Turbidity Currents and Their Deposits

Eckart Meiburg¹ and Ben Kneller²

¹Department of Mechanical Engineering, University of California at Santa Barbara, Santa Barbara, California 93106; email: meiburg@engineering.ucsb.edu
²Department of Geology and Petroleum Geology, University of Aberdeen, Aberdeen AB24 3UE, United Kingdom; email: b.kneller@abdn.ac.uk

Key Words
sediment transport, initiation mechanism, seafloor topography, erosion and deposition, linear stability, theoretical model

Abstract
The article surveys the current state of our understanding of turbidity currents, with an emphasis on their fluid mechanics. It highlights the significant role these currents play within the global sediment cycle, and their importance in environmental processes and in the formation of hydrocarbon reservoirs. Events and mechanisms governing the initiation of turbidity currents are reviewed, along with experimental observations and findings from field studies regarding their internal velocity and density structure. As turbidity currents propagate over the seafloor, they can trigger the evolution of a host of topographical features through the processes of deposition and erosion, such as channels, levees, and sediment waves. Potential linear instability mechanisms are discussed that may determine the spatial scales of these features. Finally, the hierarchy of available theoretical models for analyzing the dynamics of turbidity currents is outlined, ranging from dimensional analysis and integral models to both depth-averaged and depth-resolving simulation approaches.
1. INTRODUCTION

1.1. The Nature of Turbidity Currents

Turbidity currents are particle-laden, gravity-driven underflows in which the particles are largely or wholly suspended by fluid turbulence. The turbulence is typically generated by the forward motion of the current along the lower boundary of the domain, the motion in turn driven by the action of gravity on the density difference between the particle-fluid mixture and the ambient fluid. The ambient fluid is generally of similar composition to (and miscible with) the interstitial fluid, and, in most natural cases on Earth’s surface, is water. Turbidity currents are nonconservative in that they may exchange particles with a loose lower boundary (i.e., a sediment bed) by deposition or suspension, and may exchange fluid with the ambient by entrainment or detrainment. Such flows dissipate mainly through the deposition of the particles. So long as the bed gradient is large enough that the turbulence generated by the forward motion of the current is sufficient to maintain the suspension, the current is said to be autosuspending. Bagnold (1962), Pantin (1979) and others [reviewed by Pantin (2001) and Parker (1982) in a similar treatment] stressed the effects of the entrainment of bed sediment into an autosuspending current, which thusly becomes catastrophically erosive, or ignitive (see numerical treatment by Blanchette et al. 2005).

Particle concentrations are often sufficiently low (0.1%–7% by volume) that particle-particle interactions play a small or negligible role in maintaining the suspension (Bagnold 1954), and from a modeling standpoint, the Boussinesq approximation is commonly valid. Nonetheless, because of the extreme difficulty in estimating particle concentrations in natural flows in the ocean (see below), there remains considerable uncertainty—and debate—concerning particle loading in large submarine turbidity currents.

1.2. The Concept of Turbidity Currents

The recognition of dense, sediment-laden currents in nature goes back to Forel (1885), who postulated that a subaqueous canyon in Lake Geneva had been created by underflows from the Rhone River. Daly (1936) suggested a similar mechanism for the formation of submarine canyons, and the term turbidity current was apparently coined by Johnson in 1939. However, Kuenen (1938, 1951; Kuenen & Migliorini 1950) conducted the first experiments on turbidity currents, recognizing their nature and their potential importance in the transport of sediment to the deep sea (and in the formation of ancient sand layers that had previously been interpreted as shallow-water deposits). That we still know so little of the nature and properties of natural turbidity currents can be ascribed to their infrequent and unpredictable occurrence, in remote and hostile environments (water hundreds to thousands of meters deep), and their destructive nature.

1.3. Significance

In a geophysical context, turbidity currents are important as agents of sediment transport into subaqueous environments such as deep lakes and oceans and, to some extent, in the shallower seas of the continental shelves. In these situations, the particles generally consist of rock or mineral fragment eroded from the land surface, transported by rivers to the shoreline, and resedimented into deeper water by turbidity currents. Calcium-carbonate particles (mainly fragments of invertebrate shells) formed in shallow marine environments can be similarly resedimented into deeper water by turbidity currents. Indeed, turbidity currents, along with submarine landslides, are the principal means by which sediment is transported from shallower to deeper water. Transport distances range from a few hundreds of meters or less (e.g., down the submerged fronts of deltas) to
thousands of kilometers on the ocean floor [e.g., the North Atlantic Mid-Ocean Channel (Klaucke et al. 1998)].

Sediments in the deep sea and in deep lakes [e.g., Lake Baikal (Nelson et al. 1995)] are largely made up of turbidites, as the deposits of turbidity currents are known. Over periods of the order of $10^4$ to $10^6$ years, these deposits may build up into vast sediment accumulations [submarine fans and related systems (Weimer & Slatt 2007)] with volumes up to millions of km$^3$ [e.g., Bengal Fan (Curray et al. 2003)]. Ancient deposits of turbidite sand, deeply buried and compacted, form an important class of hydrocarbon reservoirs (Weimer & Slatt 2007) and the host rocks for a particular type of gold deposit (Keppie et al. 1987). Turbidity currents have also been invoked for generating banded iron formations, a type of iron-ore deposit unique to the early history of Earth (Lascelles 2007). In an environmental context, turbidity currents are responsible for much of the sedimentation in reservoirs (e.g., De Cesare et al. 2001, Fan 1986), with consequent loss of water storage capacity. In the ocean, even rather small turbidity currents may damage or destroy seafloor equipment and instrumentation (e.g., Inman et al. 1976, Khripounoff et al. 2003, Prior et al. 1987), and large currents commonly damage or remove sections of submarine cables (e.g., Dengler et al. 1984, Heezen & Ewing 1952).

