Large floating clasts in turbidites:
a mechanism for their emplacement

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Abstract

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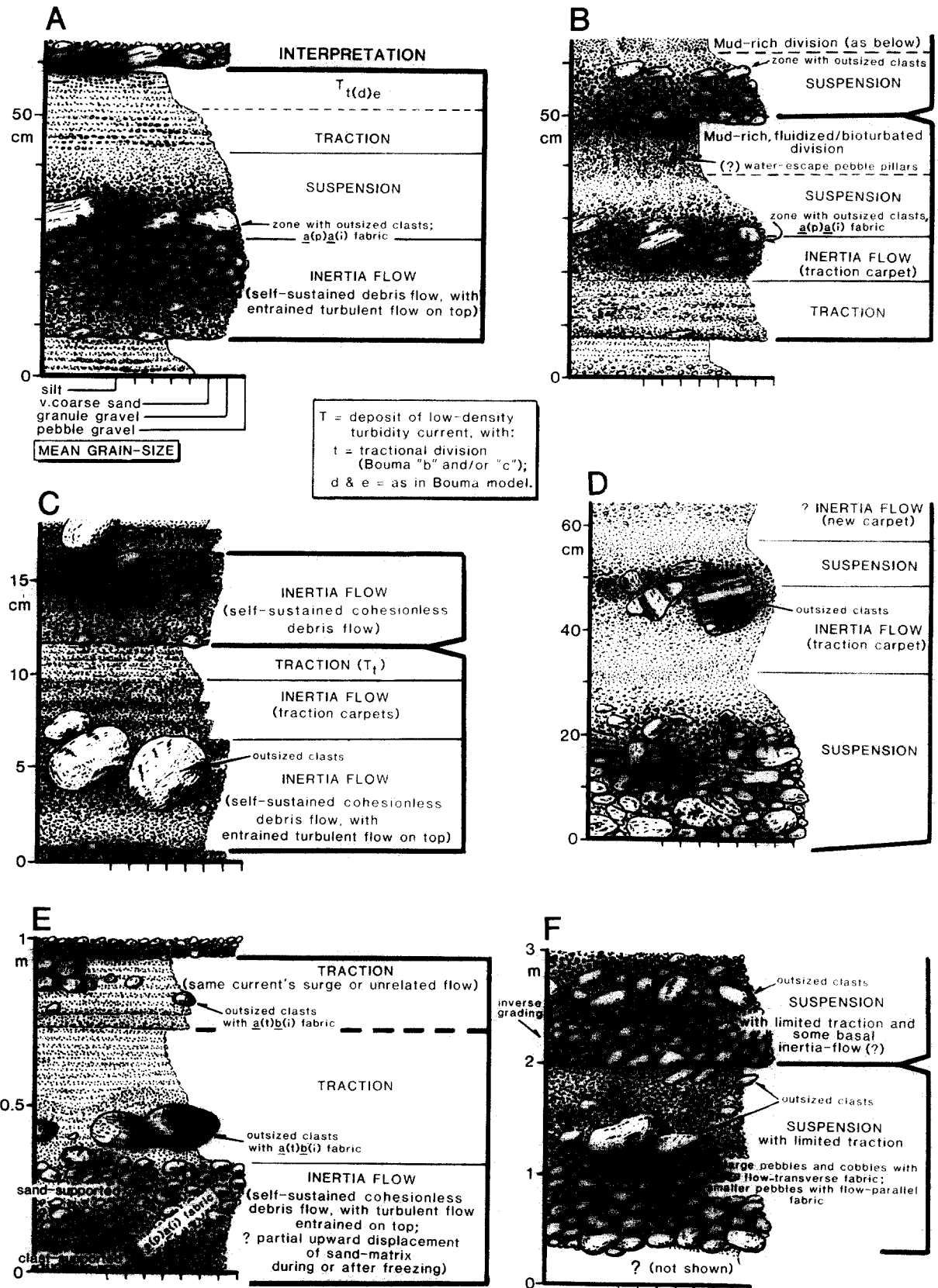
The transportation mode for large, isolated clasts "floating" in turbidites is a problem that has been difficult to resolve. New observations from experimental, high-density turbidity currents indicate that large, outsized clasts can be transported along a rheological interface which develops within the flow. The clasts "glide" along the top of an underlying, pseudolaminar inertia-flow layer, partly submerged in it, and are driven by the downflow component of turbulent shear-stresses transmitted from the overlying, faster-moving turbulent layer. As the inertia-flow layer freezes and a new one forms, or as the layer thickens, the gliding clast may be forced to a progressively higher level within the flow. With deceleration, the inertia-flow phase of the flow freezes entirely and the large clast is then trapped "suspended" above the base of the resulting turbidite. Both $a(p)a(i)$ and $a(t)b(i)$ orientations of the floating clasts are observed. It is suggested that, hypothetically, a similar mechanism for megaclast emplacement may also operate in other types of viscous sediment flows, subaqueous or subaerial, wherever there is a distinct rheological interface developed within the flow.

