio between the friction coefficients is equal to the ratio between the shear stresses at the bottom and upper interface. Assuming, as a first approximation, that the bulk density is uniform within the flow so that the shear stress varies linearly from the top to the bottom, the ratio between the two friction coefficients is also equal to the ratio of the thicknesses of the part of the flow below the velocity maximum (at which the average shear stress is zero) d_o and the thickness of the part of the flow above the velocity maximum d_i (Figure 4). The influence of slope on the densiometric Froude number is shown by using the Chézy equation for U in the definition

$$Fr = \sqrt{\frac{8S}{f_o + f_i}}. \tag{7}$$

This illustrates one of the problems in estimating f_i, since experimental evidence indicates that it, in turn, depends on Fr.

Hydraulic equations of the type given above have been widely used to

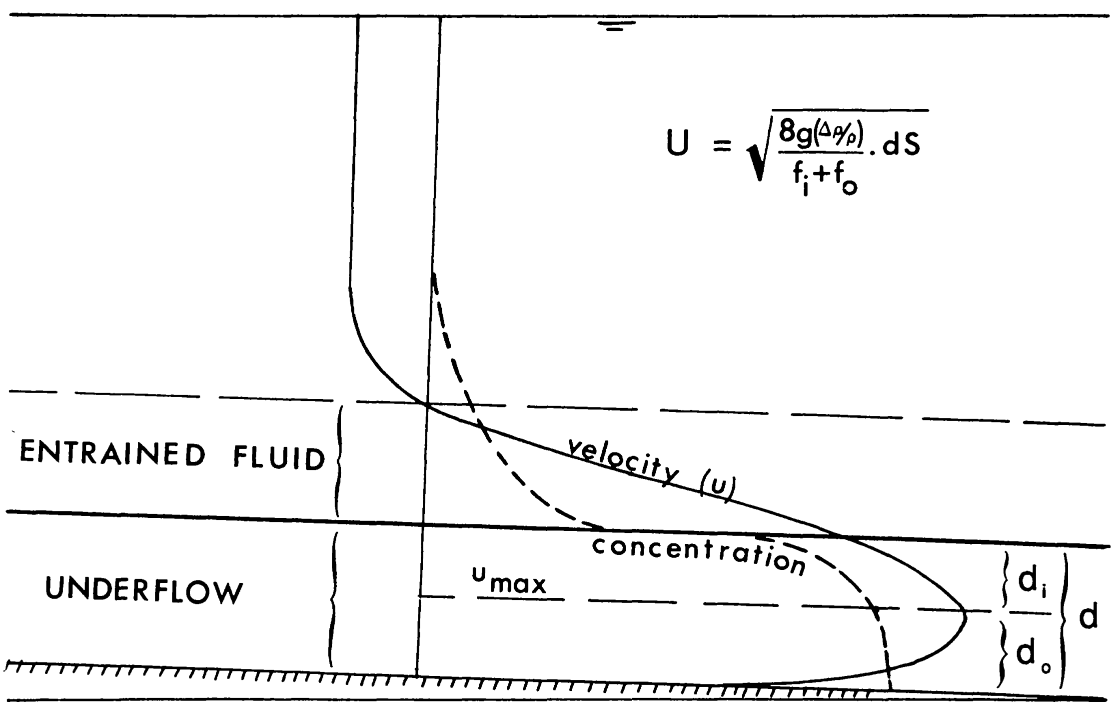

Figure 4 Uniform flow of a turbidity current. In this diagram the top of the underflow is considered to be defined by the inflection point in the sediment concentration profile. See text for discussion.

analyze the flow of turbidity currents (e.g. Kersey & Hsü 1976; Komar 1977, 1985; van Tassel 1981; Bowen et al 1984; Zeng et al 1991). Their use is, however, limited by the necessity to make educated guesses not only about the bulk density of the flow (which has generally not been measured) but also about the friction coefficients f_o and f_i. It may be expected that f_o is a function mainly of the bottom roughness, so it can be roughly estimated—though the estimate is more difficult than it is for rivers because of high sediment concentrations (see below). But f_i depends on mixing at the upper interface, and therefore on the hydraulics of the flow. It is certainly some function of the densiometric Froude number (Middleton 1966c, Fernando 1991) and is fairly small for low *Fr* and becomes large as the critical value is approached. But it is not easy to provide realistic estimates for large-scale flows. It certainly cannot be assumed that it is simply a fraction of the bottom friction, since the resistances in the two cases arise from totally different physical mechanisms and are subject to different hydraulic controls (Parker et al 1987, Normark 1989).

3. MODERN TURBIDITY CURRENTS

From the beginning, geologists have sought information about modern turbidity currents in oceans and lakes in order to understand how ancient turbidites were deposited. At first, observations were mainly of currents in reservoirs such as Lake Mead. They have been followed by a large number of other observations in reservoirs (Fan 1986; Fan & Morris 1992; Chikita 1989, 1990) and lakes (Normark 1989, Lambert et al 1976, Lambert & Giovanoli 1988, Gilbert & Shaw 1981, Smith et al 1982, Weirich 1986).

The existence of catastrophic, large-scale turbidity currents in the oceans was inferred mainly from cable-breaks, submarine topography, and the presence of sand beds with displaced shallow water fauna in deep-sea environments. Attempts to directly measure the properties of turbidity currents by monitoring currents in submarine channels were begun by Shepard in 1968 (see Shepard et al 1979, Inman et al 1976) and have continued to the present (Dengler et al 1984, Reynolds 1987). Most attempts have yielded only imperfect records, due largely to the unpredictability of the larger events and their ability to destroy installed monitoring instruments. A relatively successful program has been carried out in some fjords, where currents are, however, weaker and more continuous than those of most interest to geologists (Hay et al 1982, 1987a,b; Prior et al 1987; Zeng et al 1991; Phillips & Smith 1992).

It is impossible to give a complete summary of the observations in this review. In lakes and reservoirs, most recorded currents have been relatively continuous, extending over several days, relatively slow (less than 1 ms^{-1}), relatively thin (a few meters), and relatively low density (less than 1% by weight of sediment). They carry mainly silt and clay: The sand delivered by the source river is generally deposited in a delta at the head of the lake or reservoir. Because of the low sediment concentration, temperature may be as important in forming gravity flows as suspended sediment.

The turbidity currents in Chinese reservoirs described by Fan (1986) and Fan & Morris (1992) represent an extreme example. Suspended sediment consisted mainly of silt; concentrations were over 10% by weight in the source river, and reached 5% or more in gravity flows. Currents were up to 15 m thick and maximum velocities over 1 ms^{-1} were recorded. Fan (1986, figure 5) presents velocity and density profiles for a flow which traveled 50 km down Sanmenxia Reservoir in 1961. The slope was about 0.00025, the thickness about 2–3 m, the concentration about 6%—and fairly uniform over most of the depth, and the maximum velocity of about 0.6 ms^{-1} was located near the upper interface. This indicates there was little mixing and resistance at the upper interface ($Fr \approx 0.5$.)