This article covers initiation processes, the structure of turbidity currents as deduced from natural flows and experiments, the nature of their deposits, theoretical approaches to modeling, and some current controversies. It does not cover other types of particulate gravity currents, such as pyroclastic flows, debris flows, rock avalanches, granular flows, and snow avalanches. Various topics in gravity and turbidity-current research have previously been reviewed. First and foremost, the book by Simpson (1997) offers a beautiful and accessible introduction to the field. The chapter by Rottman & Linden (2001) reviews the basic scaling laws and force balances for idealized compositional gravity currents. Several articles by Huppert review various aspects of gravity and turbidity currents. Whereas Huppert (2000) provides a more general overview of topics related to gravity-driven geophysical flows, including the shallow-water approach for analyzing them, Huppert (1998) focuses more exclusively on box models and shallow-water equations for turbidity currents. Huppert (2006) discusses both dilute as well as concentrated particle-laden currents, along with dense granular flows. Middleton (1993) gives an elegant review of the literature on turbidity currents and their deposits, including experimental results and field data up to that time. Kneller & Buckee (2000) review experimental data and theory from a geological perspective. The recent article by Parsons et al. (2007) describes the range of sediment gravity flows in the ocean, and to some degree we take this as our starting point, although we offer a somewhat different perspective.

2. SIMPLIFIED THEORETICAL MODELS

2.1. Dimensional Analysis

Much of the elegant experimental and theoretical research carried out over the past two decades by the groups at Cambridge, University of Southern California, University of California at San Diego, and elsewhere demonstrates the ability of dimensional analysis to provide fundamental insight into the dynamics of gravity and turbidity currents [see the partial reviews by Rottman & Linden (2001) and Huppert (2006)]. First and foremost, the celebrated result that the front velocity $u_f$ of a gravity current of excess density $\Delta \rho$ in an ambient of density $\rho_0$ is proportional to the square root of the reduced gravity $g = g \Delta \rho / \rho_0$ and the front height $b$,

$$u_f \propto (gb)^{1/2}, \quad (1)$$
follows from dimensional considerations of the balance between inertial and buoyancy forces alone. The classical analysis by Benjamin (1968) finds that the proportionality factor, commonly referred to as the Froude number $Fr$ of the current, has a value of $\sqrt{2}$ for inviscid flows in deep ambients. A more recent, alternative theoretical treatment by Shin et al. (2004) yields $Fr = 1$. For environments of finite depth $H$, the experiments by Huppert & Simpson (1980) determine the dependence of $Fr$ on the ratio $b/H$.

2.2. Integral Models

For finite-volume releases such as the classical lock-exchange configuration (Simpson 1997), conceptually simple integral or box models can reproduce several aspects of experimentally observed flows (Huppert & Simpson 1980). These models neglect the entrainment of ambient fluid and assume that the released fluid will evolve in the form of constant-area rectangles, so that variations in the direction of the current are neglected. For gravity currents governed by a balance of gravitational and inertial forces, one finds that the front location evolves as $t^{1/3}$. This phase typically commences after the front has traveled $O(5-10)$ lock lengths at constant velocity (the so-called slumping phase). During the late stages of the flow, when viscous forces become important, the front location depends on time as $t^{1/5}$. These scaling laws are in agreement with the experimental observations of Huppert (1982), Huppert & Simpson (1980), Rottman & Simpson (1983), and others. Dade & Huppert (1995), as well as Gladstone & Woods (2000), apply corresponding integral models to particle-driven lock-exchange flows to obtain estimates of the current length versus time. However, these authors also point out that the respective $Fr$ values depend on whether the interstitial fluid is fresh or saline, due to differences in the current shapes and structures.

2.3. Shallow-Water Models

At the next level of complexity, one finds so-called depth-averaged or shallow-water models, first introduced for compositional gravity currents by Rottman & Simpson (1983), and later extended to turbidity currents by Bonnecaze et al. (1993) and by Parker et al. (1986). These models are reviewed in detail by Huppert (1998, 2006) and Parsons et al. (2007), so we provide only a brief summary here. The shallow-water approach typically neglects viscous forces and assumes that only small vertical accelerations are present, so that the pressure field is purely hydrostatic. At the top of the current, clear fluid is usually neither entrained nor detrained. Furthermore, the suspended phase is considered to be well mixed across the height of the current, so that the pressure field is purely hydrostatic. At the top of the current, clear fluid is usually neither entrained nor detrained. Furthermore, the suspended phase is considered to be well mixed across the height of the current, so that its volume fraction does not depend on the vertical location. This assumption may hold for very fine sediment, but it is questionable for coarser particles, or during the late stages of the flow when the decaying turbulence may no longer be fully able to distribute the particles across the entire current height. For the case of a deep ambient, the motion of the overlying fluid can be neglected, and the so-called single-layer, shallow-water equations hold. For shallow ambients, conversely, it is necessary to extend this approach by formulating a two-layer system that also accounts for the dynamics of the overlying fluid layer (Baines 1995). We note that the equations for the conservation of the mass and momentum of the fluid, and of the particle volume fraction, have to be closed by prescribing the front velocity, which is commonly accomplished on the basis of experimentally observed relationships among the current height, its reduced gravity, and its front velocity. Birman et al. (2009) employ a shallow-water model for overflow currents to shed light on the processes governing the formation of levees. Their investigation shows that the entrainment of ambient fluid plays an important role in determining the levee shape. Whereas negligible entrainment rates lead to exponentially decaying levee shapes, constant entrainment rates result in power-law
shapes. Gonzalez-Juez & Meiburg (2009) extend earlier shallow-water models by Lane-Serff et al. (1995, and references therein) to investigate gravity currents over submarine structures such as pipelines. Estimates of the maximum drag by the shallow-water model lie within 10% of high-resolution simulation results.