Introduction

The deposits of fully turbulent sediment gravity-flows, such as turbidity currents, characteristically display normal grading — a signature of the strong inherent tendency for the coarser grains to settle fastest through the turbulent sediment suspension. Consequently, any large or outsized clasts suspended in a turbulent flow are expected to fall out rapidly from the suspension once it decelerates, and to be either deposited directly at its base or transported farther by bed-load rolling, sliding, or possibly saltating. Although the turbulent action itself tends to homogenise the flowing sediment–water mixture, it also segregates smaller grains from the larger ones (Newitt et al., 1962; Fisher, 1971; Middleton and Southard, 1978). As

shown by Middleton (1967), it is not even necessary that the head of the turbidity current be decelerating in order to cause the settling of the coarser grains and to form a normally graded deposit (though possibly with an ungraded or even inversely graded basal part, depending on the actual settling intensity and bed-load behaviour; see Lowe, 1982). Rapid settling of the coarser material takes place due to an abrupt decrease in velocity associated with the passage of the "wave" directly following the head of the flow.

The transportation mode for the large clasts *floating* ("suspended") in turbidites is, therefore, a problem that has been particularly difficult to resolve, as recently emphasised by Hein (1982) and other authors. Many turbidites appear to contain floating megaclasts, extra- or intra-forma-

A

50 — cm

0 —

INTERPRETATION

$T_{t(d)e}$

TRACTION

SUSPENSION

zone with outsized clasts; a(p)a(i) fabric

INERTIA FLOW
(self-sustained debris flow, with entrained turbulent flow on top)

silt
v.coarse sand
granule gravel
pebble gravel

MEAN GRAIN-SIZE

T = deposit of low-density
turbidity current, with:
t = tractional division
(Bouma "b" and/or "c");
d & e = as in Bouma model.

B

50 — cm

0 —

Mud-rich division (as below)
zone with outsized clasts

SUSPENSION

Mud-rich, fluidized/bioturbated division
(?) water-escape pebble pillars

SUSPENSION
zone with outsized clasts, a(p)a(i) fabric

INERTIA FLOW
(traction carpet)

TRACTION

C

15 — cm

10 —

5 —

0 —

INERTIA FLOW
(self-sustained cohesionless debris flow)

TRACTION (T_t)

INERTIA FLOW
(traction carpets)

outsized clasts

INERTIA FLOW
(self-sustained cohesionless debris flow, with entrained turbulent flow on top)

D

60 — cm

40 —

20 —

0 —

? INERTIA FLOW
(new carpet)

SUSPENSION

outsized clasts

INERTIA FLOW
(traction carpet)

SUSPENSION

E

1 — m

0.5 —

0 —

sand-supported

clast-supported

a(p)a(i) fabric

TRACTION
(same current's surge or unrelated flow)

outsized clasts
with a(t)b(i) fabric

TRACTION

outsized clasts
with a(t)b(i) fabric

INERTIA FLOW
(self-sustained cohesionless debris flow, with turbulent flow entrained on top;
? partial upward displacement of sand-matrix during or after freezing)

F

3 — m

inverse grading

2 —

1 —

0 —

outsized clasts

SUSPENSION
with limited traction and some basal inertia-flow (?)

