The structure and fluid mechanics of turbidity currents: a review of some recent studies and their geological implications

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ABSTRACT
The literature on the structure and behaviour of gravity currents is reviewed, with emphasis on some recent studies, and with particular attention to turbidity currents, though reference is also made to comparable behaviour in pyroclastic flows. Questions of definition are discussed, in particular the distinction between dense currents, which may deposit en masse, and more dilute currents. High-density dispersions may exist as a discrete, independently moving layer beneath a more dilute flow, as the basal part of a continuous density distribution or possibly as a transient depositional layer. Existing theory appears inadequate to explain the behaviour of some high-density dispersions. Surge-type currents are contrasted with quasi-steady currents, which may be generated by a variety of mechanisms including direct feed by rivers in flood. Such fluvially generated currents provide one means of generating currents with reversing buoyancy. Geologically significant turbidity currents are impractical for direct study owing to their large scale and (often) destructive nature. Small-scale laboratory currents offer a wealth of insights into turbidity current behaviour. This paper summarizes recent experimental studies that focus on the physical structure of gravity currents, with emphasis on the velocity and turbulence structure, the vertical density distribution and the stability of stratification. Preliminary quantification of the turbulence structure (including controls on turbulent entrainment, turbulent kinetic energy, Reynolds stresses and turbulence production) has been facilitated by recent technological developments that have allowed the measurement of instantaneous fluctuations in both velocity and concentration. Laboratory models, however, generally involve substantial simplification, and require compromises in some parameters to achieve adequate scaling of the parameters of most interest. Mathematical modelling also provides important insights into turbidity current behaviour. We discuss various approaches to modelling, ranging from simple hydraulic equations to systems of partial differential equations that explicitly treat conservation of momentum, fluid and sediment mass, and turbulent kinetic energy. The application for which the model is designed (i.e. to calculate mean head velocity or to create an instantaneous two-dimensional contour plot of downstream velocity in a current) determines the complexity of the mathematical model required. The behaviour of suspension currents around topography is complex and depends upon the relative height of the topography, and upon the density and velocity structure of the current. Many interactions with topography are well described by the internal Froude number, Fr. Both reflection and deflection of currents may occur on the upstream side of topography, depending upon Fr. On the downstream side of topography, flow separation, lee waves or hydraulic jumps may occur.

Keywords Froude number, gravity current, hyperpycnal, Richardson number, stratification, topography, turbulence.

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INTRODUCTION

Density currents are produced where gravity acts upon a density difference between one fluid and another (as such they should probably be called buoyancy currents). In the case of suspension currents, such as turbidity and pyroclastic currents, the density excess is provided by suspended solid particles. Such currents are major agents of sediment transport on land, in lakes, seas and oceans. They are important not only for the potential environmental hazards they pose (reservoir sedimentation, submarine cable breakage, pollutant dispersal and volcanic hazards; see reviews by Middleton, 1993, and Simpson, 1997) but also because turbidite systems are volumetrically the most significant clastic accumulations in the deep sea (Normark et al., 1993), and include the world’s largest sedimentary bodies (e.g. the Bengal Fan, Bouma et al., 1985). Turbidite systems also form a significant component of the stratigraphic record, and include many of the world’s most important hydrocarbon reservoirs (Weimer & Link, 1991).

This review focuses mainly on selected aspects of turbidity current structure and behaviour, and emphasizes some recent studies. In particular, we discuss the density, velocity and turbulence structure of turbidity currents, attempts to measure and model them, and the behaviour of turbidity currents interacting with topography. We also highlight some of the controversies surrounding turbidity current behaviour. Many of the observations and arguments (and indeed debates) that we report are equally valid for other types of suspension current such as pyroclastic currents. We build upon the review of sediment deposition by turbidity currents by Middleton (1993), which reviews much of the turbidity current literature to that point, and also upon that of Simpson (1997), which deals with gravity currents in general. A recent review by Huppert (1998) on quantitative modelling of suspension currents emphasizes the application of box models (see below) to simple laboratory systems and more complex environmental currents. Pyroclastic flows and surges have recently been reviewed by Freundt & Bursik (1998) and Wohletz (1998), respectively.

Definition of turbidity currents and pyroclastic currents

Turbidity currents are part of a continuum of sediment-gravity flows (Middleton & Hampton, 1976). Turbidity currents are traditionally defined as those sediment-gravity flows in which sediment is suspended by fluid turbulence (Sanders, 1965; Middleton & Hampton, 1973; Lowe, 1982; Middleton, 1993; Simpson, 1997). However, the term ‘turbidity current’ was adopted to describe a natural phenomenon whose exact nature is often unclear, and in which the sediment support mechanism(s) is commonly impossible to determine in nature. Moreover, a variety of sediment support mechanisms (e.g. turbulence, hindered settling, dispersive pressure owing to grain interactions) may operate simultaneously within a sediment gravity flow. Defining suspension currents by specifying the grain support mechanism is thus problematic (see discussion below and in Mulder et al., 1997a) as it may confuse essentially descriptive terminology (e.g. ‘turbidity current’) with interpretations or assertions of physical process (e.g. ‘turbulent’) that cannot yet be supported by observation in all cases. Nonetheless, it is probable that turbulence is the primary or sole grain support mechanism at least in dilute currents, i.e. those with fractional grain concentrations of the order of a few per cent or less (the exact concentration being dependent on the balance of factors that determine the flow Reynolds number).

Definitions are further complicated by a still very incomplete understanding of the turbulence structure of suspension currents, and the considerable confusion in the geological literature between the term turbulent (i.e. disturbed by eddies) and turbid (i.e. opaque with sediment); McCave & Jones’s (1988) description of nonturbulent turbidity currents does not necessarily imply a contradiction. We define a suspension current as ‘flow induced by the action of gravity upon a (fluidal) turbid mixture of fluid and (suspended) sediment, by virtue of the density difference between the mixture and the ambient fluid’. Within this definition, it is accepted that suspension (the ‘holding up’ of grains above the bed) may involve grain support mechanisms other than fluid turbulence, i.e. it is broader in meaning than ‘turbulent suspension’. A turbidity current is a suspension current in which the interstitial fluid is a liquid (generally water); a pyroclastic current is one in which the interstitial fluid is gas. We use the terms ‘density current’ and ‘gravity current’ interchangeably to describe any buoyancy-driven current whether the density difference is due to particles or not.

The role of turbulence in suspension currents during both transport and deposition has been the

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source of considerable debate in both the sedimentological and volcanological communities. The argument has focused on whether particulate material (sediment or pyroclasts) is deposited from high-concentration currents en masse or grain-by-grain, and has tended to link the state of the current during deposition (and thus the resulting depositional style) with the longer-term transport regime that preceded it. One view is that suspension currents of relatively high concentration exist as non-Newtonian, essentially plug-like, and perhaps nonturbulent masses, supported largely by matrix strength and/or grain-to-grain interactions (e.g. Shanmugam et al., 1995; Shanmugam, 1996; Miller & Smith, 1977; Wilson, 1985, 1997). These non-Newtonian currents may be overlain by an independently moving dilute cloud, from which they are separated by a more-or-less discrete interface (e.g. Postma et al., 1988; Mulder et al., 1997; see also Mohrig et al., 1998). According to this view, the deposit of the basal layer may represent a ‘frozen’ current, formed more or less instantaneously by frictional or cohesive immobilisation.

Another view holds that the transport regime is more expanded, and, even at relatively high particle concentrations, only the basal part of the current has low levels of turbulence; in these cases there is a continuous, rather than stepped, density profile (e.g. Druitt, 1992; Branney & Kokelaar, 1992; Kneller & Branney, 1995; Hiscott et al., 1997). In this view, the transport regime is dominated by turbulent suspension, and deposition occurs by progressive aggradation. Much debate has focused on the use of geological criteria to differentiate between the deposits of different process regimes. The uncertainties involved in this approach are evident from the fact that contrasting models have been applied to the interpretation of the same deposits, including massive deep-water sands (e.g. Jackfork Group; Shanmugam & Moiola, 1995; Slatt et al., 1997) and ignimbrites (e.g. Taupo ignimbrite; Wilson, 1997; Dade & Huppert, 1996, 1997). There may, in nature, be a continuum between the two types of transport regime, though it is harder to envisage a continuum between progressive and en masse deposition. At least some high-concentration, debris-flow-like depositional regimes arise by reconcentration during collapse of a more dilute current (e.g. Vrolijk & Southard, 1997), possibly during lofting of buoyant interstitial fluid (see below). Thus a transient, late-stage, debris-flow-like transport regime may arise during collapse of a current in which the long-range transport regime was more dilute, and perhaps dominated by fluid turbulence.

The empirical relation for the effective viscosity of clay-free dispersions of sand was given by Davidson et al. (1977) as:

$$\frac{\mu_e}{\mu} = (1 - 1.35C)^{-2.5}$$

where $\mu_e$ is the apparent viscosity of the clay-free dispersion, $\mu$ is the molecular viscosity of water and $C$ is the fractional grain concentration. This relation suggests that even at concentrations of 45% by volume (approaching the maximum concentration for a fluidized dispersion of roughly uniform grain size), the effective viscosity is about an order of magnitude greater than that of water. Utilizing the criterion for turbulence in Newtonian fluids of:

$$Re = \frac{(\rho_e U d)}{\mu} = 2000$$

(where $\rho_e$ is the mean current density, $U$ is the depth-averaged current velocity and $d$ is the current thickness), a high-concentration (45% by volume), noncohesive particulate current ('cohesionless debris flow') should rapidly become fully turbulent when the product of velocity and thickness exceeds about $3 \times 10^{-2} \text{m}^2\text{s}^{-1}$. Moreover, recent experimental work suggests that a clay content of several per cent may be necessary to induce debris-flow-like behaviour in sandy dispersions moving under the influence of gravity (Marr et al., in press). Thus, on the face of it, long-range (many kilometres) nonturbulent transport of cohesionless dispersions seems unlikely on purely mechanical grounds. However, Hallworth & Huppert (1998), reporting experimental results from high-concentration particle-driven gravity currents, describe behaviour in suspensions whose initial volume fraction exceeds 0.3 that is qualitatively different from that in suspensions of lower concentration. These flows formed rapidly moving, densely compacted currents that came to an abrupt halt, forming deposits of fairly constant thickness with an abrupt ‘snout’. The dense flow was overlain by an independently moving dilute flow, formed by turbulent mixing at the upper boundary of the dense layer (compare Mohrig et al., 1998), which produced a more extensive, thin deposit. Hallworth & Huppert (1998) were unable to explain these phenomena quantitatively using existing theory. The process observed in these
experiments is analogous to that envisaged by Mulder et al. (1997a) for the 1979 event in the Var submarine canyon, off south-east France. The stepped thickness distribution of Hallworth & Huppert’s (1998) experimental deposits is strikingly similar to that reported by Tokuhashi (1979) in a study of thickness distributions of Miopliocene turbidite sandstones in Japan.