3. FIELD OBSERVATIONS

3.1. Natural Deposits

Erosion and deposition by turbidity currents are responsible for many of the features seen on the modern seafloor. Erosional features range from gulleys on the upper continental slope, a few tens of meters deep and hundreds of meters wide (see Hall et al. 2008, and references therein), to submarine canyons several kilometers wide and hundreds of meters deep (e.g., Inman et al. 1976, Weimer & Slatt 2007). Depositional features include laterally extensive, sheet-like deposits of the abyssal plains and also submarine fans, which are self-organized systems in many ways analogous to river deltas, similarly variable in form, and ranging from a few kilometers to several thousand kilometers across. Channels within these systems, tens to thousands of kilometers in length, often have levees resembling those of river channels, formed analogously by overspill from the channel onto the adjacent seafloor; deposition within the channel and on the levees often results in elevation of the channel-levee system above the surrounding fan surface (Normark et al. 1997). The sediment bodies at the termini of these channels are typically lobate with extents of a few square kilometers to a few hundred square kilometers (e.g., Deptuck et al. 2008, and references therein) and are largely deposited from unconfined flows, although some are apparently channelized to their margins (Twichell et al. 1995). The most continuous deposits often occur within bathymetrically confined regions on the seafloor small enough that turbidity currents may reach the confining topography (Gervais et al. 2006). Various bed forms similar to those produced by unidirectional flow in shallow water may be produced by turbidity currents, especially within channels. Larger-scale sediment waves may also be generated, especially where turbidity currents pass over topographic inflections such as the crests of submarine levees or the base of the continental slope (where they may be associated with plunge pools (Lee et al. 2002)), generating fields of sediment waves with heights of tens of meters, wavelengths $O(1)$ km, and crests oriented perpendicular to the flow.

Our knowledge of these systems is based largely on observations on the modern seafloor, from side-looking sonar images; multibeam bathymetry surveys; coring, shallow high-resolution seismic surveys; and increasingly from industrial seismic surveys, especially those used to assess seafloor hazards to drilling. Outcrops of ancient turbidites have generated numerous qualitative models, and much a posteriori reasoning about the nature of the flows responsible (e.g., Mulder & Alexander 2001). In a fluid mechanical context, they offer the prospect of providing benchmarking data for future numerical simulations.

3.2. Data From Natural Flows: Scale, Dynamics, and Flow Structure

Few turbidity currents in the ocean have provided much evidence of the nature of the flows themselves, and much of that evidence has until recently been indirect. Most widely known is the event of November 18, 1929, off the Grand Banks of Newfoundland (Piper et al. 1999, and references therein). This followed a 7.2-magnitude earthquake beneath the upper continental slope at 500 to 700 m water depth, accompanied by numerous seafloor failures and submarine cable breaks in the epicentral region; a number of cable breaks occurred in sequence down the continental slope over the ensuing 13 h. The total failed volume was perhaps 100 km$^3$, consisting
mainly of silt and clay. Some of the failed material transformed into turbidity currents flowing down the slope valleys, which eroded some 50 to 100 km$^3$ of sand that had accumulated over the past 10,000 years or so. Increasing wavelengths of bed forms down the first 100 km of the valley indicate an accelerating (ignitive) current, despite a decrease in gradient from 8$^\circ$ to 1$^\circ$. Below approximately 4700 m water depth, the bed is depositional, probably triggered by the radial expansion of the flow as it began to exit the valley; on the northern part of the abyssal plain, the deposit (mainly of fine sand) is $>1$ m thick, extending 450 km and becoming thinner and finer rapidly at its margins. The deposit covers an area roughly the size of Texas, with a volume estimated as 150 to 175 km$^3$, of which perhaps only 10 km$^3$ is mud (fine silt and clay); the missing mud was probably carried away by deep-ocean circulation. The maximum front velocity, estimated from the timing of cable breaks, was approximately 19 m s$^{-1}$. Indirect estimates of maximum flow thickness range from 300 to 400 m. This suggests overall Reynolds numbers of $O(10^9–10^{10})$ on the slope.

An event involving the failure of at least $8 \times 10^6$ m$^3$ of land-fill material occurred near the mouth of the Var River SE, France, in 1979 (Dan et al. 2007). This generated a turbidity current $O(10^3)$ m thick that severed submarine cables; the first, located 95 km from the source, was cut 3 h 45 min after the initial failure (indicating an average head speed of 7.4 ms$^{-1}$), and the second cable, situated at 122 km, was cut after 8 h (1.74 ms$^{-1}$, over a 0.15$^\circ$ slope) (Piper & Savoye 1993). Turbidity currents on the Congo-Zaire submarine fan have been inferred from cable breaks to occur every 1 or 2 years (Khripounoff et al. 2003, and references therein). Recent direct observations recorded a flow through a submarine fan channel at 4000 m water depth, carrying sand and plant debris, attaining velocities $>1.2$ ms$^{-1}$ at 150 m above the channel floor, and overflowing onto the surrounding sea bed at least 18 km from the channel. A cloud of suspended mud persisted at the site for several months. The current was unrelated to flooding in the Congo River. Frequent turbidity currents occur in Bute Inlet, a British Columbia fjord, associated with late spring to summer floods (Prior et al. 1987). Maximum velocities are $>3.35$ ms$^{-1}$ measured 4 m above the fjord floor, with coarse sand suspended at heights of at least 6 to 7.5 m and total flow thicknesses of more than 30 m. The currents flow at least 25 km along the fjord, and possibly as far as 40 to 50 km, over bottom slopes of generally less than 1$^\circ$.

The most complete picture of any marine currents to date comes from Monterey Canyon, California. Xu et al. (2004) measured vertical profiles of downstream velocity for four flows over the space of 1 year, at three locations down the canyon (1450 m, 2837 m, and 3223 m water depth). Two of the four flows were storm-generated; none was seismically triggered. Peak velocities (averaged over 1 h) were from approximately 0.5 to 2 ms$^{-1}$. Flow thickness increased downcanyon, and the height of the velocity maximum decreased downcanyon. The flows persisted for several hours each, but the duration of peak flow decreased downcanyon and became more surge-like. Measurements in lakes and reservoirs have been made by, among others, Best et al. (2005), Chikita (1989), Gould (1951), and Normark (1989). Normark and Best et al. both noted the development of pulsing flow with periods of a few minutes, despite steady inflow conditions.