Gravity flows measured in fjords are similar to those in reservoirs and lakes. They are pulsed, but may be relatively long-lived flows, though they are mostly generated by slumping at the fjord-head delta, rather than by the direct plunge of muddy inflow below the surface as in most lakes and reservoirs. In Queen Inlet (Phillips & Smith 1992) concentrations and velocities measured one meter above the bed in a 20 m deep channel were low (less than 5 ppt, and 0.1–0.2 ms^{-1} respectively), though there were a few surges with velocities of more than 1 ms^{-1}. Flow thickness is difficult to estimate, but in the July 27 flows it probably reached about 10 m, with the upper part of the flow having lower sediment concentrations (about 1 ppt) than the lowest meter (about 3 ppt). Channel slope was high (0.01), and *Fr* probably quite high (0.7)—though the significance of the number becomes dubious where there are large variations in concentration with depth. In Bute Inlet (Zeng et al 1991), current velocities were measured 4 m above the bottom, and flow thicknesses were measured roughly by the use of vane deflectors, but sediment concentrations were not measured directly. Speeds measured reached more than 3 ms^{-1} in short surges more than 30 m thick near the source but declined rapidly to less than 1 ms^{-1} 30 km down fjord. From these measurements and estimates of velocity, depth, and slope derived from the morphology of the channels, Zeng et al (1991) estimated average sediment concentrations in the range from nearly 2% by weight to 5 ppt (declining rapidly downflow).

Evidence for more powerful gravity surges in the oceans is provided by displacement of heavy instrument package anchors and a few surviving current meter records of speeds above 3 ms^{-1} (e.g. Dengler at al 1984).

But there are essentially no direct measurements of flow thicknesses, or velocity and sediment concentration profiles from such strong flows. Cable-breaks give evidence of head velocities reaching at least 19 ms^{-1} (Piper et al 1988).

4. INFERENCES FROM TURBIDITES

An alternative to measuring the properties of turbidity currents in models or in nature is to infer their behavior from the beds they deposit (*turbidites*). This has been attempted since the concept of turbidity currents was first used to explain ancient sandstone beds. A review of the early studies was given by Walker (1973), and more recent work is summarized by Allen (1982, vol. 2) and Pickering et al (1989).

An early problem was to distingish turbidites from beds deposited from other types of flows, including other types of gravity flows. This problem is still not completely solved: Observations on ancient deep-sea deposits indicate that some must have been deposited by other mechanisms, such as submarine debris flows, but that there are transitions between deposits that can safely be ascribed to deposition from turbidity currents and those possibly deposited by other types of flow. Transitions in sedimentary facies are often assumed to correspond to transitions in flow types. Middleton & Hampton (1973, 1976) and Lowe (1979, 1982) have stressed that the textures and structures observed in the deposit, formed during the final stages of flow or even shortly after deposition, do not necessarily tell us much about the nature of the flow while it was actively transporting sediment. The same problems exist for subaerial flows (Pierson & Costa 1987, Smith 1986).

The existence of a single model of the vertical sequence of textures and structures found in an "ideal" turbidite bed (the *Bouma sequence*) has tempted many geologists to propose an "ideal" sequence of hydraulic events leading to the deposition of that sequence. Very few turbidites show the complete Bouma sequence (Figure 1), however, and most fall into one of at least two categories: *proximal* or *distal* turbidites. Proximal turbidites show an erosional base, followed by a "massive" (Bouma A) division. The A division may pass up gradually into faint parallel lamination (the Bouma B division), which in turn may be capped by a weak development of current ripples (the Bouma C division). Distal turbidites show an erosional base, followed by development of distinct plane lamination (B division) and/or ripple cross-lamination (see Pickering et al 1989, for the many possible variations on this theme). Convolute lamination, originally recognized by Bouma (1962) as an alternate criterion for the C division, is formed by a gravitational instability acting on rapidly deposited sediment, and may be developed from either a plane bed or ripples. It should no longer be used to define the structural divisions.

The importance of not lumping proximal and distal turbidite types together is readily revealed by making inferences from the structures about rates of erosion or deposition. The structures were formed as different parts of a surge (head, body, tail) pass over a bed location. Many turbidites, both proximal and distal, have a sharp erosive base indicating that the head of the flow is not generally a region of sediment deposition. In some turbidites it can be inferred, either from the erosive marks themselves (Allen 1971d) or from the evidence of trace fossils preserved on the sole (Seilacher 1962, Wetzel & Aigner 1986), that several centimeters of erosion was involved, in others that there was time only for the removal of the least consolidated muddy material.

In proximal turbidites deposition of a massive interval that passes up gradationally into poorly developed plane lamination is most plausibly interpreted as due to a sudden change from erosion, to deposition from suspension which was so rapid that it did not permit sufficient grain movement by traction on the bed to form visible lamination. Strong experimental evidence for this interpretation has been described by Arnott & Hand (1989) and discussed by Allen (1991), and includes the experimental production not only of massive beds, but also of the high-angle grain imbrication characteristic of the A division of proximal turbidites.

In many distal turbidites, however, there is strong evidence that, after a period of bed erosion, deposition began slowly, with a relatively extended period of grain traction, and then accelerated with an increasing rate of sedimentation from suspension towards the upper part of the bed. The evidence consists in an upward change in the types of structures (well-developed plane lamination, to ripple cross-lamination showing stoss-side erosion, to ripple cross-lamination showing preservation of stoss-side lamination) and in a common upward increase in the rate of climb of ripple drift cross-lamination (Walker 1969; Allen 1971b,c; Ashley et al 1982). Flutes at the base of distal turbidites frequently trap grains that are much coarser than those in the rest of the bed. This indicates that the head of the surge was not only capable of erosion but also able to carry much coarser sediment than the body of the current. As recognized by Kuenen long ago, such evidence indicates that far-traveled turbidity currents have a well-developed longitudinal hydraulic structure. The coarsest sediment is either deposited upstream, or transfered from the body into the head of the flow. It is then recycled, either within the head itself, or by dropping out of eddies torn off the rear of the head, and falling into the faster moving body of the flow, where it will be carried once again into the head. It is thus possible for some coarse grains to bypass more proximal regions.

It is surely to be expected that different types of turbidity currents will deposit sediment in different parts of the same sedimentary basin. A single hydraulic model can have only limited utility.

5. EXPERIMENTAL STUDIES

In contrast to the large number of experimental studies of gravity flow hydraulics, and to the rapidly increasing number of theoretical studies of sediment deposition (see next section) there have been relatively few experimental studies of sediment deposition from turbidity currents. Since the work of Middleton (1967), the following constitutes a complete list: Riddell (1969), Tesaker (1969), Fan (1980), Lüthi (1980a,b, 1981), Pallesen (1983), Ravenne & Beghin (1983), Siegenthaler et al (1984), Siegenthaler & Bühler (1985, 1986), Bühler & Siegenthaler (1986), Parker et al (1987), Garcia & Parker (1988,1989), Garcia (1990), Scheiwiller (1986), Scheiwiller et al (1987), Laval et al (1988), Middleton & Neal (1989), Norem et al (1990), Altinakar et al (1990), Duringer et al (1991), and Kerr (1991). To this list should be added some studies of sediment deposition from fluid gravity flows that were made specifically to mimic deposition of sediment from turbidity currents. The most important are those by Ashley et al (1982) and Southard & Mackintosh (1981).