**INITIATION AND EVOLUTION OF TURBIDITY CURRENTS**

Normark & Piper (1991) reviewed a variety of initiation mechanisms for turbidity currents, including virtually instantaneous events such as seismic slumping (e.g. Heezen & Ewing, 1952; Morgenstern, 1967; Weaver et al., 1992; García & Hull, 1994) that generate surge-type currents. Such currents have their experimental analogue in the finite-volume currents generated by the release of a lock gate, and may consist of a more-or-less clearly differentiated head, body and tail, with a general decrease in both mean velocity and sediment concentration backwards from the head (e.g. Felix, in press). Typically, the velocity of the current passing a fixed point will very rapidly reach its maximum value as the head arrives, after which flow will wane during passage of the body and tail (e.g. Fig. 1). However, even surge-type currents may pass through periods of relatively steady flow. For example, the 1929 Grand Banks turbidity current, which was triggered by slope failure following an earthquake (Cochonat & Piper, 1995), continued for a minimum of 2–3 h (Piper et al., 1988; see also Piper et al., 1999a), and was sufficiently steady to develop large-scale bedforms in the fan valley (Hughes-Clarke et al., 1990). Sustained (steady or quasi-steady) underflows may also be generated during volcanic eruptions and the consequent remobilization of unconsolidated material (e.g. Lipman & Mullineaux, 1981; Kokelaar, 1992), or by seismically triggered subaerial sliding within the drainage basin (e.g. Syvitski & Schafer, 1996).

Steady or quasi-steady currents may have an element of inherent instability owing to internal waves, the generation and impact of eddies and longer term ‘surgings’ (e.g. Lambert et al., 1976) that may be expressed in the deposit (e.g. Duringer et al., 1991).

Several analytical and numerical models of turbidity currents have produced solutions for steady flow (e.g. Chu et al., 1979), but the application of steady flow models to the interpretation of deposits has not been generally popular, despite evidence in deposits for steady or quasi-steady flow (Kneller, 1995).

**The fluvial connection**

Many turbidite systems, especially fan systems, display obvious connections with rivers. The growth of turbidite systems occurs when fluvial systems can discharge directly to the shelf break. This direct link is generally associated with seal-level lowstands (Mutti, 1985; Shanmugam et al., 1985; Mutti & Normark, 1991; Posamentier et al., 1991; Normark et al., 1993) but may also occur during highstands if rates of delta progradation are sufficiently high (Burgess & Hovius, 1998). The initiation of turbidity currents is generally ascribed to remobilization of sediment deposited at the shelf break or in the canyon head (e.g. Kolla & Perlmutter, 1993; Mulder & Syvitski, 1996). However, turbidity currents (that may be
sustained for periods of days or weeks) can be generated directly by fluvial discharge into lakes (e.g. Lambert et al., 1976; Sturm & Matter, 1978; Weirich, 1986; Lambert & Giovanoli, 1988), and into the ocean, where they are linked to high sediment load in the source fluvial system. Examples occur on fjord delta-fronts (Prior & Bornhold, 1990; Zeng et al., 1991; Carlson et al., 1992; Phillips & Smith, 1992; see also Nemec, 1990) and in more open marine systems, notably off the Yellow River, the Zaire (Congo) River and the Santa Clara river that supplies the Hueneme Fan on the California Borderland (Heezen et al., 1964; Shepard & Emery, 1973; Reynolds, 1987; Wright et al., 1988, 1990; Piper et al., 1999b). The Yellow River generates seasonal silty hyperpycnal flow (a turbidity current formed directly from dense river effluent; Wright et al., 1990), and the Zaire River supplies a turbid underflow directly to the head of the Zaire Canyon that may flow down onto the fan surface (Eisma & Kalf, 1984; van Weering & van Iperen, 1984) as well as possibly serving less frequent sandy turbidity currents.

Rivers that generate hyperpycnal flows at the present day are generally those with high hinterland relief and exceptionally high suspended loads (e.g. the Var, which drains the south-western French Alps; Mulder et al., 1997b), enhanced where the available sediment is fine-grained such as the loess carried by the Yellow River. Mulder & Svytski (1995) calculated the likely interval between hyperpycnal flow events for a selection of the world’s river systems, and conclude that almost half could generate hyperpycnal flows with a return period of 100 years or less. This is probably an underestimate since they implicitly assumed hyperpycnal flow generation only when the depth-averaged suspended sediment concentration led to an effluent density exceeding that of sea water. However, sediment-charged rivers show considerable density stratification, and many flood events might produce both plumes and hyperpycnal flows when the depth-averaged effluent density is less than that of sea water (see, for example, McLeod et al., 1999). Even with this caveat, it is likely that the rivers that supply large fans can never generate hyperpycnal flows under the current (Holocene) hydrological regime. However, most of these fans have been inactive throughout the Holocene. Damuth et al. (1988) suggested that sinuous channels on the Amazon Fan, which developed during late Pleistocene lowstands, may have been formed by continuous underflows (see also Hay, 1987).

The association of turbidity currents with flood events does not necessarily imply hyperpycnal river effluent, even when the turbidity currents are associated with bursts of low salinity (Zeng et al., 1991). Rapid deposition at the river mouth during prolonged floods may lead to oversteepening and repeated failure of delta slopes, generating turbidity currents that include a proportion of the brackish pore water of the delta sediments.

**Ignite flows**

Both sustained and surge-type fine-grained underflows may also serve as triggering mechanisms for larger sandy flows; where such currents flow into canyons they may become auto-suspensate, which is to say that they are self-sustaining (Pantin, 1979; or ignite, *sensu* Parker, 1982), and may entrain sediment that has previously been introduced into the canyon by littoral drift, storms or smaller turbidity currents. Pantin (in press) has recently provided the first experimental evidence of auto-suspension. Canyon-flushing associated with surge-type currents initiated by slope failures may produce currents whose final volume may be several times that of the portion of the slope that has failed (e.g. Grand Banks, Piper & Aksu, 1987; Piper et al., 1988; Cochonat & Piper, 1995; Madeira Abyssal Plain, Masson, 1994). Ignite flows do not require that the triggering current is sandy, as noted by García & Parker (1993) who demonstrated experimentally the entrainment of sediment by saline currents, and concluded that ‘currents of silty mud could also entrain substantial amounts of sand and carry it to deep water’.

**Currents with reversing buoyancy**

The behaviour of currents with buoyant interstitial fluid (such as currents with warm, fresh or brackish interstitial water entering the sea, pyroclastic currents flowing into water, or lofting pyroclastic currents) has been investigated experimentally by Sparks et al. (1993), Carey et al. (1996) and Hürzeler et al. (1996), and numerically by Hürzeler et al. (1995) using a k-ε turbulence model (see below). McLeod et al. (1999) have experimentally investigated the behaviour of such currents on entering the sea, including the effect of flow stratification in splitting the current into a dense underflow and a buoyant plume. The front speed of underflows with buoyant interstitial fluid decreases more rapidly than that of currents in which the interstitial fluid has the
same density as the ambient. These underflows ultimately come to a halt as sedimentation results in a reversal of buoyancy (the bulk density drops below that of the ambient fluid), and the current lifts off, the point of lift-off remaining constant for a constant discharge. The lofted fluid carries fine sediment with it, thus potentially decoupling the coarse and fine sediment fractions. The lofted fluid forms a plume that rises to a level of neutral buoyancy (if in a stratified environment) or to the water surface, and spreads as a gravity interflow or overflow. Sediment falling from the plume produces a widespread fall-out deposit, termed hemiturbidite by Stow & Wetzel (1990) describing what they interpreted as the consequences of flow. Lofting pyroclastic currents may produce debris-flow-like deposits with abrupt fronts (e.g. Lipman & Mullineaux, 1981), and it is tempting to suggest that the deposits of hyperpycnal flows should have similar properties (i.e. somewhat debrite-like) produced by late-stage reconcentration as the current collapses. Indeed, the deposits of the youngest (latest Pleistocene) lobes of the Mississippi fan (Twichell et al., 1995) have geometries intriguingly similar to those of some lofted pyroclastic currents.

Approaches to understanding suspension currents

Prediction of erosion by turbidity currents, and of the distribution of turbidite deposits, such as their extent, thickness and grain size distribution, requires an understanding of the mechanisms of sediment transport and deposition, which are in turn dependent on the fluid dynamics of the currents. Middleton (1993) states that ‘a better understanding of turbidity current dynamics is required to account for the many diverse phenomena reported from modern environments and ancient turbidite systems’. However, the dynamics of suspension currents are highly complex owing to feedback between turbulence and the suspended sediment. Consequently, a greatly simplified approach is generally taken when modelling suspension currents, commonly by depth averaging.

Furthermore, laboratory experiments investigating the dynamics of gravity currents have established that the distribution of turbulence and shear stresses is significantly different from that in open-channel flows (see section on turbulence structure below and Kneller et al., 1997, 1999; Buckee et al., in press). The turbulence structure affects the way in which sediment suspension is transported and hence the way in which turbidity current behaviour and sediment deposition is modelled.

Since Kuenen (1937) and Kuenen & Migliorini (1950) first demonstrated experimentally that turbidity currents were the most probable agent of transport of clastic sediment into the deep oceans, subsequent research has followed three main paths: the study of turbidite systems, laboratory experiments and theoretical modelling. Turbidite systems are beyond the scope of this paper and the reader is referred to reviews by Mutti & Normark (1991), Mutti (1992) and Normark et al. (1993), and articles in Weimer et al. (1994) and Pickering et al. (1995). The extreme complexity of most turbidite systems, and indeed individual turbidite beds, has precluded the development of quantitative models of turbidity current behaviour inferred solely from their deposits.

Study of the mechanics of turbidity currents has been limited by the difficulties implicit in studying them in the natural environment. Turbidity currents in the oceans may reach velocities of tens of metres per second (e.g. Mulder et al., 1997a) and heights of hundreds of metres (e.g. Heezen & Ewing, 1952) and even rather modest currents have damaged or destroyed equipment deployed for the purposes of studying them (e.g. Zeng et al., 1991). Consequently, most of what is known about large natural turbidity currents (i.e. those significant in terms of sediment transfer to the deep sea) is inferred from indirect sources, such as submarine cable breaks and heights of deposits above submarine valley floors (e.g. Piper & Savoye, 1993). Direct measurements of pyroclastic currents, with the exception of frontal velocities, are more hazardous still. A few smaller, natural turbidity currents have been visualized or monitored (Hay et al., 1982; Chikita, 1989, 1990; Zeng et al., 1991), but our knowledge of natural suspension currents remains fragmentary. Small-scale laboratory experiments therefore offer one of the best means of studying suspension current dynamics. Advances in experimental technology have increased our understanding from broad descriptions of gravity current morphology (e.g. Middleton, 1966b; Allen, 1985; Simpson, 1997) to
the structure of turbulence in these currents (Kneller et al., 1997, 1999; Parsons, 1998; Best et al., in press; Buckee et al., in press).

Mathematical models can also provide significant insights into current dynamics. Analytical solutions have been proposed for some aspects of gravity current behaviour (e.g. Chu et al., 1979), but in the long term numerical techniques present the best hope of understanding and predicting three-dimensional turbidity current processes and deposits. Turbidity currents are, however, highly complex phenomena; they are nonuniform, unsteady, nonlinear, free boundary flows driven by a nonconservative density difference (i.e. the density difference varies as sediment is eroded or deposited) and a combination of body and pressure forces (Allen, 1985). In pyroclastic currents one must also consider the effects of a nonconservative interstitial fluid phase (due to exsolution of magmatic volatiles, substrate boiling and combustion), and volume and pressure changes (due to heating of admixed air, and gas phase compressibility, although this can probably be neglected in currents with low Mach numbers; Furbish, 1997). In most cases there are more variables than governing equations – the so-called closure problem – and the models rely upon simplifying assumptions in order to achieve a closure. The accuracy of the individual models thus depends upon the choice and validity of the assumptions made. Experimental results provide a means of constraining some of these variables as well as providing a test for such models.