4. FLOW STRUCTURE
Turbidity currents can be differentiated into a front region (or head) and body (Figure 1b). As shown above, the rate of advance of the front is found to be virtually independent of the lower boundary slope. The motion of the fluid behind the head can be approximated with a modified form of the Chezy equation for flow in open channels, using reduced gravity (Middleton 1993), and is slope dependent. Consequently, the buoyancy flux (Turner 1973) into the head increases with increasing slope, with a concomitant effect on mixing (see below). Finite-volume releases
4.1. Velocity and Turbulence

The vertical structure of density and turbidity currents is analyzed by Stacey & Brown (1988). The mean velocity structure of turbidity currents consists of an inner (near-wall) region with a positive velocity gradient, similar to a conventional turbulent boundary layer, and an outer region (shear layer), generally 5 to 10 (or more) times thicker than the inner region, with a negative velocity gradient and shear stress of opposite sign (Figure 1b). The velocity structure has been compared with that of plane turbulent wall jets (Gray et al. 2005; Kneller & Buckee 2000; Leeder et al. 2005; Parker et al. 1987, and references therein). However, the use of $y^{1/2}$ (the height at which the downstream velocity falls to half its maximum), advocated by Launder & Rodi (1983) as a characteristic length scale for wall jets, yields a rather unsatisfactory collapse of velocity profiles compiled from different contexts (Kneller & Buckee 2000; see also Gray et al. 2005), suggesting that the shear layer deviates from a Gaussian profile. In fact for some currents, the shear layer profile is close to linear (Ellison & Turner 1959, Xu et al. 2004).
Turbulent kinetic energy profiles in turbidity currents are similar to those of saline gravity currents, being close to zero at the height of the downstream velocity maximum (Gray et al. 2005, Kneller & Buckee 2000, Kneller et al. 1999, Leeder et al. 2005), reflecting the dominance of turbulence production by shear related to the mean streamwise velocity profile.

4.2. Density
In the case of simple turbidity currents (i.e., those in which the interstitial and ambient fluids are of the same density), the density structure is determined by the distribution of suspended sediment. Many authors have shown that the highest suspended-sediment concentrations (and commonly the steepest gradients in concentration) occur immediately above the bed (Figure 1; see review in Kneller & Buckee 2000). Parker et al. (1987) found the vertical distribution of suspended sediment to have a much weaker dependence on the ratio of the shear velocity to the sediment fall velocity than is the case in open-channel suspensions. Baas et al. (2005) showed that suspended-sediment distribution is highly unsteady and considered it to be controlled largely by the ratio of particle-settling velocity to the upward-directed components of local turbulent velocity associated with coherent flow structures. Leeder et al. (2005) proposed a criterion for the maintenance of suspension based on the ratio of maximum vertical turbulent stress to the immersed weight of suspended load over unit bed area.

4.3. Entrainment
Parsons & Garcia (1998) have shown that the entrainment of ambient fluid into the head of gravity currents is dependent on a Reynolds number based on the cube root of the buoyancy flux into the head. Entrainment into the body is a function of the overall Richardson number (Ellison & Turner 1959). Based on experiments with turbidity currents, Parker et al. (1987) proposed an empirical relation

\[ \varepsilon_w = 0.075 / (1 + 718 R_i^{1.4})^{0.5}, \]  

where

\[ R_i = \frac{b g}{u^2} \left( \frac{\partial \bar{w}}{\partial z} \right), \]

in which \( \varepsilon_w \) is the entrainment coefficient (entrainment velocity normalized by mean downstream velocity), \( b \) is the current height, and \( u \) is the mean streamwise velocity.

Evidence of very low entrainment rates on the ocean floor (Birman et al. 2009, Srivatsan et al. 2004), borne out by extremely long run-out distances of channelized flows, suggests high Richardson numbers and thus subcritical flow on the low gradients of basin floors (in contrast to flows on continental slopes (Parsons et al. 2007)). This argues for stable density stratification in the shear layer, i.e., gradient Richardson numbers, \( R_i \), sufficiently above the critical value of 0.25 to suppress mixing:

\[ R_{ig} = \left( \frac{g}{\partial \bar{w}/\partial z} \right) \left( \frac{\partial \bar{u}/\partial z}{} \right)^2, \]

where \( z \) is the vertical coordinate. Various authors have investigated the flow of turbidity currents into confining topography, where the flow thickness and topographic height control the interaction (Brunt et al. 2004, Lamb et al. 2004) and also the effects of gradient changes on flow behavior and deposition (Garcia & Parker 1993), concluding that hydraulic jumps need not occur (Gray et al. 2005), but where they do, they may generate upstream facing steps (Kostic & Parker 2006).
effects of reversing buoyancy were reviewed by Kneller & Buckee (2000, and references therein),
and Al-Musallami & Al-Ja’aidi (2008) recently investigated the turbulence structure of lofting
flows.

In summary, the above observations of a complex internal flow structure suggest that high-
resolution simulations of turbidity currents can provide insight beyond that gained from simplified
theoretical models.

5. DEPTH-RESOLVING NUMERICAL SIMULATIONS

Over the past decade, large-scale, depth-resolving simulations have begun to contribute to our
understanding of gravity and turbidity currents. Perhaps the first highly resolved direct numerical
simulation (DNS) of compositional gravity currents was conducted by Härtel et al. (2000) for
the lock-exchange configuration (see also Cantero et al. 2007, Ooi et al. 2007). More recently,
highly resolved simulations have also been conducted for particulate gravity currents. Toward
this end, Felix (2002) introduced a two-dimensional Reynolds-averaged model for a boundary-
layer approximation of the Navier-Stokes equations. He employs this model to simulate several
large-scale historical turbidity currents, such as the Bute Inlet and Grand Banks flows. A similar
approach is taken by Kassem & Imran (2004) and by Huang et al. (2005), whose simulations are
reviewed in detail by Parsons et al. (2007). These authors employ a finite-volume model on a
grid that is recalculated after every time step to allow for the temporal variation of the bottom
topography in response to erosion and deposition. This is a costly approach; in general, it may be
more promising to employ a grid that does not change with time and to represent the evolving
bottom topography via an immersed boundary approach (Mittal & Iaccarino 2005).