outsized clasts

SUSPENSION
with limited traction

large pebbles and cobbles with flow-transverse fabric; smaller pebbles with flow-parallel fabric
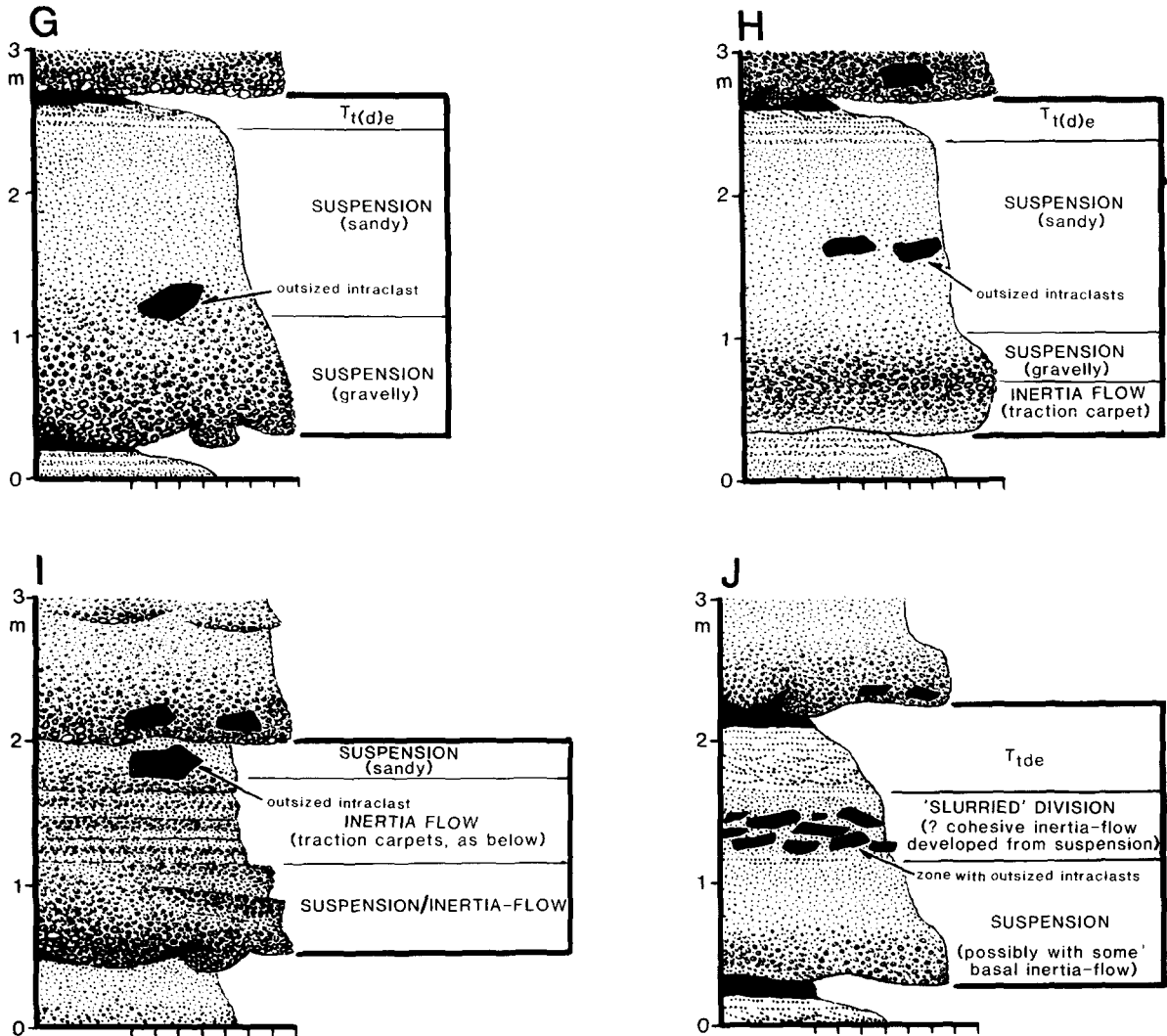
? (not shown)