EXPERIMENTS ON TURBIDITY CURRENTS

Scaling
Mathematical models provide a valuable method for understanding and predicting the consequences of the complex relationships and feedback mechanisms involved in turbidity currents. However, physical data from field observations, or more practically from experiments, are still required in order to test the simplifying assumptions necessary in such models. The advantages of using experimental models are offset by the difficulties arising from the necessary simplifications involved in scaling. Scaling of natural currents to a laboratory level is a well-documented issue in fluid dynamics (see, e.g. Middleton & Southard, 1984; French, 1986). In a recent review paper, Peakall et al. (1996) outline four approaches to the scaling of physical models: 1:1 replicas of the field prototype; Froude number similarity; distorted scale modelling; and unscaled experimental analogues. The large scale of turbidity currents in the oceans that produce significant sedimentary deposits (e.g. current thicknesses of $10^5$–$10^6$ m and velocities of $10^4$–$10^5$ m s$^{-1}$; see, e.g. Heezen & Ewing, 1952) makes it impossible to study these experimentally at a 1:1 scale; consequently, scaled laboratory experiments present the best way to study turbidity currents. The results of unscaled experimental analogue models of large turbidity currents (e.g. Alexander & Morris, 1994), although qualitatively informative, are difficult to apply quantitatively. Distorted scale experiments, for example where unrealistically high slopes are used to obtain appropriate bed shear stresses (e.g. Postma et al., 1988) may be the only way of reproducing some aspects of the prototype, but clearly necessitate some circumspection in application of the results.

Froude scale modelling
Froude scale modelling is based on a similarity approach in which the current is fully characterized by a series of dimensionless variables. As long as the values of the dimensionless variables in laboratory currents are known to be comparable with those of the natural current, the experiment is adequately scaled with respect to the parameters included in that variable, allowing modelling of large-scale phenomena in the laboratory (e.g. Middleton, 1966a). Dimensional analysis depends upon identification of the controlling variables, which are then grouped into a smaller number of dimensionless parameters such as the flow Reynolds number (Re), which reflects the ratio of inertial to viscous forces, and the Froude number (Fr), which reflects the ratio of inertial to gravitational forces acting on a fluid flow. Currents that share the same values of Re and Fr are said to be dynamically similar. In the case of Froude scale modelling, Fr takes the same value as the prototype, but Re is relaxed (making the assumption that the effects of viscosity can be neglected if the current is fully turbulent). Froude scale models are only appropriate for turbidity current modelling where the model is fully turbulent (Re > 2000) which is sometimes difficult to achieve in the laboratory, and even then assumes that turbulence is self-similar across the range of scales represented by the model and prototype. In gravity currents, the densimetric Froude number
(Fr') is used, since the action of gravity depends upon the fractional density difference between the current and the ambient fluid:

\[ Fr' = \frac{\bar{U}}{d(g')^{1/2}} \]  

(3)

where \( \bar{U} \) is the depth-averaged current velocity, \( d \) is some characteristic length scale (normally the current thickness in gravity currents), and the reduced gravity \( g' = g(\Delta \rho / \rho_a) \), where \( g \) is the acceleration due to gravity, \( \Delta \rho \) is the density difference between the current and the ambient fluid and \( \rho_a \) is the density of the ambient fluid.

**Scaling sediment-laden currents**

Experimental studies of sediment-laden turbidity currents must also be scaled to the natural system using a dimensionless settling velocity (making the assumption that settling velocity adequately describes the particle hydrodynamics), that is the ratio of terminal settling velocity of the sediment grains to some velocity scale that is considered characteristic of the current (Middleton, 1966a; Laval et al., 1988). Electrostatic forces may affect very fine sediments (such as chalk, charcoal or clay) which have the correctly scaled settling velocity, but this problem can be solved by using glass beads or silica flour (e.g. Parker et al., 1987; García, 1993). However, all fine sediments experience capillary forces once they have settled, and these can only be reduced by using larger grain sizes. Better sediment analogues can be obtained by using grains of reduced density (e.g. Middleton, 1966b,c), but the consequent reduction in the bulk density of the current necessitates an increase in sediment concentration. Changing sediment concentration or grain size affects particle behaviour, which may no longer scale with the whole current behaviour (Middleton, 1966a; Peakall et al., 1996). Many experimental studies substantially simplify the system by neglecting sediment deposition and erosion, and use brine (or other dense solutions) to create the density difference (Laval et al., 1988). However, while saline currents are believed to be dynamically similar to fine-grained turbidity currents (Stacey & Bowen, 1988a), the vertical density structure (and hence current dynamics) may be considerably different in coarse-grained turbidity currents.

**Review of experimental work**

Table 1 shows the majority of experimental work carried out on the dynamics of gravity currents (thus excluding a substantial set of experiments primarily concerned with sediment deposition from lock-exchange currents; see, e.g. Alexander & Morris, 1994; Kneller & McCaffrey, 1995; Gladstone et al., 1998). Table 1 highlights the way in which technological advances have led to increasingly detailed studies of gravity current dynamics. Initial experimentation focused on the anatomy of gravity currents, especially the head (e.g. Keulegan, 1957; Middleton, 1966b; Britter & Simpson, 1978). More recently, the development of facilities in which steady currents can be generated has allowed the body of experimental currents to be examined (e.g. Parker et al., 1987; García, 1993, 1994). However, flow measurements were made primarily using intrusive equipment (e.g. micropropellers), which disrupt the flow, and so only mean flow properties rather than the turbulence structure were determined. In the last three years (see Table 1), the application of several new experimental techniques, e.g. laser Doppler anemometry (LDA) in combination with refractive index (RI) matching, acoustic Doppler velocimetry (ADV) and ultrasonic Doppler velocity profiling (UDVP), has allowed the preliminary quantification of the turbulence structure both in the head and in the body of laboratory gravity currents (e.g. Kneller et al., 1997, 1999; Parsons, 1998; Best et al., in press; Buckee et al., in press). There are still technical limitations to this work, specifically the difficulty in measuring instantaneous density fluctuations, and the lack of high-resolution velocity data in particulate currents (LDA has not successfully been used in sediment-laden density currents, and UDVP measures at frequencies < 30 Hz, too low for a full determination of the turbulence spectrum; Parsons, 1998).

**THE ANATOMY OF GRAVITY CURRENTS**

**The head**

Gravity currents are described as having a well-defined head, body and in some cases a tail. The dynamics of the head are important because they may set a boundary condition for the current as a whole (Simpson & Britter, 1979). Both Allen (1971) and Middleton (1993) suggest that the head is a ‘locus of erosion’, and therefore important sedimentologically. The head has an overhanging ‘nose’ (Fig. 2), a result of the no-slip condition at the lower boundary (Briter & Simpson, 1978) and frictional resistance at the
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<td>Middleton (1966b)</td>
<td>Solute &amp; sediment-laden</td>
<td>Head, body</td>
<td>Continuous</td>
<td>Slow-motion photography</td>
<td>n/a</td>
<td>Simple laws</td>
</tr>
<tr>
<td>Britter &amp; Simpson (1978)</td>
<td>Solute (Saline)</td>
<td>Head</td>
<td>Arrested head</td>
<td>Conductivity meter (no results shown)</td>
<td></td>
<td>Mean flow properties, mixing rate, Kelvin-Helmholtz billows</td>
</tr>
<tr>
<td>Simpson &amp; Britter (1979)</td>
<td>Solute (Saline)</td>
<td>Head</td>
<td>Arrested head</td>
<td>Conductivity meter (no results shown)</td>
<td></td>
<td>Mean flow properties, sediment entrainment</td>
</tr>
<tr>
<td>Parker et al. (1987)</td>
<td>Sediment-laden</td>
<td>Body</td>
<td>Continuous</td>
<td>Micropropellor</td>
<td>Siphon sampling</td>
<td>Mean flow properties, sediment entrainment &amp; deposition, entrainment of ambient</td>
</tr>
<tr>
<td>García (1993)</td>
<td>Solute &amp; sediment-laden</td>
<td>Body</td>
<td>Continuous</td>
<td>Micropropellor</td>
<td>Optical probe</td>
<td>Mean flow properties, hydraulic jump, deposits</td>
</tr>
<tr>
<td>García &amp; Parker (1993)</td>
<td>Solute &amp; sediment-laden</td>
<td>Body</td>
<td>Continuous</td>
<td>Micropropellor</td>
<td>Optical probe</td>
<td>Mean flow properties, sediment entrainment</td>
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<tr>
<td>García (1994)</td>
<td>Sediment-laden</td>
<td>Body</td>
<td>Continuous</td>
<td>Micropropellor</td>
<td>Siphon sampling</td>
<td>Mean flow properties, sediment deposition</td>
</tr>
<tr>
<td>Altinakar et al. (1996)</td>
<td>Sediment-laden</td>
<td>Body</td>
<td>Modified lock-exchange</td>
<td>Indirectly from arrested head facility</td>
<td></td>
<td>Mean flow properties</td>
</tr>
<tr>
<td>García &amp; Parsons (1996)</td>
<td>Solute (Saline)</td>
<td>Head</td>
<td>Arrested head</td>
<td>Micropropellor</td>
<td>Siphon sampling</td>
<td>Mixing rate</td>
</tr>
<tr>
<td>Hallworth et al. (1996)</td>
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<td>Head</td>
<td>Lock-exchange</td>
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<td>n/a</td>
<td>Entrainment into the head</td>
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<tr>
<td>Kneller et al. (1997)</td>
<td>Solute (RI matched)</td>
<td>Head, body</td>
<td>Lock-exchange</td>
<td>LDA, video</td>
<td>n/a</td>
<td>Mean flow properties, turbulence, reflections</td>
</tr>
<tr>
<td>Parsons (1998)</td>
<td>Solute (Saline)</td>
<td>Head</td>
<td>Arrested head</td>
<td>ADV, laser-induced fluorescence (LIF)</td>
<td>Conductivity probe</td>
<td>Turbulence, entrainment of ambient, mixing rate</td>
</tr>
<tr>
<td>Kneller et al. (1999)</td>
<td>Solute (RI matched)</td>
<td>Head, body</td>
<td>Lock-exchange</td>
<td>LDA, video</td>
<td>n/a</td>
<td>Mean flow properties, turbulence</td>
</tr>
<tr>
<td>Best et al. (in press)</td>
<td>Sediment-laden</td>
<td>Head, body</td>
<td>Modified lock-exchange</td>
<td>UDVP, video</td>
<td>n/a</td>
<td>Mean flow properties, turbulence, reflections</td>
</tr>
<tr>
<td>Buckee et al. (in press)</td>
<td>Solute (RI matched)</td>
<td>Body</td>
<td>Continuous</td>
<td>LDA</td>
<td>Conductivity probe</td>
<td>Mean flow properties, turbulence, stratification</td>
</tr>
</tbody>
</table>
upper boundary. At the rear of the head a series of transverse vortices (Fig. 2) are present, identified as Kelvin–Helmholtz instabilities (Bitter & Simpson, 1978; see below).