To avoid the uncertainties associated with determining empirical constants in Reynolds-
averaged Navier-Stokes models, Necker et al. (2002) explore much smaller, laboratory-scale flows
in the lock-exchange configuration via highly resolved, three-dimensional DNS (Figure 2b). These
authors consider dilute distributions of particles with negligible inertia that are smaller
than the smallest length scales of the buoyancy-induced fluid motion. The suspended phase is
described in an Eulerian fashion, via a convection-diffusion equation for the local particle number
density.

They observed that in three-dimensionally evolving currents, particles sediment out more
rapidly than in their two-dimensional counterparts, which points to the important role played by
spanwise instabilities. Regarding the final deposit profile, they observed excellent agreement with
De Rooij & Dalziel’s (1998) corresponding laboratory data. Such high-resolution DNS simulations
can be interrogated for quantitative information that is not readily accessible experimentally, such

![Figure 2](image-url)

**Figure 2**

(a) Experimental turbidity-current front in the laboratory, showing the overhanging nose that corresponds to the height of the streamwise velocity maximum. Scale in centimeters. Figure taken from Baas et al. 2005. (b) Direct numerical simulation of a turbidity current. The current structure is visualized by an isosurface of the concentration field. Figure taken from Necker et al. 2002.
as energy budgets. Surprisingly, the authors noted that over a wide range of parameters, roughly half the initial potential energy is lost in the small-scale Stokes flows around the sedimenting particles, so it is not available for convective transport and mixing. Although Necker et al.’s (2002) simulations do not account for erosion and resuspension, they demonstrate that the largest shear stresses are exerted on the bottom walls initially by the large-scale, spanwise Kelvin-Helmholtz rollers and later by the lobe and cleft structures at the front.

In a follow-up study, Necker et al. (2005) analyzed the differences between shallow- and deep-water flows with regard to the energy budgets and mixing behavior. Blanchette et al. (2006, 2005) further extended this line of work to eroding and resuspending turbidity currents, based on the experimentally measured relationship (Garcia & Parker 1993) among particle flux, bed shear stress, settling velocity, and particle Reynolds number. They observed that particles eroded over the length of the current are transferred to the current head, where they can lead to an acceleration of the flow, thus increasing the local bed shear stress and potentially rendering the current self-sustaining. Despite the availability of experimental correlations, the detailed mechanisms by which a current detaches individual grains from a sediment bed are still poorly understood.

Boegman & Ivey (2009) argue that not only is the shear stress at the bed surface crucial for lifting particles from the bed, but so is the structure of the flow field above. In addition, the coupling between coherent structures within the turbidity current and corresponding features in the porous sediment bed below gives rise to interesting questions that are currently wide open. Some hope derives from recent advances in the computational modeling of flows with many suspended particles (Pan et al. 2002, and references therein), which hopefully will soon allow for detailed simulations of the detachment process. Ferry & Balachandar (2001) developed an extension of the numerical methodology for treating dilute suspensions containing particles with weak inertia by means of an expansion of the particle velocity field in the dimensionless particle response time (Stokes number). This allows for an Eulerian treatment of the particle velocity field that is able to capture such effects as sedimenting-particle ejection from vortex cores and particle trapping in stretched vortices (Marcu & Meiburg 1996, Marcu et al. 1995, Martin & Meiburg 1994, Raju & Meiburg 1995). Recently, Cantero et al. (2008) applied this approach to turbidity currents. Although the above simulations account for suspended sediment only, bed-load transport is known to play an important role as well in many flows, especially in determining the character of the final deposit. In a recent investigation, Schmeeckle & Nelson (2003) developed a computational model for bed-load transport by tracking large numbers of individual particles based on a variant of the equation derived by Maxey & Riley (1983). Interactions among particles, including collisions, are modeled as well. Keeping in mind the fairly restrictive limitations under which the Maxey-Riley equation holds, it may be attractive to base future detailed models on full Navier-Stokes simulations of many-particle suspensions. Simulations along these lines open up the interesting opportunity to obtain direct information about a variety of properties of the sediment bed, among them the spatially varying distribution of particle sizes, porosity, and permeability. Such information will be extremely valuable in building hydrocarbon reservoir models.

A practical concern lies in the potentially destructive impact of gravity and turbidity currents on submarine installations such as pipelines and well heads. Gonzalez-Juez et al. (2009b) performed two-dimensional Navier-Stokes simulations for such flows around circular cylinders mounted above a wall. Their simulations confirm the experimentally observed impact, transient and quasisteady stages (Ermanyuk & Gavrilo 2005), and provide insight into the mechanisms linking flow structures to unsteady lift and drag forces. The investigations by Gonzalez-Juez et al. (2009a) and Gonzalez-Juez & Meiburg (2009) extend this line of work to three-dimensional flows and rectangular shapes, as well as bottom shear stress and scour information.
Frequently, there exists considerable uncertainty regarding the formulation of realistic initial conditions in numerical simulations of turbidity currents. This represents the motivation for taking a closer look at the mechanisms responsible for triggering such flows.

6. INITIATION MECHANISMS

6.1. Sediment Failure

The initiation of turbidity currents depends on the formation of a sediment suspension. Since the 1950s, it has been recognized that turbidity currents can be initiated by sediment failures on the slope (e.g., Gorsline et al. 2000, Heezen & Ewing 1952); this occurs through the dilution and transformation of the resulting submarine landslides or debris flows (Hampton 1972, Normark & Piper 1991, Parsons et al. 2007). More recent work has evaluated the mechanisms of this transition, which occurs either by shearing or detachment of material from the surface of the debris flow, or by the initiation of turbulence within the body of the flow, which depends on a critical ratio of dynamic stress to shear strength, in turn dependent on the proportion and type of clay present (Felix & Peakall 2006, Marr et al. 2001, Mohrig & Marr 2003). Many such submarine failures are initiated by earthquakes (Gutierrez-Pastor et al. 2009, Heezen & Ewing 1952), but in some cases they simply result from deposition on a slope, leading to oversteepening and failure (Girardclos et al. 2007).