Fig. 1. Examples of deposits of inferred high-density turbulent flows containing large (outsized), floating clasts. The sketches show: A. Subaqueously deposited, sediment gravity-flow unit from the lacustrine Karlskaret fan delta, Devonian Hornelen Basin, Norway. B. Deposits of high-density turbidity currents from the marine Espiritu Santo Gilbert-type fan delta, Pliocene, Spain (see Postma and Roep, 1985; similar deposits from other mass flow-dominated fan deltas are reported by Postma, 1984). C. Deposits of cohesionless debris-flows (density-modified grain-flows) and associated, high-density turbidity currents from the lacustrine Domba fan delta, Devonian Hornelen Basin, Norway (see Facies C in Nemec et al., 1984). D. Main portion of a deposit of high-density turbidity current from the Wollaston Foreland Group submarine fan, Jurassic-Cretaceous, Greenland (drawn from Surlyk, 1984, fig. 18; interpretation slightly modified). E. Typical deposit of a sediment gravity-flow from the Carmelo Formation submarine canyon-fill, Paleocene Point Lobos, California (drawn from the description by Clifton, 1984, and his fig. 14; interpretation added). F. Deposits of subaerial, highly concentrated turbulent flows ("hyperconcentrated flood-flows") from the Neogene Deschutes Formation, Oregon (drawn from the description by Smith, 1986, and his fig. 2B; interpretation added). G-I. Deposits of high-density turbidity currents ("fluxoturbidites") from the Palaeogene Ciężkowice Sandstone, Silesian Nappe Flysch, Polish Carpathians (compiled from Leszczyński et al., 1986, figs. 4 and 5). J. A relatively common type of turbidites in the Palaeogene Magura Nappe Flysch, Polish Carpathians (after Porębski, written communication, 1985; interpretation added).

tional, that are more than an order of magnitude larger than the actual mean grain size of their "host" deposit (Fig. 1). Reported examples include the deposits of inferred high-density turbid-ity currents that contain isolated, floating mega-clasts up to a few decimetres or even a few metres in their longest dimension (e.g., Piper, 1970; Davies and Walker, 1974; Stanley et al., 1978; Winn and

Dott, 1978, 1979; Hein, 1982; Massari, 1984; Clifton, 1984). The occurrence of large "suspended" intraclasts seems particularly puzzling, as such semi-consolidated components should easily become disintegrated in a fully turbulent, concentrated sediment suspension.

On the basis of our observations from a series of flume experiments, we propose a likely mechanism for the transport and deposition of large clasts which are floating *within*, not merely dragged at the base of, a high-density turbidity current. The implications may also be relevant to other high-concentration turbulent flows, either subaqueous or subaerial.

Laboratory experiment: observations and inferences

The small-scale experiments were performed by the serior author (G.P.), using natural pebbly sand in a tilted, glass-walled flume (4 m long, 60 cm deep and 25 cm wide), at the Geological Institute, University of Amsterdam. The sediment was of glacio-fluvial origin, and was taken from the "Ullenberg" pit of Veluwe, The Netherlands (for sedimentological details, see Postma et al., 1983). An admixture of chalk powder (5 vol.%) was subsequently added, when preparing the experimental sediment-water mixture of a uniform concentration of approximately 35–40 vol.% solids. The chalk powder was added to prevent an instantaneous settling of the suspension in the sediment-tank compartment of the flume, before opening the valve (ca. 10 s time-lapse). The grain-size distribution of a bulk sediment sample is given in Table 1. Pebbles (quartz and crystalline rock fragments) had a maximum size of about 2.5 cm (*a*-axis), were generally subrounded and had a low sphericity. The temperature of water in the flume was 15°C.

The technique and filmed results of the experiments were presented by the senior author at the I.A.S. 4th European Meeting (1983) in Split, Yugoslavia, and discussed in a later paper (Postma, 1984). In the present paper, we refer to those aspects of the experimental turbidity currents that relate to the transport mechanism of outsized "floating" clasts.

TABLE 1

Grain size composition of the sediment used in the laboratory experiment

Grain size fraction	Wt.%
Gravel coarser than 16 mm	5
Gravel, 4–16 mm	40
Granules (2–4 mm)	15
Very coarse sand	6
Coarse sand	6.5
Medium sand	10
Fine sand	10.5
Very fine sand	4.5
Unsieved fines:	
silt and clay	1
chalk powder (added)	1.5

The experimental turbidity currents moved down a relatively steep, sand-coated slope of 25°. Observations were essentially confined to the upper ("proximal") part of the slope. During one run, the mechanism of pebble transport was studied by means of close-up high-speed photography (70 frames per second) covering a 35 cm segment of the flume at a distance of 1.3 m downslope from the valve (see position indicated in Fig. 2). This permitted analysis of both the flow's behaviour and the transportation paths of "tracer" pebbles at various levels within the flow. Flow velocity was averaged over the time interval between consecutive film frames.