Middleton (1966b) performed the first comprehensive set of experiments on the heads of brine currents and turbidity currents in the laboratory. He investigated the effects of slope on the velocity and shape of the head, finding that for low slopes (angles <2-3°) the head velocity is adequately described by Keulegan’s (1957) formula in which the head velocity is independent of the slope. Further experiments suggest that dimensionless head velocity is only weakly dependent on slope (for slopes of 5–90°) because, with increasing slope, higher gravitational forces are counteracted by increased frictional resistance at the upper boundary owing to increased rates of ambient entrainment (Bitter & Linden, 1980). The downstream velocity in the body of the current, which does depend on the slope, has been shown to be up to 30–40% faster than the head velocity (Middleton, 1966b,c; Kneller et al., 1997, 1999; Best et al., in press). Consequently, the head height increases with slope, as the body velocity increases and material moves more rapidly into the head (Hopfinger & Tochin-Danguy, 1977; Bitter & Linden, 1980; Simpson, 1997).

Mixing of the current with the ambient fluid is an important process that occurs at the head, primarily by the detraining of dense fluid out of the back of the head in a series of transverse vortices (e.g. Allen, 1971; Bitter & Simpson, 1978; Simpson & Bitter, 1979). Bitter & Simpson (1978) used slow-motion pictures to trace the shapes of the billows at the upper boundary, showing that the ratio of billow amplitude to wavelength was the same as for Kelvin–Helmholtz billows. These billows can grow in wavelength and amplitude to about the size of the head, and collapse as dense fluid is detrained into a ‘mixing region’ of approximately the same thickness as the billow amplitude (Bitter & Simpson, 1978; Simpson & Bitter, 1979). Middleton (1966b) investigated the motion of fluid and particles within, and in front of, the head, showing that flow lines diverge within the head and that the head displaces the ambient fluid. Further experiments on the displacement of ambient fluid by the head of the current led to the description of the ‘lobe and cleft’ structure at the base of the head, attributed to a gravitational instability formed when less dense fluid is over-run by, and incorporated into, the head (e.g. Simpson, 1969, 1972; Allen, 1971). Simpson (1969) visualized the lobe and cleft structure (Fig. 2), demonstrating that the clefts extend a substantial distance back from the head. Recent work, however, shows that lobe spacing scales with the flow thickness and that the development of the lobe and cleft structures is probably attributable to a secondary instability associated with Kelvin–Helmholtz vortex breakdown (Schowalter et al., 1994; Parsons, 1998). Simpson & Bitter (1979) used dye tracers to visualize the ambient fluid being
entrained into the head at its base. They calculated a flux of light ambient fluid moving under the head and found it accounted for only 1% of the total mixing, the rest occurring at the upper boundary.

The body

Few physical descriptions of the body of gravity currents exist in the literature. Ellison & Turner (1959) describe the body as a region of steady downstream velocity which has a thin, dense layer of fluid near the base of the current which, with increasing downstream velocity, mixes with the ambient fluid at the upper boundary as an 'irregular succession of large eddies'. Similarly, Britter & Simpson (1978) and Simpson & Britter (1979) used visual observations and measurements to divide gravity currents, behind the head, into two distinct regions; the body, which is the lower dense layer, and a region of less dense, mixed fluid that has been mixed out of the head of the current. A similar structure was inferred by Kneller et al. (1999) based on the effect of mixing on the refractive index of brine currents. Some controversy surrounds the nature of the mixed region above the dense layer; it has been suggested that this is not strictly part of the gravity current but should be described as a zone of 'clouded water' entrained by the underflow (Middleton, 1966b; based in part on Kuenen, 1951; Blanchet & Villatte, 1954).

MEAN FLOW PROPERTIES OF GRAVITY CURRENTS

Downstream velocity profile

Figure 3 shows a typical vertical profile of the downstream velocity and serves to define some of the terms used. The shape of the downstream velocity profile in experimental and natural gravity currents (Chikita, 1990; Altinakar et al., 1996; Simpson, 1997) is similar to the velocity profile in turbulent plane wall jets (Lauder & Rodi, 1983). Gravity currents (and wall jets) are described as having an inner and outer region divided by the velocity maximum. The inner, wall-bounded region has a positive velocity gradient and is generally less than half the thickness of the outer region, which has a negative velocity gradient. The height of the velocity maximum is controlled by the ratio of the drag forces at the upper and lower boundaries (Middleton, 1966c; Kneller et al., 1997), and in many experimental currents is found to occur at about 0.2–0.3 of the height of the current (Altinakar et al., 1996; Kneller et al., 1997, 1999; Best et al., in press). For gravity currents traveling over a rough bed (e.g. turbidity currents over bedforms or gravel beds) the height of the velocity maximum will be raised compared with those that are dominated by drag at the upper boundary (as in supercritical currents, e.g. García, 1993; Buckee et al., in press) which tends to lower the level of the velocity maximum. Despite this variability, it is possible to collapse velocity profiles from many different experimental currents (Fig. 4), by using a characteristic length scale, \( y_{1/2} \) (defined as the distance between the bed and the height in the outer region at which the downstream velocity is half the maximum downstream velocity, see Fig. 3; Buckee et al., in press).

Concentration profiles

Gravity currents are density stratified (i.e. there is a vertical gradient in concentration), having a dense basal layer of fluid and sediment (or dense fluid), with a less dense, more homogeneous mixed region above. A number of models for the distribution of density in gravity currents have been proposed (see discussion in Peakall et al., in press; Fig. 5). Initial two-layer models (e.g. Middleton, 1966b, 1993; Simpson

Structure of turbidity currents  

& Britter, 1979) were based on visual observations of saline currents (e.g. Middleton, 1966b; Britter & Simpson, 1978; Simpson & Britter, 1979) and measurements in clay suspension currents (e.g. Blanchet & Villatte, 1954) and saline currents (e.g. Britter & Simpson, 1978; Simpson & Britter, 1979). The inflection point in the density distribution was observed to occur well above the level of the velocity maximum, and divided the ‘body’ of the current from the ‘mixed’ fluid detrained from the head (Fig. 5a). However, concentration measurements from both experimental sediment-laden currents (e.g. García, 1993, 1994; Altinakar et al., 1996) and saline currents (e.g. Parsons, 1998; Buckee et al., in press) demonstrate that the density structure is somewhat more complex. Two main types of sediment concentration profile have been observed (Peakall et al., in press). A smooth profile (Fig. 5b) is commonly seen in low-concentration, weakly depositional currents (Altinakar et al., 1996; García, 1990, 1994) and in saline gravity currents (Ellison & Turner, 1959; Buckee et al., in press). These currents are nonetheless highly stratified, with a density gradient that is greatest near the base of the current and decreases rapidly around the level of the velocity maximum. The second class of density distribution has a ‘stepped-concentration’ profile (Fig. 5c) and is commonly observed

**Fig. 4.** Collapse of downstream velocity data using characteristic height $y_{1/2}$ (defined as the height in the outer region at which the downstream velocity is half the maximum velocity, see Fig. 3). The data represent both supercritical and subcritical currents.

**Fig. 5.** Schematic diagram showing various characteristic density/concentration profiles (dashed lines) in density currents; the same downstream velocity profile (solid line) has been used in each case for reference. In each case the $y$-axis represents a normalized height and the $x$-axis normalized concentration and velocity. (a) A two-layer model type concentration profile, dividing the flow into a constant density lower region (the current) and an upper region of fluid detrained from the head (e.g. Britter & Simpson, 1978; Simpson & Britter, 1979; Middleton, 1993). (b) A smooth profile, characteristic of low-concentration, weakly depositional flows (e.g. Altinakar et al., 1996; García, 1990, 1994). (c) A stepped concentration profile observed in erosional flows (e.g. García, 1993). (d) A Rouse-type distribution of sediment grain-sizes observed in turbidity currents (e.g. García, 1994), in which coarse material is concentrated towards the lower part of the flow whereas fine-grained material is more evenly distributed throughout the depth of the flow (modified from Peakall et al., in press).

in erosional currents (García, 1993) or currents interpreted to have a high entrainment rate at the upper boundary (Peakall et al., in press). Experiments in which vertical grain size distributions have been measured (García, 1994) show that fine-grained material is more uniformly distributed in the vertical than the coarse material, which tends to become concentrated in the lower part of the current. Theoretical and experimental studies of turbidity currents (Stacey & Bowen, 1988a,b; García & Parker, 1993; García, 1994; Altinakar et al., 1996), and limited field data (e.g. Chikita, 1989; Normark, 1989; Zeng & Lowe, 1997), suggest that they have patterns of vertical sediment distribution rather similar to those of shear flows, in which the vertical sediment concentration profile obeys a power law distribution (given by Rouse, 1937; see Middleton & Southard, 1984) with an exponent that is directly proportional to the ratio of the current shear velocity, $U_c$, to the grain settling velocity, $U_s$ (Middleton & Southard, 1984). Relatively low values of $U_c/U_s$ predict high near-bed concentrations of suspended sediment, decaying rapidly upwards (Fig. 5d). This has two important consequences; firstly, any polydisperse (i.e. multiple grain-size) current must be stratified in terms of both density and grain-size, particularly if it is depositional or close to being so; secondly, the grain-size range near the base of the current is broader than that higher in the current.

**TURBULENCE STRUCTURE OF GRAVITY CURRENTS**

The head: temporal variations, vertical structure and entrainment

Data on the turbulence structure in the head of a gravity current have been presented by Kneller et al. (1997, 1999) and Parsons (1998). Time series of instantaneous downstream velocity clearly record the arrival of the head and the passage of large, low-frequency Kelvin–Helmholtz billows superimposed on a period of quasi-steady motion (Kneller et al., 1997, 1999; Fig. 1). Instantaneous velocities, associated with large eddies, were found to exceed the maximum mean downstream velocity by up to 50%. Similarly, large instantaneous peaks in Reynolds stress were recorded. Two areas of high negative Reynolds stress were identified – at the top of the head, related to the Kelvin–Helmholtz billows, and beneath the nose of the current, attributed to the entrainment of over-ridden ambient fluid which is buoyant within the current (Kneller et al., 1999). The turbulence structure in gravity current heads, in which Fr is always less than one, is dominated by shearing at the upper interface. This is reflected in the distributions of turbulent kinetic energy and density fluctuations, both of which have a maximum at the upper boundary of the head (Kneller et al., 1997, 1999; Parsons, 1998; Best et al., in press).