6.2. Rivers, Flood, and Storms

The generation of turbidity currents has also been attributed to rivers in flood. Suspended-sediment concentrations in river outflows are typically up to a few kilograms per m$^3$, falling to a few grams per m$^3$ in the far field (e.g., Zaire River (Eisma & Kalf 1984), Amazon River (Rockwell Geyer & Kineke 1995)). In plumes generated by river outflows into freshwater lakes, the contribution of suspended sediment to the (negative) buoyancy is of the same order as that due to temperature differences, and river-generated underflows (so-called hyperpycnal flows) are common (e.g., De Cesare et al. 2001, Lambert & Giovanoli 1988), especially during floods, when suspended-sediment concentrations are high (Nash 1994). River plumes discharging into the ocean are typically positively buoyant because the density difference due to salinity normally greatly exceeds the density contribution due to suspended sediment. Sedimentation from such plumes may generate turbidity currents in one of two ways.

First, sediment may settle convectively from the plume at rates up to two orders of magnitude higher than Stokes settling velocities (McCool & Parsons 2001) and may generate a bottom-propagating turbidity current (Maxworthy 1999; Parsons et al. 2001, 2007; see below); simple scaling relations predict sediment settling velocities in agreement with those of sediment from a natural river plume (Eel River, California (McCool & Parsons 2001)). Maxworthy (1999) found the surface current to behave quite similarly to a nonparticulate current of equal density. In certain parameter ranges, however, the evolution of the current is strongly affected by the loss of particles and interstitial fluid at its lower boundary, which translates into a loss of momentum and an effective retarding stress. After an initial constant-velocity phase, the current begins to slow down as a result of this loss of momentum, until it comes to a complete stop. Simultaneously, vigorous plumes containing particles and interstitial and ambient fluid develop at the underside of the current. Upon reaching the floor of the experimental tank, they form a secondary turbidity current propagating horizontally along this wall. This secondary current still contains some of the interstitial fluid that was dragged downward by the particles. This interstitial fluid is subsequently...
released from the secondary turbidity current through upward-moving plumes. After having lost most of its particular matter, the surface current eventually becomes sufficiently buoyant to start moving forward again. Maxworthy presents scaling arguments for the early, intermediate, and late stages of the flow.

A second mechanism involves the resuspension of muddy sediment lost from the plume at or close to the river mouth due to flow expansion and the rapid decay of bottom-generated turbulence where the plume detaches from the sea bed at a saline front; this may be combined with flocculation of clays on contact with saltwater, leading to rapid sedimentation and the formation of fluid muds [dense suspensions with >10 kg m\(^{-3}\) of sediment (e.g., Kineke & Sternberg 1995)]. The slopes of most continental shelves are too low to sustain autosuspension (Wright & Friedrichs 2006), but tidal currents (Ogston et al. 2008, Wright et al. 1990) or waves (Traykovski et al. 2000, Warrick et al. 2008) may generate turbulence sufficient to maintain this sediment in suspension, or to resuspend it after a short period of residence (hours to months) on the continental shelf (Palanques et al. 2006a, Warrick et al. 2008); once suspended, it moves across the low gradient of the shelf as a hybrid gravity current, to be redeposited further out on the shelf, or eventually to find its way into deeper water (Wright & Friedrichs 2006). These mechanisms have recently been extensively reviewed by Parsons et al. (2007).

However, the high suspended-sediment loads developed during river floods may occasionally lead to river discharges whose bulk density exceeds that of coastal waters (approximately 1025–1030 kg m\(^{3}\)), producing a sediment-laden underflow. Such events have been recorded historically, most notably in Taiwan (Dadson et al. 2005, Milliman & Kao 2005), and implied elsewhere (Mulder et al. 2003). Their frequency has been predicted on the basis of historic data for river discharge and suspended-sediment load (Mulder & Syvitski 1995). With the present hydrologic regime and sea level, they are restricted to a few rivers with large ranges in discharge, which drain elevated and/or easily erodible catchments; this excludes the world’s largest rivers.

Except where discharging directly into a canyon head, high suspended-sediment concentrations in rivers are generally insufficient on their own to generate hyperpycnal currents (Warrick et al. 2008, Wright et al. 1990). Elsewhere, a contribution from waves or tides is necessary to maintain the sediment in suspension or resuspend it as it traverses the shelf (Palanques et al. 2006a). Even where wave action is a result of the same storm that generated the flood, the cross-shelf transit leads to a delay of hours to days before the turbidity current can be detected in the canyon. This is so even in locations where the river mouth is only a few kilometers from the canyon head [e.g., Salinas and Santa Clara rivers, California (Warrick et al. 2008, Xu et al. 2004)]; nonetheless, the interstitial water in the current is typically warmer and less saline than the ambient seawater, implying that river water is involved, and the cross-shelf gravity flow has maintained some integrity.

The generation of hyperpycnal currents has occasionally been detected where the river outflow discharges directly into the head of a submarine canyon (Kuehl et al. 2004), a situation more common at times of lowered sea level (most recently during Pleistocene glacial periods) when rivers discharge directly to the top of the continental slope. The transfer of sediment to the deep sea by turbidity currents, generated by this and other mechanisms, is held to have been far more frequent during such sea-level low stands (Blum & Hattier-Womack 2009, Weimer & Slatt 2007).

Storms are implicated in other mechanisms of turbidity-current generation, on coastlines with both wide and narrow shelves. Downwelling is a phenomenon associated with strong onshore winds that produce a setup (a piling up of water at the shoreline) that results in a deep offshore counterflow; in summer, when seawater is stratified, downwelling is inhibited by buoyancy, but in winter, when water is well mixed and of homogeneous density, winter storm-induced downwelling flow (occurring every few years) may suspend sufficient sediment to become gravity driven and autosuspending on steeper slopes (Palanques et al. 2006a). Downwelling may also be a mechanism for sediment transport.
by which tropical storms generate turbidity currents (Dengler et al. 1984). Sustained cold winter winds can also generate cold dense water on the shelf that cascades down the slope, flushing sediment out of submarine canyons en route, in events lasting from a few hours to a week [e.g., Gulf of Lions (Canals et al. 2006)].