The scaling of laboratory experiments was discussed by Middleton (1966a) and Middleton & Neal (1989). Besides ensuring that the current is fully turbulent, the most important dimensionless numbers are the densiometric Froude number, and the ratio between the sediment settling velocity w and some characteristic velocity of the flow (generally U or u_h). As large turbidity currents generally have speeds in the range of 1–10 ms^{-1} and laboratory flows only reach about 0.1 ms^{-1}, this means that settling velocity should be reduced in the model by a factor of 10 to 100. To model a sand with a settling velocity of 5 cms^{-1} one should use a silt—but this raises several problems: Such sediments may exhibit cohesion, beds composed of silt are hydraulically smooth—not rough like beds of sand, and deposits produced by the experimental flows are very thin and difficult to analyze. To overcome this problem, Middleton (1967), Middleton & Neal (1989), Riddell (1969), and Garcia (1990) used sediment of sand size but reduced density (plastic or coal). Unfortunately, this also reduces the bulk density of the flow, an effect that can only be offset by raising the volume concentration. Laval et al (1988) tried mixed experiments where the density contrast was produced using dissolved salt, but sand was added to the flow to act as a tracer. Garcia & Parker (1988) studied erosion of a sand or silt bed by a saline flow.

Middleton (1967), Riddell (1969), and Middleton & Neal (1989) studied deposition of sand from surges released from a lock into a horizontal channel. One surprising aspect of these flows is that they deposit a bed of sediment that is at first almost uniform in thickness, and wedges out only where the flow nears the end of its run out. Middleton & Neal (1989) found that increasing the volume concentration increased both thickness and length of the bed in about the same proportion, and the data for several grain sizes and two different densities could be collapsed onto a single line by using the ratio between the bed thickness and the cube root of the initial volume of suspension. For unit width and 20% volume concentration the equation is:

$$\log(w/u_{\mathrm{h}}) = 1.75 + 2.0 \log[t/(LH)^{1/2}], \tag{8}$$

where L and H are the length and height of the lock from which the suspension was released. In words, the result is that the dimensionless bed thickness is proportional to the square root of the dimensionless settling velocity. Extrapolation of this relationship to much larger scales seems possible: Turbidite beds on abyssal plains (which have slopes of less than one in a thousand) have remarkably constant thickness, and application of the experimental equation yields estimates of current speed in the right range. But the result needs much more testing and is poorly understood theoretically.

The studies by Ravenne & Beghin (1983), Laval et al (1988), Lüthi (1980a,b, 1981), Garcia & Parker (1989), Duringer et al (1991), and Kerr (1991) were exploratory in nature. Laval et al (1988) studied "pure" surges (produced by using a large H/L ratio), which decelerate rapidly. Only Hauenstein & Dracos (1984), Beghin & Olagne (1991), and Lüthi have made large-scale experiments on any kind of gravity flow in which the flow was allowed to expand in the transverse direction as it flowed down a slope (in other words, experiments on a negatively buoyant wall-jet.) Hauenstein & Dracos (1984) reported that steady discharge saline inflows, on slopes less than 15°, did not show a constant angle of spreading, though a constant angle of 30° was observed by Beghin & Olagne (1991) for saline surges released onto slopes greater than 15°. In experiments using quartz silts, Lüthi (1991) found that dilution accounted for most of the density decrease of the flow, with sediment deposition being only of secondary importance. Spreading, combined with dilution by mixing, led to rapid deceleration of the flows even down relatively steep slopes and to a nearly exponential decrease in bed thickness with distance from the source.

The experiments by Altinakar et al (1990) and Garcia (1990) were concerned mainly with near-uniform flow of turbidity currents down slopes. Together with the older studies of Michon et al (1955) they provide most of the measurements of velocity and sediment concentration profiles available for such flows. Profiles were also measured by Scheiwiller (1986), but only for flows on slopes higher than 30°. Discussion of these results is deferred to the next section.

6. THEORY

6.1 *Introduction*

Most of the theoretical discussion of turbidity currents has addressed two-dimensional, subcritical, nearly-uniform flow down a slope. The basic problem is how the suspended sediment interacts with the flow, changing the turbulent structure of the flow, and therefore the mixing with ambient fluid at the upper interface and the velocity and sediment concentration profiles in the flow. These in turn determine whether the flow will grow in volume, due to incorporation of ambient fluid, and whether the sediment concentration will increase or decrease, due to erosion from or deposition on the bed. If an initial cloud of suspended sediment on a slope loses sediment and becomes more dilute, then it may either reach some quasi-equilibrium, or it may be dissipated completely (flows in the "subsiding" field of Pantin 1979; flows that fail to "ignite" in the analysis by Parker 1982). Alternatively, the flow may accelerate and erode more sediment from the bed (i.e. it may be in the "exploding" field of Pantin 1979, or it may "ignite" in the analysis of Parker 1982) until it reaches some quasi-equilibrium state where flow volume and sediment concentration change only very slowly, if at all, so long as the slope remains constant.

Models of such flows fall into two major categories: those that use vertical averaging techniques, so that at each position along the travel path, the flow is characterized by a single velocity and density (sediment concentration), and those that attempt to derive the velocity and density profiles from a turbulence model.

The first category consists of models which expand the simple Chézy-type equation (Section 2.3) to flows where the slope, depth, and other parameters vary slowly down flow. The way in which the parameters vary is a consequence either of conservation equations or of the assumptions of the model. There are numerous examples including Johnson (1962), Hopfinger & Tochon-Danguy (1977; applied to the Oahu turbidity current by Dengler & Wilde 1987), Chu et al (1979), Fan (1980), Siegenthaler & Bühler (1985), Siegenthaler & Bühler (1986), Kirwan et al (1986), Caserta et al (1990), Norem et al (1990), and the models of Pantin and Parker referenced above. The models can be classified by the number of equations involved: for example the Kirwan et al model uses a single, conservation of momentum equation, and the Johnson and Parker et al (1987) models use four equations, for conservation of momentum, sediment mass, fluid mass, and energy. Also some models assume strongly non-Newtonian materials (Dengler & Wilde 1987, Norem et al 1990) so are more appropriate for submarine debris flows than for turbidity currents.

The second category uses models of turbulent suspension and mixing to predict the evolution of velocity and density profiles down the flow path, from some assumed initial condition. These models require use of turbulence models, either implicitly or explicitly. Turbulence models make use of the time-averaged equations of motion of a turbulent fluid. These include the Reynolds stress terms, which cannot be predicted from the Navier-Stokes equations (thus giving rise to the turbulence "closure problem"). A large variety of possible models has been developed that attempt to predict these terms for practical use. The hydraulics applications have been reviewed by Rodi (1984). One simple model, proposed by Gelfenbaum (1988), is discussed below. Another model was developed by Stacey & Bowen (1988a,b). A somewhat more complex model, the k-ε model, uses equations for the total turbulent energy k, and its rate of dissipation ε to obtain closure. This model has been modified to allow for density stratification and applied to turbidity currents by Scheiwiller et al (1987), Kupusović (1989), and Eidsvik & Brørs (1989).