García & Parsons (1996) and Parsons (1998), demonstrated experimentally that mixing in gravity current heads is Reynolds number dependent; dimensionless mixing rates decrease with decreasing Reynolds number. In large (high-Re) currents, which are fully turbulent, entrainment is primarily due to secondary instabilities (i.e. Kelvin–Helmholtz vortex breakdown) in the upper part of the current. In small currents (i.e. low-Re currents in which all the turbulent subranges are not present, for instance the laboratory experiments of Britter & Simpson, 1978), entrainment due to secondary instabilities (i.e. Kelvin–Helmholtz vortex breakdown) is less significant as viscous effects become more important. Care must be taken in laboratory currents therefore to ensure that the Reynolds number is sufficiently high (> 2000) to minimize such viscous effects as they may compromise the scaling approach taken. Hallworth et al. (1996) examined entrainment into the head of lock-exchange currents, using a pH neutralization technique whereby the alkali current entrained acidic ambient fluid. The amount of dilution in the head could be quantified using pH indicator solution. Hallworth et al. (1996) suggest that in lock-exchange currents, total entrainment at a given distance from the source is independent of the reduced gravity and related primarily to the initial volume. Experiments on gravity currents travelling over a free surface (i.e. the water–air interface) led Hallworth et al. (1996) to the conclusion that approximately two-thirds of the total entrainment into the head, for currents over a rigid surface, occurs by overriding of ambient fluid. These conclusions are in contrast to most other results which show that entrainment is a function of the densiometric Froude number (e.g. Ellison & Turner, 1959) and hence the initial reduced gravity. Furthermore, many experimental studies have shown that entrainment at the lower boundary
by overriding of ambient fluid is relatively insignificant (e.g. Simpson & Britter, 1979; Parsons & García, 1995; Parsons, 1998; see above). The contradictory results presented in Hallworth et al. (1996) probably reflect the significant differences in the turbulence structure of gravity currents flowing over fixed (i.e. the bed) and free surfaces (i.e. the water–air interface), and the increased influence of viscosity in low-Re (<2000) currents.

**The body: temporal variations**

Time series of downstream velocity in the body of quasi-steady gravity currents reveal the presence of large coherent structures that advect with the current (Kneller et al., 1997; Best et al., in press). Instantaneous downstream velocities may be up to 40% higher than the maximum mean downstream velocity in the body (Buckee et al., in press), and therefore equivalent to, or higher than, instantaneous velocities in the head. This result suggests that the body of the current may play a significant role in sediment entrainment, as such high turbulent velocities imply high Reynolds stresses (Kneller et al., 1997) and hence high potential erosion rates. Spectral analysis of velocity time series (Kneller et al., 1999) shows that low-frequency (<10Hz) motions dominate the energy budget at all depths in the current.

**The body: vertical structure**

Laboratory experiments show that turbulence intensities are highest at the top of the gravity current (e.g. Kneller et al., 1997, 1999; Best et al., in press), and related to large-scale shearing and mixing at the upper boundary. Low turbulence intensities are observed around the level of the velocity maximum, or the high density gradient near the base of the current (Best et al., in press; Buckee et al., in press). Similarly, turbulent kinetic energy is low at around the level of the velocity maximum (Kneller et al., 1999; Buckee et al., in press). In currents with high levels of shear at the upper boundary (e.g. lock-exchange currents, supercritical currents) maximum turbulent kinetic energy occurs towards the top of the current (Kneller et al., 1999; Buckee et al., in press; Fig.6). However, in subcritical currents without a counterflow of ambient fluid (owing to replacement of fluid in the lock), maximum turbulent kinetic energy occurs towards the base of the current and Reynolds stresses at that point are high and positive, reflecting an upward transfer of momentum (Buckee et al., in press), with the level of zero Reynolds stress being...
displaced above the velocity maximum. Negative Reynolds stresses in the upper part of the current are associated with mixing of the ambient fluid downward into the current. Significant instantaneous variations may occur in the vertical Reynolds stress profile however, reflecting flow unsteadiness and the passage of large eddies (Kneller et al., 1997).

Energy budget in gravity currents

Buckee et al. (in press) calculated rates of turbulent kinetic energy production in steady experimental gravity currents from shearing of the mean flow and from buoyancy flux. The results showed that the highest rates of turbulence production occurred due to shearing near the lower boundary, where the velocity gradient is highest. The results also suggested that turbulence production by shear is more significant in currents that are more highly density stratified (e.g. currents in which vertical mixing is inhibited; subcritical currents) while turbulence production by buoyancy flux is more important in currents that are less stratified (e.g. currents in which there is more significant vertical mixing; supercritical currents). The study by Buckee et al. (in press) was limited, however, by poor vertical resolution for the calculation of velocity gradients and the fact that the entire energy budget could not be resolved owing to lack of spatial resolution in the downstream direction.

DENSITY STRATIFICATION IN GRAVITY CURRENTS

Definitions

A stable density stratification is one in which the overall density decreases upwards (Turner, 1973). However, a current that is stratified on a large scale may, on closer examination, reveal poorly stratified regions as well as well-stratified regions. The gradient Richardson number (\(Ri_g\)) is calculated with respect to local density gradients and highlights regions that are less stably stratified than the whole:

\[
Ri_g = \frac{-gd\rho_y/dy}{\rho_0(dU_f/dy)^2}
\]  

where \(\rho_0\) is some reference density (normally taken as that of the ambient fluid) and \(U_f\) is the mean downstream velocity at height \(y\) in the flow. A widely accepted criterion for the stability of stratification is that if \(Ri_g > 0.25\) locally, then the stratification is sufficiently strong to significantly inhibit vertical mixing at that level (Turner, 1973). However, the presence of a stable stratification does not necessarily imply the absence of turbulence. A stable stratification tends to inhibit vertical mixing, and an external energy supply is therefore required to generate turbulence and prevent its rapid dissipation (e.g. Tritton, 1988).

Stability of stratification in gravity currents

García (1993) and Buckee et al. (in press) calculated \(Ri_g\) above the velocity maximum in both subcritical and supercritical experimental gravity currents. Typically, supercritical currents were found to have \(Ri_g\) in the range 0.13–0.23, below the critical value of 0.25, whereas for subcritical currents \(Ri_g\) had a range between 0.54 and 0.84. The difference in stability between the two flow regimes reflects the greater degree of mixing that occurs at the upper boundary of supercritical currents (Ellison & Turner, 1959). Furthermore, Buckee et al. (in press) calculated vertical profiles of \(Ri_g\) for two experimental currents (Fig. 7), showing that the strength of the stratification varies with height in gravity currents. Stacey & Bowen (1988a,b) calculate theoretical \(Ri_g\) profiles for supercritical currents which are in reasonable agreement with the data of Buckee et al. (in press) for the region above the velocity maximum. At the level of the velocity maximum, where the velocity gradient is zero, calculated \(Ri_g\) goes to infinity. In currents with a smooth sediment concentration profile, the stratification below the velocity maximum is generally stable. However, when the concentration profile is stepped (e.g. Fig. 5c) there may be a region of less stable stratification close to the bed.

THEORETICAL MODELS OF GRAVITY CURRENT BEHAVIOUR

Mathematical modelling of gravity currents takes a range of forms from simple hydraulic equations and box models to highly complex turbulence models (see Table 2 and below). Similarly, mathematical models may be used for a wide range of purposes, from providing estimates of current velocity and thickness to predicting turbidite geometries and grain size distributions,
to modelling the vertical structure of turbulence in a gravity current. In order to model gravity current dynamics, a series of equations must be solved: the conservation of fluid mass, the conservation of sediment mass and the conservation of momentum (known as the Navier–Stokes equation). In addition, an equation for conservation of turbulent energy may be employed. The Navier–Stokes equation is a nonlinear partial differential equation which can be solved for turbulent currents, such as gravity currents, by the application of simplifying assumptions (e.g. Tennekes & Lumley, 1972; Tritton, 1988).

Simple models

The first models of the hydraulic behaviour of turbidity currents were one-equation models based on a modified form of the Chézy equation for steady, uniform flow, which can be derived from the Navier–Stokes equations (Komar, 1977; Stacey & Bowen, 1988a):

\[ \mathcal{U} = \sqrt{\frac{8g'Sd}{f_b + f_1}} \]  

(5)

where \( \mathcal{U} \) is the depth-averaged mean current velocity, \( S \) is the slope (\( \sin \alpha \)), \( d \) is the thickness of the current, \( g' \) is the reduced gravity and \( f_b \) and \( f_1 \) are the Darcy–Weisbach friction coefficients at the lower and upper boundaries of the current, respectively (Daly, 1936; Kuenen, 1952; Hinz, 1960; Middleton, 1966b,c, 1993; Middleton & Southard, 1984). It has been suggested that for large turbidity currents \( f_b + f_1 \) is approximately 0.01 (Middleton & Southard, 1984), where \( f_1 > f_b \), and that both terms are likely to be functions of the densiometric Froude number (Middleton, 1966a; Chikita, 1989). Modified Chézy equations have frequently been used to model the bulk flow properties of natural turbidity currents (e.g. Kuenen, 1952; Bowen et al., 1984; Kirwan et al., 1986; Mulder et al., 1998). Some of these models have been extended in order to study turbidity current evolution (e.g. on the Navy submarine fan, Bowen et al., 1984) and erosion and deposition (e.g. in the Saguenay fjord, Mulder et al., 1998).

Box models

The primary utility of box models is to provide a simple and rapid way to model the rate of propagation of finite-volume, low-concentration, surge-type currents and their broad deposit characteristics. Box models are most effective for modelling currents with relatively low concentrations of small heavy particles which are transported by, and sedimented from, turbulent currents (turbidity currents) (for a review see

Fig. 7. Calculated gradient Richardson numbers for experimental gravity currents (Fr > 1, solid line and Fr < 1, dashed line) taken from Buckee et al. (in press). (a) The horizontal lines represent the level of the velocity maximum for each current. In (b) a detail of (a) is shown co-plotted with a theoretical curve for Ri_g (dotted line) for a current in which Fr > 1 (adapted from Stacey & Bowen, 1988b). It should be noted that the curves derived from experimental data are less smooth than the theoretical curve owing to their poorer vertical resolution. The line Ri_g = 0.25 highlights the value above which the flow is locally stably stratified.

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Table 2. Summary of mathematical models of gravity currents.