Yet another mechanism may occur when canyons cut close to the shoreline, where wind and wave setup, possibly combined with standing edge waves (all of which are associated with high winds and waves), produce oscillations with a dominant downcanyon component; when sand is present in the canyon head (supplied by long-shore currents), these oscillations culminate in an energetic downcanyon current. Inman et al. (1976) described this situation in detail for Scripps Canyon, California. Sediment in canyon heads or gullies on the upper slope that acts as a fuel for turbidity currents is commonly supplied by wind- and tide-driven currents, to be subsequently remobilized by turbidity currents. Mastbergen & Van Der Berg (2003) suggested that the progressive retreat of steep failure-generated slopes in fine sands (a process known as breaching) may be instrumental in the generation of sustained turbidity currents during the flushing of canyons.

6.3. Other Mechanisms

Other, nonmeteorological, events on land may also be linked to the formation of turbidity currents. Earthquake-triggered subaerial landslides can introduce large quantities of sediment into river systems (Dasdon et al. 2005). The breaching of glacially dammed lakes may also generate turbidites via catastrophic floods (Brunner et al. 1999). Volcanically triggered subglacial lake breakouts (jökulhlaups) constitute a special case of such events, which may also generate turbidity currents when they enter the ocean (Geirsdóttir et al. 2000). Volcanic eruptions can generate turbidity currents directly when pyroclastic flows enter the sea (Trofimovs et al. 2008) or indirectly when ash falls and pyroclastic flows introduce large quantities of ash into river systems; eruptions are often accompanied by high rainfall associated with eruption column convection, leading to floods with extremely high suspended-sediment discharges that form hyperpycnal flows on entering the ocean (Nelson et al. 1988).

Anthropogenic turbidity currents include the effects of mine tailings being dumped into Lake Superior (Normark 1989); landfill, such as may have contributed to the Vår turbidity current in 1979 (Dan et al. 2007); the dumping of dredged material at a canyon head (Xu et al. 2004); and trawling along a canyon wall (Palanques et al. 2006b).

7. TURBIDITY CURRENT/SEDIMENT BED INTERACTION

Because the channels and gullies created by turbidity currents play an important role as pathways for sediment transport down the continental slope, it is desirable to obtain insight into the processes underlying their formation (Parsons et al. 2007). Interestingly, gullies, channels, sediment waves, and other features on the seafloor frequently appear in straight, evenly spaced patterns, which suggests the presence of an underlying, coupled hydrodynamic/sediment-driven instability. The hypothesis of an instability mechanism at the heart of submarine channel inception has spawned a number of investigations employing depth-averaged flow and sediment transport models, starting with the classical work of Smith & Bretherton (1972) (see also Hall et al. 2008, Parsons et al. 2007, and references therein).

A disadvantage of depth-averaged approaches in this regard lies in their inability to capture the detailed interaction between the sediment bed and the three-dimensional flow structures above. Specifically, potential coupling mechanisms between the spanwise and vertical velocity
components, on the one hand, and the erosion process, on the other hand, cannot be explored with this approach. Colombini & Parker (1995) showed that such coupling mechanisms are important with regard to the formation of longitudinal topographical features by bed-load transport. A related experimental investigation was conducted by Wang & Cheng (2005). Colombini & Parker (1995) further elaborated on this concept with a view toward generating small-amplitude streak features of a few grain diameters.

Hall et al.'s (2008) recent investigation aimed to explore the importance of two-way coupling mechanisms between transverse turbidity-current flow structures and suspended sediment for the formation of submarine gullies and channels. Toward this end, the authors conducted a linear stability analysis based on the full three-dimensional Navier-Stokes equations, rather than depth-averaged equations. They identified a conceptually simple and physically intuitive stability criterion that states that, for instability to occur, the suspended-sediment concentration of the base flow needs to decay more slowly away from the sediment bed than does the shear stress inside the current. Under such conditions, an upward protrusion of the sediment bed will find itself in an environment where erosion decays more quickly than sedimentation, so that it will keep growing. The authors demonstrated that this destabilizing effect of the base flow is modulated by the stabilizing perturbation of the suspended-sediment concentration, and by the shear stress due to a secondary flow structure in the form of counter-rotating streamwise vortices. As pointed out by Nielsen & Teakle (2004), measurements in river flows over bed forms typically show sediment diffusivities that are larger than the eddy viscosities, so that the conditions for instability are satisfied. For a representative current height of $O(10-100)$ m, the linear stability analysis provides the most-amplified wavelength in the range of 250–2500 m, which is consistent with field observations reported in the literature. The above Navier-Stokes-based analysis could serve as a starting point for a secondary instability analysis to gain insight into the frequently observed meandering evolution of submarine channels [see also Imran et al.'s (1999) nonlinear model and Yu et al.'s (2006) experiments, which are the first to report channelization by turbidity currents and mudflows at the laboratory scale].

Interestingly, the base flow instability mechanism identified by Hall et al. (2008) should also apply to the formation of streamwise sediment waves by turbidity currents and bottom flows carrying suspended sediment. This hypothesis is borne out by Hall's linear stability analysis (B. Hall, personal communication). Their results confirm Flood's (1988) earlier analysis linking the formation of sediment waves to the presence of internal waves in the density-stratified region above the sediment bed [see also the analysis of lee waves by Queney (1948)]. For supercritical flows over an erodible bed, Parker & Izumi (2000) proposed an alternative mechanism for the generation of streamwise periodic structures, so-called cyclic steps. Their analysis, based on the shallow-water equations, demonstrates the evolution of slowly upstream migrating features, each of which is associated with a head cut and a related hydraulic jump. Taki & Parker (2005) presented laboratory experiments demonstrating this mechanism, whereas Sun & Parker (2005) discussed corresponding nonlinear shallow-water simulations.