The experimental turbidity currents moved with an average head velocity of the flow of approximately 100 cm/s. They all represent surge-type flows, with characteristically unsteady, non-uniform behaviour (Fig. 2). Approximate calculations indicate that the bulk Richardson number, Ri (Lüthi, 1981), falls between 0.3 and 0.4 (densimetric $Fr = 1.8$ to 1.6). This implies also some very high values for the bulk-friction factor (c_f), between 0.13 and 0.18, apparently reflecting strong friction at the flow–water interface (momentum equation indicates negligibly small friction at the bottom of the flow; see Lüthi, 1981, p. 103 and his fig. 7). The numerical result is consistent with the observed behaviour of the flow's upper surface, involving strong return flow and "back-thrusting" of turbulent suspension clouds (Fig. 2).
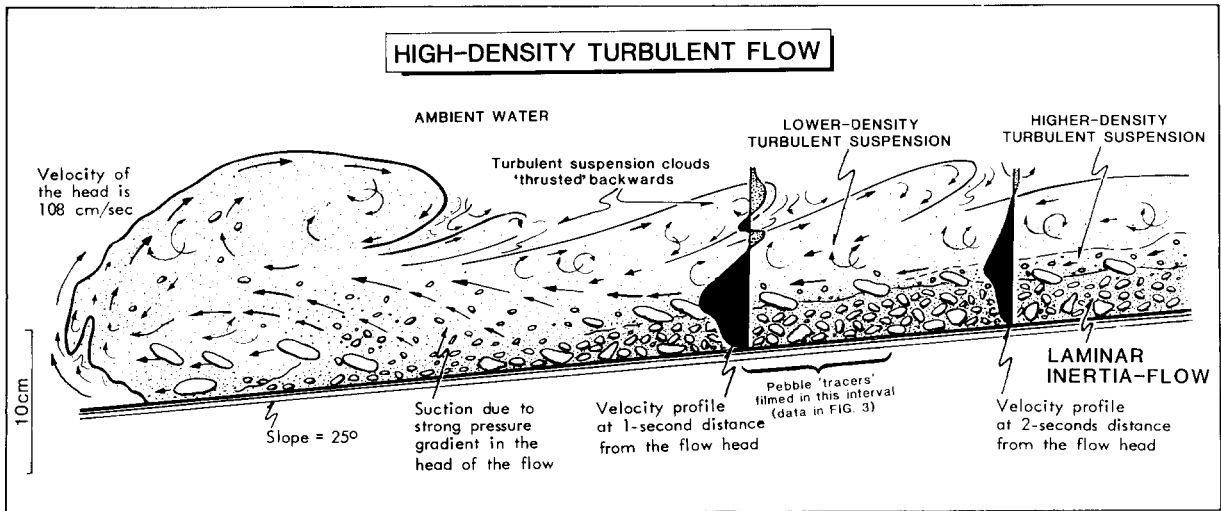
Fig. 2. A drawing from a high-speed motion picture of the experimental, high-density, surge-type turbidity current discussed in this study. Note the bipartition of the flow into a denser, non-turbulent lower layer (inertia-flow carpet) and turbulent upper layer, with the larger clasts "gliding" along the interface between these two layers.

In the experimental flows, rapid partial settling of sediment particles resulted immediately in a denser, highly concentrated layer of pebbly sand near the flow's base, herein called the (laminar) *inertia-flow layer*. In contrast to the faster moving, sandy turbulent suspension above, the basal inertia-flow layer (Fig. 2) displayed pseudolaminar behaviour, because its turbulence had been suppressed by the high particle concentration (Bagnold, 1954; Lowe, 1982). This basal, non-turbulent layer developed as a wedge thickening continuously upflow within the head region of the flow and for a short distance behind it, but having an almost constant thickness, of around 5 cm, farther upflow, within the body of the flow (Fig. 2). Notably, the basal layer in this latter region constituted more than one-third to one-half of the flow's total thickness (Fig. 2).