All of these are two-dimensonal *forward models*, i.e. they predict the evolution of a turbidity current as it flows in a channel down a varying slope, assuming a (generally constant) specified source of suspension. They do not directly address the main problem of interest to geologists: how to reconstruct the properties of the current from the observed properties of the bed that it deposits. It is possible to calculate some properties of the bed produced from most of the models, though none of the authors has actually gone through this exercise, but to simulate a given bed would require multiple iterations of the model, and the result would probably not be unique. One advantage of the equation of Middleton & Neal (1989) is that it can be used to infer the current speed directly from the measured volume, size distribution, and thickness of a given bed (provided it was deposited on a flat, confined basin floor).

For discussion, three aspects of the problem may be distinguished: (*a*) the autosuspension criterion; (*b*) velocity and concentration profiles in the lower part of the flow, and sediment exchange with the bed; and (*c*) velocity and concentration profiles in the upper part of the flow, and mixing at the upper interface.

6.2 *Autosuspension*

Bagnold (1962) argued that there should exist a condition in which the input of gravitational energy is just sufficient to sustain the turbulence necessary to keep the sediment in suspension and to overcome friction at the upper and lower interface. As it is the suspended sediment that supplies the gravitational energy to drive the flow and create the turbulence, it may be said that the sediment "suspends itself," hence the term *autosuspension*. From a simple energy balance Bagnold (1962) derived the necessary condition to be

$$w \leqslant SU. \tag{9}$$

Pantin (1979), Parker (1982), Seymour (1986), and Stacey & Bowen (1988b, 1990) proposed modifying this criterion to

$$w \leqslant eSU, \tag{10}$$

where e is an efficiency coefficient, less than unity, calculated in various ways by different authors. Southard & Mackintosh (1981) devised an experimental test for Bagnold's original criterion and could find no evidence to support it: There is a disagreement between them and other authors whether this was because their test was not sensitive enough to detect the reduced effect of a small value of e in Equation (10) or because the whole concept of autosuspension is based on a faulty concept of suspension energetics (Paola & Southard 1983).

If an autosuspension criterion does exist, field evidence examined by Seymour (1986) suggests $e < 0.1$. Unfortunately the interpretation of the scanty field data is not at all clear. In the oceans, large turbidity currents certainly travel on low slopes for hundreds of kilometers without depositing all their sediment. One may argue that this is evidence for at least a near-approach to the autosuspension condition. On the other hand, there is evidence that even large flows deposit sediment along their complete travel-path (e.g. Pilkey et al 1980, Piper et al 1988). For this to be consistent with autosuspension, one must suppose that, in regions of near constant slope, sediment deposition took place only from the tail of the flow.

6.3 *Lower Part of the Flow*

The simplest theory of sediment suspension is the diffusion theory of Rouse, based on the mixing-length theory for velocity distribution in a turbulent boundary layer (for a review see Middleton & Southard 1984). This theory assumes that the velocity profile is independent of the sediment concentration—an assumption long known to be incorrect. The problem is bad enough in rivers, where suspended sediment concentrations are generally low, but becomes critical for turbidity currents. There is a large literature which cannot be discussed here: Some recent contributors are Smith & McLean (1977), Noh & Fernando (1991a), Villaret & Trowbridge (1991), McLean (1992), and Umeyama & Gerritsen (1992).

Perhaps the simplest solution is that proposed by Smith & McLean

(1977) and tested by Gelfenbaum & Smith (1986) and again by Villaret & Trowbridge (1991) using a larger data set. These authors use a depth-dependent turbulent diffusion coefficient, as does the classical theory, but modified by the stable stratification induced by the suspended sediment profile and calculated by iteration (Villaret & Trowbridge provide direct approximation methods.)

In a slug of suspended sediment starting to move down a slope some particles will settle to the bed. The critical question is then whether or not they will remain on the bed or be eroded and taken back into suspension. The problem of entrainment of bed material into suspension (rather than into tractive motion, which is the usual "initiation of motion" problem in sediment transport) was studied by Akiyama & Fukushima (1985) and their results were used to construct a theory of turbidity currents by Akiyama & Stefan (1985) and Parker et al (1987). In these theories, as in many other theories of turbidity currents, the velocity and sediment concentration in the model is averaged over the depth, rather than derived explicitly. Parker et al (1987; see also Akiyama & Stefan 1985) argued that three equation models, based on depth-averaged conservation equations for fluid mass, sediment mass, and momentum, were inadequate because they did not take into account the turbulent energy balance. They proposed adding a fourth equation to link the sediment entrainment to the turbulent energy k rather than to the mean velocity of the flow.

6.4 *Upper Part of the Flow*

A steady gravity current flowing down a uniform slope mixes with the ambient fluid at its upper interface. This raises problems in defining exactly what constitutes the current itself: It cannot be all the fluid moving down the slope, because some of the overlying fluid is dragged along by the flow, even though it has nearly the ambient density. Mixing at the interface increases the fluid discharge downslope, so that the volume (and thus the depth) of the flow gradually increases in that direction—therefore it is not possible to have a strictly uniform gravity flow (Stacey & Bowen 1988a), and the flow depth cannot reasonably be calculated from the upslope discharge and the velocity averaged up from the bed. Most (but not all) authors have measured the depth from the bed to the inflection point of the concentration profile (Figure 4). A few authors prefer to include all of the profile that differs significantly in density from the ambient fluid: This implies by definition that mixing goes only one way. Ambient fluid may be mixed into the gravity current, but denser fluid cannot be mixed into the ambient fluid.

Factors governing the rate of mixing have been reviewed by Fernando (1991). In geophysical flows, there is always mixing, therefore no sharp interface between the denser and ambient fluid. Mixing across the zone where both the density ρ and the velocity u are functions of elevation z is therefore better related to the *gradient Richardson number*

$$Ri = \frac{-g}{\rho} \frac{(\partial \rho / \partial z)}{(\partial u / \partial z)^2} \tag{11}$$

than to the densiometric Froude number [in the case of a sharp interface, the Richardson number can be defined as $Ri = (1/Fr)^2$ (Turner 1979)]. The critical condition for stability is $Ri > 0.25$ (Miles 1990). Gelfenbaum (1988), using the results of Geyer & Smith (1987), proposed that mixing processes maintain a constant Richardson number only slightly larger than critical, and therefore a linear velocity and concentration profile in the upper part of the flow. The thickness of the upper layer can then be calculated from the critical value of the Richardson number and the velocity and sediment concentration at the velocity maximum (the top of the lower layer). The entrainment of ambient fluid into the flow, and its downslope evolution (including erosion or deposition of sediment), can be calculated by using the equations of motion, and the velocity and sediment concentrations for the lower part of the flow derived from the stratification-corrected diffusion theory, There is, however, still no generally accepted theory of entrainment in stratified shear flows (Fernando 1991).