8. OUTLOOK AND OPEN QUESTIONS

Experimental investigations of turbidity currents are inevitably limited by scale. Because experiments at any but the smallest scales involve the use of common fluids for practical reasons, it is generally not possible to maintain all the dimensionless parameters within ranges appropriate to modeling large-scale flows in the environment. Typically one is restricted to considering limited aspects of whole-flow behavior, in which the relevant parameters can be maintained within a critical range [e.g., above the threshold values of the Reynolds number where similarity applies (Parsons...
ANRV400-FL42-07 ARI 13 November 2009 12:7

& Garcia 1998]) and others are relaxed. Nonetheless, the use of large facilities offers a scope to better define the limits to similarity. Comparable issues arise with the scaling of particles and their settling velocity; the use of smaller grains may produce nonscaled surface and electrostatic effects, whereas the use of larger low-density grains requires higher fractional concentrations and distorts the scaling relations between particles and turbulent length scales. Problems similarly arise with nonscalable bed forms such as ripples that are commonly generated in the laboratory. Nonetheless, experiments do offer considerable scope for the verification of numerical simulations, especially in areas such as sediment erosion mechanisms.

On the modeling side, there are substantial challenges waiting to be addressed as well. Deeper insight into the initiation process by which a slope failure evolves into a turbidity current is required to formulate suitable initial and boundary conditions for the early stages of the flow. Here the main difficulties lie in understanding the mechanisms that govern the initial fluidization of the sediment bed, a process that involves the interaction of densely packed particles that may or may not be cohesive with the interstitial fluid. Toward this end, it may be helpful to incorporate recent advances from the field of granular flows (Forterre & Pouliquen 2008, Huppert 2006, Lajeunesse et al. 2004, and references therein). Even during the later stages of the flow, the boundary layer of the turbidity current right above the bed can involve dense particle concentrations, so that particle-particle interactions cannot be neglected. To a first order, the effects may be captured by allowing both the effective viscosity of the suspension and the particle-settling velocity to depend on the local volume fraction of the particles. However, the true dynamics frequently will be substantially more complex, involving the interaction of suspended load and bed load, the exchange of particles between the current and the bed, and possibly non-Boussinesq flow effects. Specifically, the development of advanced erosion models will be highly beneficial for improving the fidelity of numerical simulations. Once the particles are in suspension, their interaction with the fluid turbulence still involves open questions. Whereas both experimental (Aliseda et al. 2002, and references therein) and computational (Bosse et al. 2006, and references therein) investigations of particles in homogeneous turbulence suggest that particle settling should be enhanced by two-way coupling effects, turbidity-current experiments and field observations generally indicate that particles are kept in suspension by the turbulence for long times, which allows turbidity currents to travel over very long distances.

An understanding of the interaction of gravity and turbidity currents with the background stratification and the related internal wave fields is just beginning to emerge (Maxworthy et al. 2002). Similarly, the ability of turbidity currents to form topographical features on the seafloor, and their subsequent interaction with these features, should motivate further linear stability investigations and nonlinear numerical simulations. Investigations along these lines should shed light on such issues as the meandering of channels, and the dynamics of turbidity currents propagating through seafloor topography, where substantial reflection and wave generation may occur. Similarly, the interaction of turbidity currents with a free surface represents a relevant research direction, due to the documented ability of near-shore slope failures and the ensuing turbidity currents to generate tsunamis (Dan et al. 2007).

Given that many hydrocarbon reservoirs consist of turbidity-current deposits, it will be attractive to couple the flow simulation to a realistic substrate model for the sediment bed that accounts for spatially varying distributions of particle size, porosity, and permeability. These properties could then feed into a reservoir model that, in turn, would form the basis of subsequent porous-media flow simulations. Addressing the above goals will require the integration of complementary research approaches from the fields of geology and fluid mechanics, involving field observations and measurements, laboratory experiments, the development of fundamental models, high-resolution simulations, and linear stability analysis.
FUTURE DIRECTIONS

1. Improved erosion models need to be developed that can be employed in numerical simulations.
2. Researchers will have to undertake high-resolution numerical simulations that track large numbers of individual particles to gain insight into the influence of particle-particle interactions.
3. The coupling between the evolution of the turbidity current and that of the underlying substrate needs to be explored.

DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

ACKNOWLEDGMENTS

E.M. thanks Prof. Tony Maxworthy for introducing him to the field of gravity and turbidity currents and his Ph.D. students, postdocs, and collaborators for their contributions, especially Vineet Birman, Francois Blanchette, Peter Burns, Mike Glinsky, Esteban Gonzalez-Juez, Brendon Hall, Carlos Hartel, Leonhard Kleiser, Chris Lerch, Lutz Lesshafft, Paul Linden, James Martin, Mohamad Nasr, Frieder Necker, and Moshe Strauss. Funding for this work has been provided by NASA, the National Science Foundation, BHP Billiton Petroleum, and BG Group. E.M. furthermore gratefully acknowledges the hospitality of Prof. Greg Ivey and the Geophysical Fluid Dynamics group at the University of Western Australia, and support by the Gledden Foundation, during an extended visit that allowed him to focus on the writing of this article. B.K. acknowledges the generous support of BG Group and interactions with many Ph.D. students, collaborators, and colleagues, notably Clare Buckee, Bill McCaffrey, Jeff Peakall, Maarten Felix, Henry Pantin, and Mike Leeder. Inevitably in an article of this type there are omissions. Some are a result of choices forced by space limitations; others are mere oversights. To those whose work is not adequately acknowledged, we offer our apologies.

LITERATURE CITED


<table>
<thead>
<tr>
<th>Year</th>
<th>Authors</th>
<th>Title</th>
<th>Journal/Citation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>Kuenen PH.</td>
<td>Density currents in connection with the problem of submarine canyons.</td>
<td><em>Geol. Mag.</em> 75:241–49</td>
</tr>
<tr>
<td>1951</td>
<td>Kuenen PH, Migliorini CI.</td>
<td>Turbidity currents as a cause of graded bedding.</td>
<td><em>J. Geol.</em> 58:91–127</td>
</tr>
</tbody>
</table>


www.annualreviews.org • Turbidity Currents and Their Deposits 155