This highly concentrated bed-load zone, or "moving bed" in engineering terminology, is thought to represent a relatively thick "traction carpet" (*sensu* Dżułyński and Sanders, 1962, p. 88; see also Middleton, 1970) or a sort of ballistic dispersion ("rheological layer" *sensu* Moss, 1972, p. 162), whereby grains move above the bottom but are supported by mechanisms other than an upward component of turbulence. Support is probably provided by some combination of dis-

persive pressure (due to grain collisions), hindered settling (due to quasi-static grain-to-grain contacts associated with the carpet's high concentration and high apparent viscosity), and enhanced buoyant lift (due to interstitial mixture of water and finer grain fractions). High shear stresses, necessary to maintain the inertia-flow grain dispersion (*sensu* Carter, 1975), were provided by the downslope component of gravity and by the less-concentrated, faster-moving turbulent suspension passing above (Fig. 2). For the studied flows, the shear stress associated with the base of the upper, turbulent-flow layer (Fig. 2) has been roughly calculated as ranging between 490 and 860 dyn/cm^2, assuming steady-uniform flow conditions and using the formula (for similar usage see Hiscott and Middleton, 1979; Hein, 1982):

$$\tau_0 = \Delta\rho \cdot g \cdot d' \cdot \sin \phi$$

where τ_0 = shear stress at the base of the turbulent flow layer; $\Delta\rho$ = density contrast between this layer and the ambient fluid (water) above, here estimated as 0.2–0.3 g/cm^3; g = acceleration due to gravity; d = thickness of the turbulent layer, here measured as 5–7 cm (flow's body); and ϕ = slope angle (25° in the present case).

For comparison, the gravitationally derived shear stress in the laminar inertia-flow layer has

been estimated according to the formula (Middleton and Southard, 1978):

$$\tau = \rho \cdot g \cdot d' \cdot \sin \phi$$

where ρ = density of the subaqueous laminar flow (assumed as 1.4 g/cm^3); d' = height in the flow, above its base; and other symbols as above.

The calculated shear stress averages 860 dyn/cm^2, with a maximum of about 1730 dyn/cm^2 for the base of the flow (where, in turn, the transfer of the turbulent stress from above is at a minimum). This implies that the applied shear stresses within the inertia-flow layer were considerably high, obviously due to the relatively steep slope on which the experimental flows were generated.

The high experimental slope was adopted primarily to prevent, or rather delay, the "freezing" of the inertia-flow layer, by producing high shear stresses at the base of the turbulent flow and within its underlying "carpet". In relating our small-scale laboratory observations to natural flows or deposits, it is important to realise that the much larger scale of natural turbidity currents results in similar or even much larger shear stresses at the base. For example, Komar (1970) calculated that a shear stress of 1800 dyn/cm^2 was necessary for a turbidity current to move 20 cm clasts, and

that such stress could be developed by flows of density 1.10 g/cm^3 with thickness of 40 m on a slope of 1°.

The lack of recognisable grain-size segregation of sand in the basal layer of the experimental flows suggests that its high concentration resulted in high apparent viscosity (see Dangeard et al., 1965; for a semi-empirical, direct functional relationship between the two parameters, see Roscoe, 1953; and Bagnold, 1956). Differential viscosity effects probably dominated, because with increasing grain concentration, apparent viscosity increases several orders of magnitude above that of water, whereas density may increase only by a factor of about 2 (Fisher, 1971). Consequently, the development of the highly concentrated basal layer gave rise to the flow bipartition, between the more viscous, non-turbulent inertia-flow layer and the less viscous, turbulent upper layer, with a visually well-defined boundary between the two layers (Fig. 2). We infer that this *interface* played the role of an important physical discontinuity within the flow body, reflecting an abrupt vertical gradient in flow concentration (density) and apparent viscosity.