7. CONCLUSIONS

Turbidity currents are very complex hydraulic phenomena. Those of greatest interest to geologists are large-scale catastrophic events which are extremely difficult to observe and measure in nature. Though small-scale models have been used successfully to explore some aspects of gravity flow hydraulics, there are serious limitations on their use to reproduce sedimentation from such flows directly. The complexity of the interaction between sediment suspension, turbulence, and mixing at the top of the flow extends present turbulence and numerical models up to and beyond their known limits. It seems that progress in the future will depend upon more detailed monitoring of natural flows at intermediate scales (in lakes and fjords), renewed experimentation with sediment-bearing flows at larger laboratory scales than generally attempted in the past, and further application of forward models which attempt to predict velocity and sediment concentration profiles, rather than use depth-averaging techniques. Verification of model assumptions is generally attempted only piece-by-piece, and it will be very difficult to obtain field data that is sufficiently complete to allow testing of the whole model. These models must not only be verified in large-scale experiments, but also used to predict turbidite size distribution and bed-thickness. Such predictions, though not yet attempted, could be made using several existing models.

A fundamental problem that still remains concerns the general nature of turbidity currents: Can a cloud of suspended sediment achieve some sort of equilibrium state (autosuspension), where it moves downslope without either eroding or depositing sediment? Or, are suspension clouds, once created, doomed to rapid dissipation by spreading, mixing with ambient fluid, and sediment deposition?

Though large uncertainties remain, remarkable progress has been made in the tracing and interpretation of ancient turbidites, and in understanding the basic hydraulics of gravity currents. The problems that remain, though difficult, have been vigorously attacked in the past ten years by workers trained in several different disciplines (geology, oceanography, meteorology, fluid mechanics) using techniques unknown to an earlier generation. The future seems hopeful.

ACKNOWLEDGMENTS

I thank Marcelo Garcia, Donald Lowe, Gary Parker, and Jianjun Zeng for their comments on an earlier version of the manuscript. This review was written while I was on sabbatical leave at the Department of Geological Sciences, University of Washington. My research on turbidity currents has been supported by grants from the National Science and Engineering Research Council of Canada.

Literature Cited

Akiyama. J., Fukushima, Y. 1985. Entrainment of noncohesive bed sediment into suspension. *Third Int. Symp. on River Sedimentation, Univ. Miss.*, pp. 804–13

Akiyama, J., Stefan, H. 1985. Turbidity current with erosion and deposition. *J. Hydraul. Eng.* 111: 1473–96

Allen, J. R. L. 1971a. Mixing at turbidity current heads, and its geological implications. *J. Sediment. Petrol.* 41: 97–113

Allen, J. R. L. 1971b. A theoretical and experimental study of climbing ripple cross-stratification, with a field application to the Uppsala Esker. *Geogr. Ann.* A53: 157–87

Allen, J. R. L. 1971c. Instantaneous sediment deposition rates deduced from climbing-ripple cross-lamination. *J. Geol. Soc. London* 127: 553–61

Allen, J. R. L. 1971d. Transverse erosional marks of mud and rock: their physical basis and geological significance. *Sediment. Geol.* 5: 167–384

Allen, J. R. L. 1982. *Sedimentary Structures: Their Character and Physical Basis.* New York: Elsevier Co. 2 Vol., 593 pp. and 663 pp.

Allen, J. R. L. 1991. The Bouma division A and the possible duration of turbidity currents. *J. Sediment. Petrol.* 61: 291–95

Altinakar, S., Graf, W. H., Hopfinger, E. J. 1990. Weakly depositing turbidity current on a small slope. *J. Hydraul. Res.* 28: 55–80

Arnott, R. W. C., Hand, B. M. 1989. Bedforms, primary structures and grain fabric in the presence of sediment rain. *J. Sediment. Petrol.* 59: 1062–69

Ashley, G. M., Southard, J. B., Boothroyd, J. C. 1982. Deposition of climbing-ripple beds: a flume simulation. *Sedimentology* 29: 67–79

Bagnold, R. A. 1962. Auto-suspension of transported sediment: turbidity currents. *Proc. R. Soc. London* 265A: 315–19

Beghin, P., Olagne, X. 1991. Experimental and theoretical study of the dynamics of powder snow avalanches. *Cold Reg. Sci. Technol.* 19: 317–26

Bell, H. S. 1942. Density currents as agents for transporting sediments. *J. Geol.* 50: 512–47

Bouma, A. H. 1962. *Sedimentology of Some Flysch Deposits: A Graphic Approach to Facies Interpretation*. New York: Elsevier. 168 pp.

Bourgeois, J., 1990. Kuenen, Philip Henry. In *Dictionary of Scientific Biography*, ed. F. L. Holmes, 17(Suppl. II): 509–14. New York: Scribener's

Bowen, A. J., Normark, W. R., Piper, D. J. P. 1984. Modelling of turbidity currents on Navy Submarine Fan, California Continental Borderland. *Sedimentology* 31: 169–85

Bühler, J., Wright, S. J., Kim, Y. 1991. Gravity current advancing into a flowing fluid. *J. Hydraul. Res.* 29: 243–57

Campbell, C. S. 1990. Rapid granular flows. *Annu. Rev. Fluid Mech.* 22: 57–92

Caserta, A., Mieli, E., Salusti, E. 1990. On a model of bottom erosion by dense water steady veins. *Geophys. Astrophys. Fluid Dyn.* 55: 117–35

Chikita, K. 1989. A field study on turbidity currents initiated from spring runoffs. *Water Resources Res.* 25: 257–71

Chikita, K. 1990. Sedimentation by river-induced turbidity currents: field measurements and interpretation. *Sedimentology* 37: 891–905

Chu, F. H., Pilkey, W. D., Pilkey, O. H. 1979. An analytical study of turbidity current steady flow. *Mar. Geol.* 33: 205–20

Crook, N. A., Miller, M. J. 1985. A numerical and analytical study of atmospheric undular bores. *Q. J. R. Meteorol. Soc.* 111: 225–42

Daly, R. A. 1936. Origin of submarine canyons. *Am. J. Sci.* 31: 410–20

Dengler, A. T., Wilde, P., Noda, E. K., Normark, W. R. 1984. Turbidity currents generated by Hurricane Iwa. *Geo-Mar. Lett.* 4: 5–11

Dengler, A. T., Wilde, P. 1987. Turbidity currents on steep slopes: application of an avalanche-type numeric model for ocean thermal energy conversion design. *Ocean Eng.* 14: 409–33

Droegemeier, K. K., Wilhelmson, R. B. 1985. Kelvin-Helmholtz instability in a numerically simulated thunderstorm outflow. *14th Conf. on Severe Local Storms*, Indianapolis, Ind., pp. 151–54

Duringer, P., Paicheler, J. C., Schneider, J. L. 1991. Un courant d'eau continu peut-il générer des turbidites? Résultats d'expérimentations analogiques. *Mar. Geol.* 99: 231–46

Eidsvik, K. J., Brørs, B. 1989. Self-accelerating turbidity current prediction based upon $(k-\varepsilon)$ turbulence. *Cont. Shelf Res.* 9: 617–27

Fan, J. 1980. Analysis of the sediment deposition in density currents. *Scientia Sinica* 23(4): 526–38

Fan, J. 1986. Turbid density currents in reservoirs. *Water Int.* 11: 107–16

Fan, J., Morris, G. L. 1992. Reservoir sedimentation. I: Delta and density current deposits. *J. Hydraul. Eng.* 118(3): 354–69.