<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Model</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kuenen (1952)</td>
<td>Chézy equation</td>
<td>Calculated properties of the 1929 Grand Banks current</td>
</tr>
<tr>
<td>Ellison &amp; Turner (1959)</td>
<td>Depth-averaged</td>
<td>Used in conjunction with laboratory experiments to examine entrainment of ambient fluid at the upper boundary of saline currents</td>
</tr>
<tr>
<td>Kirwan et al. (1960)</td>
<td>General analytical solution</td>
<td>Calculated evolution of the 1929 Grand Banks current and the Orleansville current in order to model the origin of the currents</td>
</tr>
<tr>
<td>Hinze (1960)</td>
<td>Analytical solution in which eddy viscosity is assumed to be constant in the vertical direction</td>
<td>Models density and velocity profiles above the velocity maximum. Predicts velocity and flow thickness for the Grand Banks current in reasonable agreement with other researchers</td>
</tr>
<tr>
<td>Chu et al. (1979)</td>
<td>Depth-averaged</td>
<td>Examines the effect of changing certain variables (sediment exchange with the bed, slope, ambient entrainment and lower boundary drag) on flow evolution</td>
</tr>
<tr>
<td>Bowen et al. (1984)</td>
<td>Modified Chézy equation, including terms for ambient entrainment and Coriolis force</td>
<td>Modeled flow evolution of turbidity currents on the Navy submarine fan</td>
</tr>
<tr>
<td>Parker et al. (1986)</td>
<td>Depth-averaged (3- and 4-equation models)</td>
<td>Examines the theory of self-acceleration (or autosuspension). The 4-equation model appears to predict self-acceleration realistically</td>
</tr>
<tr>
<td>Stacey &amp; Bowen (1988a,b)</td>
<td>Algebraic turbulence closure, eddy viscosity and eddy diffusivity vary vertically with the gradient Richardson number (with a linear bridge across the velocity maximum)</td>
<td>Models vertical profiles of velocity, density and gradient Richardson number Results compare well with laboratory experiments The second paper uses the model to examine self-acceleration and flows bearing multiple grain sizes</td>
</tr>
<tr>
<td>Eidvik &amp; Brars (1980)</td>
<td>k-ε turbulence closure model</td>
<td>Predicts vertical distribution of velocity, density, turbulent kinetic energy and eddy viscosity Results compare well with experimental data Predicts self-acceleration</td>
</tr>
<tr>
<td>Brars &amp; Eidvik (1992)</td>
<td>Reynolds stress turbulence closure model</td>
<td>Predicts vertical distribution of velocity, density, turbulent kinetic energy and eddy viscosity Results compare well with experimental data</td>
</tr>
<tr>
<td>Dade &amp; Huppert (1994)</td>
<td>Box model</td>
<td>Scaling arguments used to apply the model to the deposits of 2D (channelized) turbidites Predicts run-out length and deposit thickness Results compare well with experimental data</td>
</tr>
<tr>
<td>Zeng &amp; Lowe (1997)</td>
<td>Depth-averaged, and accounts for vertical distribution of sediment</td>
<td>Models flow evolution and deposit characteristics</td>
</tr>
<tr>
<td>Kubo et al. (1998)</td>
<td>Box model</td>
<td>Compares model results with palaeocurrent velocities calculated from grain size data from an extensive turbidite in the Miocene–Pliocene Kyusumi Formation, Japan</td>
</tr>
<tr>
<td>Mulder et al. (1998)</td>
<td>Modified Chézy equation, including terms for ambient entrainment and sediment exchange with the bed</td>
<td>Modelled flow evolution of a 28-day flood in the Saguenay fjord in 1663 Compared model deposit characteristics (run-out distance, thickness, grain size) with real data</td>
</tr>
<tr>
<td>Felix (in press)</td>
<td>Mellor-Yamada 2½ turbulence closure model, includes influence of particles on turbulence</td>
<td>Models vertical and downstream distribution of velocity, density, turbulent kinetic energy and eddy viscosity Results compare well with experimental data Models turbidite deposits (thickness, grain size distributions)</td>
</tr>
</tbody>
</table>
Huppert, 1998). Box models are not based directly on the Navier–Stokes equations; instead the current is modelled as an evolving series of rectangles with no horizontal or vertical variation in current properties (Huppert, 1998). The rate of propagation is controlled by the Froude number of the head, which is assumed to be constant in the range 0.7–0.8 (e.g. Middleton, 1993; Huppert, 1998). In this way a dimensional form of the solution can be found which can be solved analytically. Box models simulate currents which are very similar to lock-exchange experiments and achieve good results (except near the point of initiation) when the model results are compared with the deposits of strongly depositional lock-exchange currents (Bonnecaze et al., 1993; Huppert, 1998). These models are adequate for modelling such simple cases, but have rarely been tested against natural deposits owing to the difficulty in constraining initial conditions. Dade & Huppert (1994, 1995) have made use of box models to simulate a known deposit (Black shell turbidite) and calculate initial current volumes and sediment concentrations. Kubo et al. (1998) tested Dade & Huppert’s (1995) box models and Bonnecaze et al.’s (1995) axisymmetric model against well-constrained data from an extensive turbidite bed of the Miocene–Pliocene Kiyosu, Japan. Using grain size data to calculate palaeocurrent velocities, Kubo et al. (1998) showed that the box models significantly underestimated current velocity. This may be partly because box models fail to account for the difference between the velocity of the head and that of fluid within the body.

An inherent problem with box models that also applies to other depth-averaged models (see below) is the assumption that the vertical sediment concentration profile is uniform (as a result of vertical mixing by turbulence); in fact any polydisperse current must be density stratified (Fig. 5d) and this assumption fails. Similarly, box models do not explicitly account for turbulence, or for entrainment of ambient fluid at the upper boundary, both of which are fundamental. Furthermore, Piper et al. (1999a) show that even a current generated by slope failure (e.g. the 1929 Grand Banks current), which might be regarded as a surge-type current, may produce ‘sustained flow over many hours’ which cannot be simulated by a box model. Perhaps neither the deposits produced in lock-exchange experiments nor the results of box models realistically reflect the patterns of deposition beneath more complex natural turbidity currents.

More complex depth-averaged models

In turbidity current modelling, the Navier–Stokes equations are commonly solved in a depth-averaged form. The assumption is made that there are no significant vertical variations in the properties of the current and therefore the governing equations are integrated over the height of the current. This assumption results in one mean value for each variable (such as velocity and sediment concentration) at each point downstream. As it is clear that the flow properties of turbidity currents vary dramatically with height, some researchers impose vertical profiles (or shape-factors) for velocity and sediment concentration (Turner, 1973; Pantin, 1979; Parker et al., 1987; García, 1994). While this approach does take into account some significant properties of the current, the shape-factor is generally taken as constant in these numerical experiments while in natural turbidity currents both velocity and sediment concentration profiles may evolve with time. Such models are therefore quasi-two-dimensional, not truly two-dimensional as the vertical variation in properties is not actually modelled. Depth-averaged models account for turbulence by employing an empirically derived function for the entrainment of ambient fluid at the upper boundary (e.g. see Ellison & Turner, 1959; Fukushima et al., 1985). Depth-averaged models are generally solved numerically (i.e. by discretizing the governing equations, e.g. Parker et al., 1986; Zeng & Lowe, 1997). However, by making many simplifying assumptions (such as steady flow), the depth-averaged Navier–Stokes equations may be reduced to terms that may be solved analytically (see Stacey & Bowen, 1988a). While complex numerical solutions may require considerable computational expense, they may be more appropriate than analytical solutions for gravity currents which are typically highly unsteady.

In the turbidity current literature, three-equation depth-averaged models are most common, in which the equations for fluid mass, sediment mass and fluid momentum are solved simultaneously (Ellison & Turner, 1959; Fietz & Wood, 1967; Chu et al., 1979; Parker et al., 1986; García, 1994; Zeng & Lowe, 1997). In four-equation models, conservation of turbulent energy is also taken into account (Johnson, 1962; Parker et al., 1986). Most depth-averaged models have been used to study aspects of the evolution of turbidity currents. Ellison & Turner (1959) developed an early model for studying entrain-
ment of ambient fluid at the upper boundary of saline currents. Their theoretical results were combined with experimental data to produce a semi-empirical entrainment function and a series of equations that could be solved to predict current velocity and thickness. Chu et al. (1979) modelled flow evolution, examining the transition from supercritical flow on a slope to subcritical flow on the basin floor via a hydraulic jump, and the effect of changing variables such as slope, sediment exchange with the bed and entrainment at the upper boundary on this development. Self-acceleration (or auto-suspension after Pantin, 1979) has been studied theoretically in a similar way (Parker et al., 1986); a four-equation model linking sediment entrainment to mean turbulent energy appeared to give realistic results. A recent model (Zeng & Lowe, 1997) predicts current development and sediment deposition, producing detailed models of deposit thickness and grain-size distribution in both the downstream and the vertical directions.

Depth-averaged models, while not actually modelling the fluid dynamic processes within turbidity currents, can produce relatively accurate predictions of current evolution and deposit characteristics which provide valuable insights for geologists concerned with ancient turbidity current deposits.

Models incorporating turbulence

A second approach to solving the Navier–Stokes equation for turbulent currents is to use a model that accounts for turbulence and hence can resolve vertical current properties. This can be achieved analytically; for instance, Hinze (1960) reduced the Navier–Stokes equations to model a steady current with constant eddy viscosity in the vertical profile. Similarly, Stacey & Bowen (1988a) assumed eddy viscosity and eddy diffusivity were constant in the vertical direction and produced an analytical solution for unsteady currents. However, they demonstrated that this approach is of limited value since the velocity and sediment concentration profiles are decoupled in the model, which is unrealistic. Stacey & Bowen (1988a) also developed a numerical model in which eddy diffusivity and eddy viscosity vary vertically as a function of the gradient Richardson number. This model produced vertical velocity and sediment concentration profiles that compared well with Ellison & Turner’s (1959) experimental data, and gradient Richardson number profiles that compare well with Buckee et al. (in press; Fig. 7). Several more complicated turbulence closure models have been applied to turbidity currents (for a review of turbulence closure models see Rodi, 1980). Eidsvik & Brørøs (1989) used a k–ε model (Rodi, 1980), and Brørøs & Eidsvik (1992) used a Reynolds stress model; both models produce realistic results when compared with experimental data sets (e.g. Fig. 6). Interestingly, some features predicted by these models, such as low turbulent kinetic energy at the level of the velocity maximum, were initially dismissed by the authors as unrealistic but have since been observed in laboratory experiments (Parsons, 1998; Kneller et al., 1999; Best et al., in press; Buckee et al., in press; Fig. 6). More recently, Felix (in press) has adapted the Mellor–Yamada level 2½ closure model (Mellor & Yamada, 1982) to include the influence of suspended sediment on turbulence and to model current development in both the vertical and the downstream directions. This approach allows a unique two-dimensional representation of the properties of the whole current (Fig. 8), including the head, and consequently can predict spatial variability (in terms of thickness and grain size) of turbidite deposits.

A turbulence-based model is appropriate for a situation in which there is significant vertical variation in the current properties, and hence is ideal for use with turbidity currents, and may lead to better understanding of turbidity current dynamics (Stacey & Bowen, 1988a). A limitation of the use of turbulence models is that the methods are currently very costly in terms of computer power and time. In many cases the required model output (such as grain size distributions or deposit thickness) does not warrant the computational power required, given the uncertainties over initial conditions of natural currents in the environment, and a more simple depth-averaged model would serve the purpose in modelling.

EFFECTS OF TOPOGRAPHY ON TURBIDITY CURRENTS

In any system in which the topography encountered by sediment gravity flows is not completely flat, the topography exerts a potentially major influence on deposition (e.g. Miller & Smith, 1977; Pickering et al., 1989; Fisher, 1990; Thornburg et al., 1990; Apps et al., 1994), either
Fig. 8. A 2D time-slice of a turbidity current travelling from left to right, (a) downstream velocity (in m s^{-1}) and (b) total volume concentration fraction (adapted from Felix, in press). The model current is initiated as a static suspension (1000 m x 50 m) at 3% volume concentration, composed of three grain sizes in equal proportions. This image is taken 2000 s after initiation, after the current has passed over an obstacle of 100 m x 5 m. Notice the high velocity core near the front of the current in (a) and the general decrease in mean velocity and sediment concentration backwards from the head.

by controlling the nonuniformity of the currents or by confining and 'ponding' them, either partially, as in a valley, or completely (e.g. Wilson & Walker, 1985; Branney & Kokelaar, 1992; Rothwell et al., 1992; Hodgson et al., 1992; Prather et al., 1999). The behaviour of turbidity currents and pyroclastic currents around obstructing topography varies with the forward velocity of the current, the obstacle height, the current density and, most significantly, the density stratification within the current (Long, 1953, 1955, 1970; Valentine, 1987; Grue, 1992; Lawrence, 1993; Castro & Snyder, 1993; Lane-Serff et al., 1995; see also Muck & Underwood, 1990; Alexander & Morris, 1994). This topographic interaction has major implications for the spatial distribution of sediment in the deep sea, for the interactions of unconfined currents with intrabasinal highs and basin margins, and for channellized currents with channel margins and levées. It is also relevant to hazard mitigation in volcanic areas.