Fernando, Harindra J. S. 1991. Turbulent mixing in stratified fluids. *Annu. Rev. Fluid Mech.* 23: 455–91

Flood, R. D., Damuth, J. E. 1987. Quantitative characteristics of sinuous distributary channels on the Amazon deep-sea fan. *Geol. Soc. Am. Bull.* 98: 728–38

Garcia, M. H. 1990. Depositing and eroding sediment-driven flows: turbidity currents. *Univ. Minn., St. Anthony Falls Hydraul. Lab. Proj. Rep. No.* 306. 179 pp. + 2 Append.

Garcia, M., Parker, G. 1988. Entrainment of bed sediment by density underflows. In *Hydraulic Engineering, Proc.* 1988 *Natl. Conf., Am. Soc. Civil Eng.*, ed. S. R. Abt, J. Gessler, pp. 270–75

Garcia, M., Parker, G. 1989. Experiments on hydraulic jumps in turbidity currents near a canyon-fan transition. *Science* 245: 393–96

Gelfenbaum, G., Smith, J. D. 1986. Experimental evaluation of a generalized suspended-sediment transport theory. In *Shelf Sand and Sandstones, Can. Soc. Petroleum Geol. Mem.*, ed. J. R. Knight, 2: 133–44

Gelfenbaum, G. R., 1988. *Mechanics of Steady Turbidity Currents*. PhD thesis. Univ. Wash., Seattle. 137 pp.

Geyer, W. R., Smith, J. D. 1987. Shear instability in a highly stratified estuary. *J. Phys. Oceanogr.* 17: 1668–79

Gilbert, R., Shaw, J. 1981. Sedimentation in proglacial Sunwapta lake, Alberta. *Can. J. Earth Sci.* 18: 81–93

Haase, S. P., Smith, R. K. 1989a. The numerical simulation of atmospheric gravity currents. Part I: Neutrally-stable environments. *Geophys. Astrophys. Fluid Dyn.* 46: 1–33

Haase, Sabine P., Smith, Roger K. 1989b. The numerical simulation of atmospheric gravity currents. Part II: Environments with stable layers. *Geophys. Astrophys. Fluid Dyn.* 46: 35–51

Hand, B. M. 1974. Supercritical flow in turbidites. *J. Sediment. Petrol.* 44: 637–48

Harleman, D. R. F. 1961. Stratified flow. In

Handbook of Fluid Dynamics, ed. V. L. Streeter, Chap. 26. New York: McGraw Hill

Hauenstein, W., Dracos, Th. 1984. Investigation of plunging density currents generated by inflows in lakes. *J. Hydraul. Res.* 22: 157–79

Hay, A. E. 1983. On the frontal speeds of internal gravity surges on sloping boundaries. *J. Geophys. Res.* 88(C1): 751–54

Hay, A. E. 1987a. Turbidity currents and submarine channel formation in Rupert Inlet, British Columbia, 1. Surge observations. *J. Geophys. Res.* 92(C3): 2875–81

Hay, A. E. 1987b. Turbidity currents and submarine channel formation in Rupert Inlet, British Columbia, 2. The roles of continuous and surge-type flow. *J. Geophys. Res.* 92(C3): 2883–900

Hay, A. E., Burling, R. W., Murray, J. W. 1982. Remote acoustic detection of a turbidity current surge. *Science* 217: 833–35

Hein, F. J. 1982. Depositional mechanisms of deep-sea coarse clastic sediments, Cap Enragé Formation, Quebec. *Can. J. Earth Sci.* 19: 267–87

Hesse, R., Rakofsky, A. 1992. Deep-sea channel/submarine-yazoo system of the Labrador Sea: a new deep-water facies model. *Am. Assoc. Petroleum Geol. Bull.* 76: 680–707

Hopfinger, E. J. 1983. Snow avalanche motion and related phenomena. *Annu. Rev. Fluid Mech.* 15: 47–76

Hopfinger, E. J., Tochon-Danguy, J.-C. 1977. A model study of powder-snow avalanches. *J. Glaciol.* 19: 343–56

Huppert, H. E. 1991. Buoyancy-driven motions in particle-laden fluids. In *Of Fluid Mechanics and Related Matters*, Proc. of a Symp. Honouring John Miles on his Seventieth Birthday, Scripps Inst. Oceanogr. Ref. Ser. 91-24, ed. R. Salmon, D. Betts, pp. 141–59.

Inman, D. L., Nordstrom, C. E., Flick, R. E. 1976. Currents in submarine canyons: an air-sea-land interaction. *Annu. Rev. Fluid Mech.* 8: 275–310

Johnson, M. A. 1962. Turbidity currents. *Science Prog.* (*London*) 50: 257–73

Kerr, R. C. 1991. Erosion of a stable density gradient by sedimentation-driven convection. *Nature* 353: 423–25

Kersey, D. G., Hsü, K. J. 1976. Energy relations of density-current flow: an experimental investigation. *Sedimentology* 23: 761–89

Keulegan, Garbis H. 1958. Twelfth progress report on model laws for density currents: the motion of saline fronts in still water. *U.S. Natl. Bur. Stand. Rep.* 5831, 29 pp.

Kirwan, A. D. Jr., Doyle, L. J., Bowles, W. D., Brooks, G. R. 1986. Time-dependent hydrodynamic models of turbidity currents analysed with data from the Grand Banks and Orleansville events. *J. Sediment. Petrol.* 56: 379–86

Kneller, B., Edwards, D., McCaffrey, W., Moore, R. 1991. Oblique reflection of turbidity currents. *Geology* 14: 250–52

Komar, P. D. 1971. Hydraulic jumps in turbidity currents. *Geol. Soc. Am. Bull.* 82: 1477–88

Komar, P. D. 1972. Relative significance of head and body spill from a channellized turbidity current. *Geol. Soc. Am. Bull.* 83: 1151–56

Komar, P. D. 1977. Computer simulation of turbidity current flow and the study of deep-sea channels and fan sedimentation. In *The Sea, Ideas and Observation on Progress in the Study of the Seas*: 6, *Marine Modelling*, ed. E. D. Goldberg, I. N. McCave, J. J. O'Brien, J. H. Steele, pp. 603–21

Komar, P. D. 1985. The hydraulic interpretation of turbidites from their grain sizes and sedimentary structures. *Sedimentology* 32: 395–407

Kuenen, P. H. 1966. Matrix of turbidites: experimental approach. *Sedimentology* 7: 267–97

Kupusović, T. 1989. A two dimensional model of turbulent flow applied to density currents. In *Computational Modelling and Experimental Methods in Hydraulics* (*Hydrocomp* '89), ed. Č. Maksimović, M. Radojković, pp. 169–78. New York: Elsevier

Labaume, P. , Mutti, E., Séguret, M., Roselle, J. 1983. Mégaturbidites carbonatées du bassin turbiditiques de l'Eocène inférieur de moyen sud-pyrénéen. *Bull. Soc. Géol. France Ser.* 7 25: 927–41

Lambert, A. M., Kelts, K. R., Marshall, N. F. 1976. Measurements of density underflows from Walensee, Switzerland. *Sedimentology* 23: 87–105

Lambert, A. M., Giovanoli, F. 1988. Records of riverborne turbidity currents and indications of slope failures in the Rhone delta of Lake Geneva. *Limnol. Oceanogr.* 33: 458–68