Run-up heights

For obstacles that are much larger than the current, oceanographic data suggest that run-up distances in nature may be many hundreds of metres (Dolan et al., 1989; Muck & Underwood, 1990; Lucchi & Camerlenghi, 1993). Similarly, pyroclastic currents may surmount topographic ridges several hundreds of metres high (e.g. Miller & Smith, 1977; Wilson & Walker, 1985). According to Lane-Serff et al. (1995), a finite volume of fluid related to the head may surmount the obstacle if H/ (the ratio of the obstacle height to the current body thickness) is less than 4 or 5, or if the obstacle height is less than 1.5 times the height of the head of the current, according to Muck & Underwood (1990). Both these studies reported the results of experiments with saline currents. However, it seems likely that in relatively poorly stratified currents, the maximum run-up height is probably dependent on the bulk Froude number of the current:

\[ h_{\text{max}} = \frac{1}{2} \left( \frac{U^2}{g'} \right) \]  

(Rottman et al., 1985). For strongly stratified currents, Kneller & McCaffrey (1999) developed Allen’s (1985) energy balance arguments to suggest an expression for maximum run-up
Fig. 9. Nomenclature of three types of bores defined by Rottman & Simpson (1989) (modified from Edwards, 1993).

height that takes account of both the velocity and the density profiles of the current:

\[ h_{\text{max}} = y + \frac{\rho_y u_y^2 (1 - E)}{2g\Delta\rho_y} \]  

(7)

where \( u_y \) is the downstream component of velocity at initial height \( y \), \( g \) is the acceleration due to gravity, \( \Delta\rho_y \) is the density difference between the fluid at initial height \( y \) and the ambient fluid, and \( E \) is the fractional energy loss due to friction. The value of \( y \) that yields the highest run-up of any parcel of fluid in the current is given by:

\[ \frac{d}{dy} \left[ y + \frac{\rho_y u_y^2 (1 - E)}{2g\Delta\rho_y} \right] = 0 \]  

(8)

\( h_{\text{max}} \) is thus a function of the velocity and density profiles, and cannot be generalized in systems in which the density stratification is significant – as in most sediment-laden systems.

The blocking of turbidity currents by topography

Complete blocking is when none of the current gets over the top of the topography. For two-dimensional systems, i.e. where the obstacle axis is perpendicular to flow and effectively infinite in extent, Rottman et al. (1985) give the criterion for complete blocking as:

\[ \frac{U}{g'd} = 0.5(H_c - 1)^2 \left( \frac{H_c + 1}{H_c} \right) \]  

(9)

Partial blocking occurs where the relative obstacle height is less and part of the forward current continues over the obstacle. Woods et al. (1998) provide an analysis of the criteria for full and partial blocking of unstratified currents by two-dimensional topography, based on the concept of critical flow velocity \( (u_c, \text{Eq. 11}) \) and critical flow thickness \( (h_c, \text{Eq. 13}) \) that are used to nondimensionalize the mean flow velocity \( (\overline{U}) \) and obstacle height \( (H) \), thus:

\[ U_o = \frac{\overline{U}}{u_c} \]  

(10)

where

\[ u_c = (g'Q)^{1/3} \]  

(11)

(where \( U_o \) is the nondimensional downstream velocity and \( Q \) is the volume flux), and,

\[ D_0 = H/h_c \]  

(12)

where

\[ h_c = Q/u_c \]  

(13)

where \( D_0 \) is the nondimensional height. The minimum height for complete blocking is dependent on \( U_o \) and \( D_0 \), and is equal to the thickness of the bore generated upstream (Woods et al., 1998, and see below), which is \( \sim 2.2h_c \) for critical currents. The situation is more complex for three-dimensional topography; flow probably goes around the obstacle when the densiometric Froude number is much less than 1, but stratification plays an important role (see below).

The effect of blocking is to generate a disturbance upstream of the obstacle, which consists of an internal bore (a more or less abrupt downstream increase in current thickness with an accompanying drop in velocity) that migrates upstream. Where the current is completely blocked, the fluid velocities within the bore may be negative (i.e. upstream) and it therefore constitutes a current ‘reflection’ (Edwards, 1993; Best et al., in press).
Kneller et al. (1991), Edwards (1993) and Kneller (1995) have shown that in three-dimensional situations, these ‘reflections’ propagate perpendicular to the reflecting surface. The bore is often referred to as a moving internal hydraulic jump, though it does not necessarily separate regions of supercritical and subcritical flow (defined as regions where the densiometric Froude number is, respectively, greater than or less than unity).

The bore may take different forms (Fig.9) determined by the ratio of the height of the bore to the thickness of the forward current through which it moves (Fig.9; Wood & Simpson, 1984; Edwards, 1993; Edwards et al., 1994; Best et al., in press; see also Simpson, 1997; Rottman & Simpson, 1989). This ratio in turn depends on the speed of the forward current, the current density and the upstream slope of the obstacle (Edwards et al., 1994). The strongest, Type C in the nomenclature of Edwards et al. (1994), consists of an undercutting turbulent gravity current head that generates a bore within the overlying dense fluid (Fig.9a). Considerable mixing with the ambient fluid may occur at the upper boundary of a Type C bore (see, e.g. Best et al., in press). Type B, sensu Edwards et al. (1994) is an undular bore, and has a rather smoother front with distinct undulations behind it (Fig.9b). Lastly, Type A bores consist of a series of internal solitary waves (Fig.9c). These types form a continuous spectrum. To date, no investigation has examined the effect of varying density stratification on the type of bore produced, but it seems likely that strong bores are associated with stronger stratification.

Multiple palaeocurrent directions within single beds have been reported from many turbidite systems, both ancient (Marjanac, 1990; Kneller et al., 1991; Edwards, 1993) and modern (Pickering et al., 1992), and are probably present in many more. These multiple-palaeocurrent directions have been interpreted as ‘reflection’ of the turbidity current from topography at the basin margin; comparable interactions with topography have also been described for pyroclastic currents (Valentine et al., 1992; Woods et al., 1998). In some cases, the ‘reflections’ are more or less diametrically opposed to the forward current (e.g. Ellis, 1982; Pickering & Hiscott, 1985), while in many cases they are roughly perpendicular. Often the changes in current direction occur in association with abrupt reversals in grading (Pickering & Hiscott, 1985), and these reversals or repetitions have been used to imply current reflection.

Kneller et al. (1997) discussed the possibility that internal solitary waves of Type A may be responsible for the creation of wave ripples in deep water – as undoubtedly occurs (Ricci Lucchi, 1981; Fujioka et al., 1989). In lock-exchange experiments, surge-type currents generated only a short train of 3–4 waves; it is not clear whether a sustained current would create a more continuous wave train. It is unlikely that such a short wave train could generate ripples from a flat bed, but they would readily modify the form of pre-existing current ripples. Particle velocities within a solitary wave are given by:

$$U = \frac{Hg'(h + H)^{0.5}}{3h}$$  \hspace{1cm} (14)

(Allen, 1982, p. 35) where $h$ is the thickness of the lower fluid layer (in this case the residual current), $H$ is the wave height and $g'$ is reduced gravity $g[\Delta \rho/\rho_s]$. Edwards et al. (1994) demonstrated that $H$ shows some dependence upon the slope of the obstruction at which the wave was generated. Conservative values of $g'$ (density differences of a few per cent) and $h$ and $H$ (of the order of a few metres) yield estimates of velocity and bed shear stress capable of moving sand in traction.

A number of examples in the rock record consist of symmetrical ripples associated with what in all other respects look like deep-water sequences (Ricci Lucchi, 1981; Kneller et al.,...
1991); at least some of these may have been deposited well below storm wave base. Deep-sea photographs of symmetrical ripples in water depths of thousands of metres (e.g. Fujioka et al., 1989), possibly formed by this mechanism, cast doubt on the reliability of wave ripples as water-depth indicators.

**Stratified flow around 3D topography**

Where the relative height of the obstacle is small (of the same order as the inner region of the current), the behaviour of the current around three-dimensional topography can probably be treated in the same way as stratified boundary layer flows. In these circumstances, behaviour both upstream and downstream of topography can be related to an internal Froude number (see, e.g. Baines, 1995), generally defined as:

\[
Fr_i = \frac{U}{NH}
\]  

(15)

where \(U\) is the current velocity, \(H\) is the obstacle height, or some more general length scale \((y_s)\) such as the dimensionless depth \((\eta)\) over which a known density difference \((\Delta \rho_i)\) occurs (Valentine, 1987):

\[
y_s = \frac{\Delta \rho_i d}{(d \rho / d \eta)}
\]  

(16)

where \(\eta = (y/d)\). \(N\) is a parameter known as the Brunt–Väisälä frequency, or simply the buoyancy frequency, which is the frequency of unforced gravity waves on the density stratification, given by:

\[
N^2 = \left(\frac{\partial \rho / \partial y}{g \rho_a}\right)
\]  

(17)

where \(g\) is the acceleration due to gravity, \(\partial \rho / \partial y\) is the vertical component of the density gradient, and \(\rho_a\) is a reference density (normally taken as the density of the ambient fluid).

Hunt & Snyder (1980) used an internal Froude number criterion to describe strongly stratified atmospheric flow over hills (see also Baines, 1979; Snyder et al., 1985). For small values of \(Fr_i\), a critical plane (‘dividing streamline’ of Hunt & Snyder, 1980) exists within the current, above which fluid particles have sufficient kinetic energy to move up and over obstructing topography, whereas in the denser, lower regions of the current, fluid particles have insufficient kinetic energy to rise up the topography, and are deflected around it. The height of this dividing plane, \(H_s\), is approximated by:

\[
H_s = H (1 - \theta Fr_i)
\]  

(18)

where \(\theta\) is constant with a value of about 1 (Hunt & Snyder, 1980), although it has been assigned values from 0.7 to 2.0, and may depend on the shape of the obstacle (Baines, 1979; Snyder et al., 1985). For small values of \(Fr_i\), all the fluid below the dividing plane moves around the barrier, while that above this level moves over it in an essentially two-dimensional pattern (Fig. 10). For values of \(Fr_i\) greater than \(1/\theta\) there exists no dividing plane, and fluid at all levels moves over the obstacle; where there are gaps between obstacles, only the fluid opposite the gap moves through it, i.e. there is effectively no deflection (Baines, 1979).

**Downstream effects of topography on gravity current behaviour**

The downstream behaviour of a current is also dependent upon the internal Froude number
(Fig. 11), although the threshold values of $F_r$ separating these different behaviours varies with the form of the topography in three dimensions. In the simplest case, where the topography is small enough that $F_r$ is very large ($\gg 1$; H. likely to be less than $\sim 0.1$), simple divergence of streamlines occurs on the downstream side of the topography (Long, 1955). This is likely to be the case over small downstream steps in topography.

At slightly lower $F_r$ (Fig. 11b), flow separation occurs downstream of the topography (Lawrence, 1993), in exactly the same fashion as occurs downstream of ripples and dunes in shear flows. The probable velocity and density scales in turbidity currents mean that flow separation is likely in similar situations, i.e. downstream of scours and large bedforms such as dunes or gravel waves, assuming currents are of sufficiently low density that turbulence is not suppressed near the bed. The higher velocities of pyroclastic currents lead to flow separation on a much larger scale beneath blasts and surges (e.g. Lipman & Mullineaux, 1981).

At still lower $F_r$ (larger obstacle sizes and density gradients, lower velocities; Fig. 11c), flow over the lee side becomes supercritical (Castro & Snyder, 1993), and lee-waves occur downstream of the obstacle, whose wavelength is a function of the internal Froude number (see also Huppert & Miles, 1969). Flood (1988) suggested a lee-wave origin for fine-grained sediment waves, generally having wavelengths of 500–3000 m and heights of 10–100 m (McCave & Tucholke, 1986; see also Blumtack & Weatherly, 1989), that occur in many slope, base of slope and levee settings. A lee-wave model satisfactorily explains the commonly observed up-current migration of mud waves. Additionally, in some cases these waves can plausibly be related to turbidity current activity in the lee of topography, e.g. mud-waves on the Var sedimentary ridge formed by turbidity current over spill from the adjacent fan valley (cf. Savoye et al., 1993; who suggest an antidune origin), and sediment waves on the right-hand levee of the Hueneme fan channel (Piper et al., 1999b), in both cases apparently fed by hyperpycnal flow. Wherever such waves appear to migrate upslope, it is likely they have formed beneath gravity-driven currents.

**Hydraulic jumps**

At the lowest values of $F_r$ a hydraulic jump occurs downstream of the obstacle (Fig. 11d), where fluid goes through an abrupt downstream deepening; this is effectively a breaking lee wave. For three-dimensional cases, downstream wave breaking is more likely for objects with a larger width-to-length ratio (Castro & Snyder, 1993).

Several authors have described the existence and mechanics of internal hydraulic jumps in stratified or multilayer experimental systems (Long, 1953; Yih & Guha, 1955; Wood & Simpson, 1984), and the effects of such jumps on sediment deposition from both steady (García & Parker, 1989; García, 1993) and highly unsteady experimental turbidity currents (Alexander & Morris, 1994; see below). The possibility of hydraulic jumps in natural systems has been postulated by Middleton (1970), amongst others (see also Menard, 1964; Komar, 1971, 1975; Hand, 1974, 1975), and the analytical model of steady turbidity current flow of Chu et al. (1979) suggested the existence of a hydraulic jump between the uniform flow of the slope or channel ($F_r > 1$) and the nonuniform flow of the basin plain ($F_r < 1$). Experimental data from García & Parker (1989) and García (1993) suggest that while increased bedload deposition coincides with the jump (i.e. where shear velocity falls), the zone of enhanced deposition of suspended load is displaced slightly downstream of the jump. This may be a function of the lag-time for material to settle through the current (i.e. a function of $U_*/U_0$), or the decay time for the additional turbulence (and thus high Reynolds stress) associated with the jump. A number of authors invoke hydraulic jumps at slope breaks or channel mouths to explain sediment bypassing (e.g. Mutti & Normark, 1991; Mutti, 1992) possibly associated with scour (e.g. Kenyon et al., 1995; Vicente-Bravo & Robles, 1995; see also Shor et al., 1990). If the buoyancy frequency, $N$, varies vertically, several different effects may occur in different parts of the current, for example flow separation may occur on the downstream-facing slope of the topography, with a hydraulic jump forming further downstream, immediately beyond the break in slope.

**Implications for sedimentation**

Where turbidity currents interact with large obstacles and are partially blocked, they will inevitably experience a massive and rapid decrease in both competence and capacity associated with the upstream jump, and sedimentation is likely to occur here. This is supported by the observations of Alexander &
Morris (1994), Kneller (1995) and Kneller & McCaffrey (1995) that show marked localization of deposition associated with the jump, despite the sudden increase in turbulence that would be expected there. In rapidly waning currents, the jump may remain almost stationary where the current strikes the obstacle at a high angle; however, in steadier currents the jump – and associated deposition – might migrate upstream, producing a less localized deposit. Rapidly waning currents may also produce higher energy jumps (Type C or B) since they propagate through the thinning tail of the forward current and so the ratio of jump height to forward current thickness progressively increases.

Deposition may similarly be concentrated in the lee of the obstacle in association with the downstream hydraulic jump, possibly producing an abrupt downstream thickening in the deposit. However, sedimentation from currents that remain subcritical even over the lee side of the obstacle (slower, thicker or denser currents) might occur over the entire lee slope of the obstacle where flow separation occurs. Deposition beneath supercritical flow on the lee side of obstacles may produce antidune stratification, whose preservation in the sediment thus need not imply that the current was supercritical except on the downstream facing slope of the obstacle.

DISCUSSION

The debate over the nature and importance of high-density dispersions dates back at least to the introduction of the term ‘fluxoturbidite’ (e.g. Carter, 1975), and seems set to continue until experimental studies and theory together can elucidate their mobility and manner of deposition. Distinction between dense and dilute flows on the basis of their deposits alone is a hazardous business; the deposits themselves are likely to reflect only transient conditions in the depositional regime (only the lowermost part of the current above progressively aggrading deposits), and may say nothing directly about the transport regime. At any rate, existing theory seems inadequate to explain the behaviour of some highly mobile dense dispersions, and arguments based solely on the geological interpretation of deposits may be inadequate to resolve issues of process. The link between textural features of the deposit on the one hand and process on the other is somewhat indirect and in many cases poorly understood. Some geological features that may allegedly be used to differentiate between different processes are susceptible to misinterpretation. For example, normal grain-size grading in a turbidite does not necessarily arise in the same way as that produced by static settling of a multisize dispersion in the absence of any effective bed shear stress (thus depending on the vertical structure of the dispersion), but by a temporal decline in bed shear stress at a point beneath a waning (but still moving) current – hence the presence of current-induced structures in many turbidites. Such grading thus depends on the longitudinal structure of the current, which is likely to be dependent on the initiation mechanism and downstream evolution of the flow.

The relative importance of hyperpycnal flow compared with other initiation mechanisms is also unclear, and is likely to vary with the configuration of the shelf and drainage basin. The narrowing of shelves during sea-level lowstands may militate in favour of hyperpycnal flow in general, though Mulder & Syvitski (1996) considered that the changing balance of discharge and suspended sediment load reduced the likelihood of hyperpycnal flow in any given river system. Recognition of ancient deposits produced by hyperpycnal flow is still conjectural. Modelling may help to reduce the uncertainty in the interpretation of deposits, but the compromises involved in both experimental and mathematical approaches mean that the results may describe only limited aspects of the prototype. Experimental studies are also limited by the physical difficulties involved in making the required observations. Many model studies necessarily involve a certain lack of geological realism that must be taken into account when applying the results to natural systems. However, experimental observations on currents can help alert us to factors that cannot generally be neglected, such as the existence of density stratification and variations in its stability. In this context, it is again worthy of note that stable stratification does not imply the absence of turbulence.

The use of bulk (i.e. depth-averaged) parameters, such as overall flow Richardson number or even flow Reynolds number, may be inappropriate under many circumstances. Many currents are better characterized by parameters that capture the vertical density and velocity structure, such as gradient Richardson number and internal Froude number. The vertical velocity and turbulence structure of turbidity currents is
fundamental to the manner in which sediment is distributed within the current, and thus affects many aspects of deposition, especially where the presence of topography means that different levels in the current are in contact with the bed. This includes depositional topography such as the levées of submarine channels. The stratification of currents flowing down levées channels is instrumental in partitioning different grain-sizes of sediment between the channel axis and the levée (e.g. Peakall et al., in press, and references therein). However, $F_r$ also determines the degree of over spill on channel bends from the height of the dividing streamline with respect to the levée crest (Fig. 10) and the behaviour of the portion of the current that overtops the levée (Fig. 11).

Despite differences in structure between suspension currents and open channel flows, expressions such as the law-of-the-wall (when suitably modified for stratified flow; Turner, 1973), and the quadratic stress law are probably reasonable approximations when treating processes such as near-bed sediment distribution, bedload transport and erosion, which are dominated by the near-bed region of the current, since the velocity structure of the inner region of a suspension current is similar to that of a shear flow (Kneller et al., 1999).

Technological advances have facilitated increasingly useful models of density current behaviour, both experimentally and numerically. Experiments show that the turbulence structure in density currents is highly complex. Further work is required, particularly with respect to sediment-laden currents and the interaction between the turbulence structure and suspended particles. Turbulence-based numerical models appear to predict realistically the dynamic structure of gravity currents (e.g. Stacey & Bowen, 1988a,b; Eidsvik & Brørs, 1989; Brørs & Eidsvik, 1992; Felix, in press) and hence are more likely to be useful in modelling turbidite deposition. Experimental and numerical models are, however, limited by the lack of testing against environmental data, owing to the large scale of natural turbidity currents and the inherent difficulties in comparing model output with turbidite deposits for which the initial conditions are largely or wholly unknown. The future surely lies in a mix of experimental and theoretical modelling, along with increasingly quantitative field studies. Ultimately, however, there is no substitute for direct observation; full instrumentation of natural currents in the deep sea remains one of the most exciting prospects for the future.

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**NOMENCLATURE**

- $C$ Fractional sediment concentration
- $d$ Flow thickness
- $d_h$ Mean height of the bore waveform
- $D_o$ Nondimensionalized height
- $E$ Fractional energy loss due to friction
- $f_l$ Darcy–Weisbach friction coefficient at the lower boundary
- $f_i$ Darcy–Weisbach friction coefficient at the upper boundary
- $F_r$ Froude number
- $F_r'$ Densiometric Froude number
- $F_{ri}$ Internal Froude number
- $g$ Acceleration due to gravity
- $g'$ Reduced gravity
- $h$ Thickness of the lower fluid layer
- $h_c$ Critical flow thickness
- $h_{max}$ Maximum run-up height
- $H$ Obstacle height
- $H_s$ Ratio of obstacle height to current body thickness
- $H_s$ Height of critical dividing plane
- $H$ Wave height
- $k$ Turbulent kinetic energy
- $N$ Brunt–Väisälä frequency
- $Q$ Volume flux
- $Re$ Flow Reynolds number
- $Ri_g$ Gradient Richardson number
- $S$ Slope
- $u$ Instantaneous downstream velocity
- $u_c$ Critical flow velocity
- $U_y$ Component of downstream velocity at initial height $y$
- $U_s$ Shear velocity
- $U_s$ Grain settling velocity

\( U \) Particle velocity within a wave
\( \mathcal{U} \) Depth-averaged downstream velocity
\( U_{\text{max}} \) Maximum mean downstream velocity
\( U_0 \) Nondimensionalized downstream velocity
\( U_y \) Mean downstream velocity at height \( y \)
\( v \) Instantaneous vertical velocity
\( y_{1/2} \) Distance between the bed and the level in the outer region at which the downstream velocity is half the maximum downstream velocity
\( y_s \) Length scale in internal Froude number
\( \gamma \) Height
\( \theta \) Constant
\( \Delta \rho \) Density difference between the current and the ambient fluid (\( \rho_c - \rho_a \))
\( \Delta \rho_y \) A known density difference
\( \Delta \rho_f \) Density difference between the fluid at height \( y \) and the ambient fluid
\( \eta \) Dimensionless depth, \( y/d \)
\( \rho_a \) Density of the ambient fluid
\( \rho_c \) Mean current density
\( \rho_0 \) Reference density
\( \rho_f \) Current density at initial height \( y \)
\( \mu \) Molecular viscosity of water
\( \mu_s \) Apparent viscosity of a clay-free dispersion


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