Laval, A., Cremer, M., Beghin, P., Ravenne, C. 1988. Density surges: two dimensional experiments. *Sedimentology* 35: 73–84

Linden, P. F., Simpson, J. E. 1986. Gravity-driven flows in a turbulent fluid. *J. Fluid Mech.* 172: 481–97

Lowe, D. R. 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits. In *Geology of Continental Slopes*, SEPM Spec. Publ. 27, ed. L. J. Doyle, O. H. Pilkey Jr., pp. 75–82

Lowe, D. R. 1982. Sediment gravity flows, II. Depositional models with special reference to the deposits of high-density turbidity currents. *J. Sediment. Petrol.* 52: 279–97

Lüthi, S. 1980a. Die Eigenschaften nichtkanalisierter Trübeströme: Eine experimentelle Untersuchung. *Eclogae Geol. Helv.* 73: 881–904

Lüthi, S. 1980b. Some new aspects of two-dimensional turbidity currents. *Sedimentology* 28: 97–105

Lüthi, S. 1981. Experiments on non-channelized turbidity currents and their deposits. *Mar. Geol.* 40: M59–M68

Malinverno, A. , Ryan, W. B. F., Auffret, G., Pautot, G. 1988. Sonar images of the path of recent failure events on the continental margin off Nice, France. In *Sedimentologic Consequences of Convulsive Geologic Events*, ed. H. E. Clifton, *Geol. Soc. Am. Spec. Pap.* 229: 59–75

McCave, I. N., Jones, K. P. N. 1988. Deposition of ungraded muds from high-density non-turbulent turbidity currents. *Nature* 333: 250–52

McLean, S. R. 1992. On the calculation of suspended load for noncohesive sediments. *J. Geophys. Res.* 97(C4): 5759–70

Michon, X., Goddet, J., Bonnefille, R. 1955. *Etude Théoritique et Expérimentale des Courants de Densité*. Chatou, France: Lab. Natl. Hydraul. 2 vol.

Middleton, G. V. 1966a. Small scale models of turbidity currents and the criterion for auto-suspension. *J. Sediment. Petrol.* 36: 202–8

Middleton, G. V. 1966b. Experiments on density and turbidity currents, I. Motion of the head. *Can. J. Earth Sci.* 3: 523–46

Middleton, G. V. 1966c. Experiments on density and turbidity currents, II. Uniform flow of density currents. *Can. J. Earth Sci.* 3: 627–37

Middleton, G. V. 1967. Experiments on density and turbidity currents, III. Deposition of sediment. *Can. J. Earth Sci.* 4: 475–505

Middleton, G. V. 1970. Experimental studies related to problems of flysch sedimentation. In *Flysch Sedimentology in North America*, ed. J. Lajoie, *Geol. Assoc. Can. Spec. Pap.* 7: 253–72

Middleton, G. V., Hampton, M. A. 1973. Part I. Sediment gravity flows: mechanics of flow and deposition. In *Turbidites and Deep Water Sedimentation*, ed. G. V. Middleton, A. H. Bouma, *SEPM Pac. Sec. Short Course Notes*. Anaheim, Calif: SEPM. 38p.

Middleton, G. V., Hampton, M. A. 1976. Subaqueous sediment transport and deposition by sediment gravity flows. In *Marine Sediment Transport and Environmental Management*, ed. D. J. Stanley, D. J. P. Swift, pp. 197–218. New York: Wiley

Middleton, G. V., Southard, J. B. 1984. *Mechanics of Sediment Movement*. Soc. Econ. Paleontol. Mineral. Short Course Notes 3. 401 pp. 2nd Ed.

Middleton, G. V., Neal, W. J. 1989. Experiments on the thickness of beds deposited by turbidity currents. *J. Sediment. Petrol.* 59: 297–307

Miles, J. 1990. Richardson's number revisited. In *Stratified Flows, Proc. Third Int. Symp. on Stratified Flows, Feb.* 3–5, 1987, *Pasadena CA, Am. Soc. Civil Eng.*, ed. E. J. List, G. H. Jirka, pp. 1–7

Muck, M. T., Underwood, M. B. 1990. Upslope flow of turbidity currents: a comparison among field observations, theory, and laboratory models. *Geology* 18: 54–57

Mutti, E. , Ricci Lucchi, F., Séguret, M., Zanzucchi, G. 1984. Seismoturbidites: a new group of resedimented deposits. *Mar. Geol.* 55: 103–16

Noh, Y., Fernando, H. J. S. 1991a. Dispersion of suspended particles in turbulent flow. *Phys. Fluids* A3: 1730–40

Noh, Y., Fernando, H. J. S. 1991b. Gravity current propagation along an incline in the presence of boundary mixing. *J. Geophys. Res.* 96(C7): 12,586–92

Noh, Y., Fernando, H. J. S. 1992. The motion of a buoyant cloud along an incline in the presence of boundary mixing. *J. Fluid Mech.* 235: 557–77

Norem, H. , Locat, J. , Schieldrop, B. 1990. An approach to the physics and modeling of submarine flowslides. *Mar. Geotechnol.* 9: 93–111

Normark, W. R. 1989. Observed parameters for turbidity-current flow in channels, Reserve Fan, Lake Superior. *J. Sediment. Petrol.* 59: 423–31

Pallesen, T. R. 1983. *Turbidity Currents*. Tech. Univ. Denmark, Inst. Hydrodyn. Hydraul. Eng., Ser. Pap. 32. 115 pp.

Pantin, H. M. 1979. Interaction between velocity and effective density in turbidity flow: phase-plane analysis, with criteria for autosuspension. *Mar. Geol.* 31: 59–99

Pantin, H. M., Leeder, M. R. 1987. Reverse flow in turbidity currents: the role of internal solitons. *Sedimentology* 34: 1143–55

Paola, C. , Southard, J. B. 1983. Autosuspension and the energetics of two-phase flows: reply to comments on "Experimental test of autosuspension" by J. B. Southard and M. E. Mackintosh. *Earth Surf. Processes Landf.* 8: 273–79

Parkash, B., Middleton, G. V. 1970. Downcurrent textural changes in Ordovician turbidite greywackes. *Sedimentology* 14: 259–93

Parker, G. 1982. Conditions for the ignition of catastrophically erosive turbidity currents. *Mar. Geol.* 46: 307–27

Parker, G., Garcia, M., Fukushima, Y., Yu, W. 1987. Experiments on turbidity currents over an erodible bed. *J. Hydraul. Res.* 25(1): 123–47
Phillips, A. C., Smith, N. D. 1992. Delta slope processes and turbidity currents in prodeltaic submarine channels, Queen Inlet, Glacier Bay, Alaska. *Can. J. Earth Sci.* 29: 93–101
Pickering, K. T., Hiscott, R. N. 1985. Contained (reflected) turbidity currents from the Middle Ordovician Cloridorme Formation, Quebec, Canada: an alternative to the antidune hypothesis. *Sedimentology* 32: 373–94
Pickering, K., Stow, D., Watson, M., Hiscott, R. 1986. Deep-water facies, processes and models: a review and classification scheme for modern and ancient sediments. *Earth Sci. Rev.* 23: 75–174
Pickering, K. T., Hiscott, R. N., Hein, F. J. 1989. *Deep-Marine Environments: Clastic Sedimentation and Tectonics.* London: Unwin Hyman. 416 pp.
Pierson, T. C., Costa, J. E. 1987. A rheological classification of subaerial sediment-water flows. *Geol. Soc. Am. Rev. Eng. Geol.* 7: 1–12
Pilkey, O. H., Locker, S. D., Cleary, W. J. 1980. Comparison of sand-layer geometry on flat floors of 10 modern depositional basins. *Am. Assoc. Petrol. Geol. Bull.* 64: 841–56
Piper, D. J. W., Shor A. N., Hughes Clarke, J. E. 1988. The 1929 "Grand Banks" earthquake, slump, and turbidity current. In *Sedimentologic Consequences of Convulsive Geologic Events*, ed. H. E. Clifton, *Geol. Soc. Am. Spec. Pap.* 229: 77–92
Prior, D. B., Bornhold, B. D., Wiseman, W. J. Jr., Lowe, D. R. 1987. Turbidity current activity in a British Columbia fjord. *Science* 237: 1330–33
Ravenne, C., Beghin, P. 1983. Apport des expériences en canal à l'interprétation sédimentologique des dépots de cones détritiques sous-marins. *Rev. Inst. Fr. Pétr.* 38: 279–97
Reynolds, S. 1987. A recent turbidity current event, Hueneme Fan, California: reconstruction of flow properties. *Sedimentology* 34: 129–37
Riddell, J. F. 1969. A laboratory study of suspension-effect density currents. *Can. J. Earth Sci.* 6: 231–46
Rodi, W. 1984. *Turbulence Models and their Application in Hydraulics—A State of the Art Review.* Amsterdam: Int. Assoc. Hydraul. Res. 104 pp. 2nd ed.
Scheiwiller, T. 1986. *Dynamics of powder-snow avalanches.* D. Nat. Sci. Diss. Swiss Fed. Inst. Technol., Zurich (Diss. ETH 7951). 116 pp.
Scheiwiller, T., Hutter, K., Hermann, F. 1987. Dynamics of powder snow avalanches. *Ann. Geophysicae* 5B: 569–88
Seilacher, A. 1962. Paleontological studies on turbidite sedimentation and erosion. *J. Geol.* 70: 227–34
Seymour, R. J. 1986. Nearshore auto-suspending turbidity currents. *Ocean Eng.* 13: 435–47
Shanmugam, G., Moiola, R. J. 1991. Types of submarine fan lobes: models and implications. *Am. Assoc. Petrol. Geol. Bull.* 75: 156–79
Shepard, F. P., Marshall, N. F., McLoughlin, P. A., Sullivan, G. G. 1979. *Currents in Submarine Canyons and Other Seavalleys, Am. Assoc. Petrol. Geol. Stud. Geol. no.* 8.. 173 pp.
Siegenthaler, C., Hsü, K. J., Kleboth, P. 1984. Longitudinal transport of turbidity currents—a model study of the Horgen events. *Sedimentology* 31: 187–93
Siegenthaler, C., Buehler, J. 1986. The reconstruction of the paleo-slope of turbidity currents, based on simple hydromechanical parameters of the deposit. *Acta Mech.* 63: 235–44
Siegenthaler, C. , Bühler, J. 1985. The kinematics of turbulent suspension currents (turbidity currents) on inclined boundaries. *Mar. Geol.* 64: 19–40
Simpson, J. E. 1982. Gravity currents in the laboratory, atmosphere, and ocean. *Annu. Rev. Fluid Mech.* 14: 213–34
Simpson, J. E., 1987. *Gravity Currents in the Environment and the Laboratory.* New York: Wiley. 244 pp.
Simpson, J. E., Britter, R. E. 1980. A laboratory model of an atmospheric mesofront. *Q. J. R. Meteorol. Soc.* 106: 485–500
Smith, G. A. 1986. Coarse-grained nonmarine volcaniclastic sediment: terminology and depositional process. *Geol. Soc. Am. Bull.* 97: 1–10
Smith, J. D., McLean, S. R. 1977. Spatially averaged flow over a wavy surface. *J. Geophys. Res.* 82: 1735–46
Smith, N. D., Vendl, M. A., Kennedy, S. K. 1982. Comparison of sedimentation regimes in four glacier-fed lakes of western Alberta. In *Research in Glacial, Glaciofluvial, and Glaciolacustrine Systems*, ed. R. Davidson-Arnott, W. Nickling, B. D. Fahey, pp. 203–38. Norwich, UK: Geobooks
Southard, J. B., Mackintosh, M. E. 1981. Experimental test of autosuspension. *Earth Surf. Processes Landf.* 6: 103–11
Stacey, M. W., Bowen, A. J. 1988a. The vertical structure of density and turbidity currents: theory and observations. *J. Geophys. Res.* 93: 3528–42

Stacey, M. W., Bowen, A. J. 1988b. The vertical structure of turbidity currents and a necessary condition for self-maintenance. *J. Geophys. Res.* 93: 3543–53

Stacey, M. W., Bowen, A. J. 1990. A comparison of an autosuspension criterion to field observations of five turbidity currents. *Sedimentology* 37: 1–5

Tesaker, E. 1969. *Uniform turbidity current experiments.* Thesis for Licentiatus Tech. Civ. Eng. Inst. Vassbygging. Norges Tekniske Hogskole, Trondheim, Norway

Turner, J. S. 1979. *Buoyancy Effects in Fluids.* Cambridge: Cambridge Univ. Press. 368 pp. 2nd ed.

Umeyama, M. , Gerritsen, F. 1992. Velocity distribution in unform sediment-laden flow. *J. Hydraul. Eng.* 118: 229–45

Van Tassell, J. 1981. Silver abyssal plain carbonate turbidite: flow characteristics. *J. Geol.* 89: 317–33

Villaret, C. , Trowbridge, J. H. 1991. Effects of stratification by suspended sediments on turbulent shear flows. *J. Geophys. Res.* 96(C6): 10,659–80

Walker, R. G. 1969. Geometrical analysis of ripple-drift cross-lamination. *Can. J. Earth Sci.* 6: 683–91

Walker, R. G. 1973. Mopping-up the turbidite mess. In *Evolving Concepts in Sedimentology*, ed. R. N. Ginsburg, pp. 1–37. Baltimore: Johns Hopkins Press

Weirich, F. H. 1986. The record of density-induced underflows in a glacial lake. *Sedimentology* 33: 261–77

Weirich, F. H. 1988. Field evidence for hydraulic jumps in subaqueous sediment gravity flows. *Nature* 332: 626–29

Wetzel, A., Aigner, T. 1986. Stratigraphic completeness: teired trace fossils provide a measuring stick. *Geology* 14: 234–37

Xu, Q., Zhang, F. S., Lou, G. P. 1992. Finite element solutions of free-interface density currents. *Mon. Weather Rev.* 110: 230–31

Yih, C.-S., 1980. *Stratified Flows.* New York: Academic. 418 pp.

Zeng, J. , Lowe, D. R., Prior, D. B., Wiseman, W. J. Jr., Bornhold, B. D. 1991. Flow properties of turbidity currents in Bute Inlet, British Columbia. *Sedimentology* 38: 975–96