We emphasise the development of this particular interface within the flow because, in our experiments, many of the largest pebbles were ob-
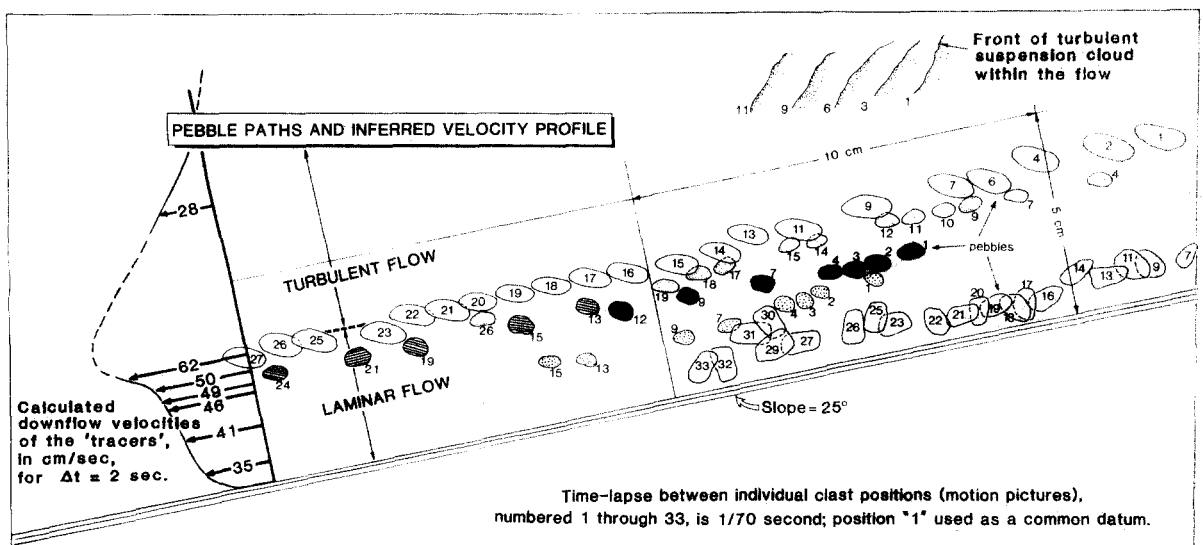

Fig. 3. Transportation paths of some large "tracer" clasts within the experimental turbidity current (example drawn from a motion picture). Note the rapidly moving largest clast "gliding" at the top of the inertia-flow layer. The variable shape/size of the individual tracers, from position to position, is due to their varying exposure at the glass wall of the flume.

served "gliding" along this boundary, directly at the top of the basal, inertia-flow layer (Figs. 2 and 3). Initially, these clasts were briefly suspended by the strong turbulence in the head region of the flow, but then were thrown far backwards into its wake. However, the inertia-flow layer developed in the body of the flow before the clasts completely settled through it. The clasts thus eventually assumed a position at the base of the turbulent flow layer, usually with partial submergence in the underlying inertia flow, whereby they immediately started to be driven by the drag force exerted from above. Apparently, the resulting downflow velocity component of the clasts was sufficiently high to prevent their settling farther through the inertia-flow layer, although similar or smaller scattered pebbles already present within this layer often showed the tendency to gradually sink through it. This mechanism of clast gliding is somewhat analogous to water-skiing, whereby the "drag force" provided by the towing motor-boat prevents the skier from sinking beneath the interface of two fluids with markedly different densities and viscosities.

The inertia-flow layer did not further increase in thickness as it began to "freeze", progressively upwards, when coming to the lower slope, where also the velocity of the overlying, turbulent flow decreased (for similar observation see Middleton, 1967). As suggested by Middleton (1967), a highly concentrated basal layer in turbidity currents behaves much like a normal (Newtonian) fluid at high rates of shear, but as the applied shear stress decreases, the apparent viscosity increases rapidly and eventually a yield-stress limit may be rapidly reached. The inertia-flow layer would thus display a quasi-plastic rheological behaviour, with essentially non-Newtonian (stress-dependent), pseudo-plastic viscometry.

An important implication is that when the overall turbulent shear stresses decrease, the gliding clast also loses its impetus, because of the reduced drag force from above and increased viscous resistance from the "moving bed" below. Hence, the clast rapidly decelerates and comes to rest at the top of the freezing inertia-flow layer, as observed in our experiments. If the upper part of the inertia-flow layer is still intensely shearing (for

example through a marked vertical gradient in apparent viscosity), the clast may partly sink into this layer, but will probably never reach its base (see previous paragraph). In either case, the clast will thus be found floating ("suspended") at a certain level *within* the resulting deposit, rather than dropped at its base.

Many, if not most, of the "gliding" clasts may actually overpass the underlying inertia flow, as observed in our laboratory flows, and be carried into the flow-head region, where they may be resuspended and thrown backwards by the turbulent lift, if the turbulence is still sufficiently strong. Thus, the clasts become candidates for another phase of gliding on the inertia-flow layer behind the head of the flow. Alternatively, the large clasts, once in the head region, may remain at the base of the flow and be transported farther by bed-load traction, or they may be deposited immediately, in either case assuming the more "typical" position at the base of the resultant deposit. In our experiments, this first mode of frontal entrapment produced concentrations of large pebbles in the flow-head region (Fig. 2). When suddenly dropped (?*en masse*) as a "lag" and left unreworked, such a local accumulation of large clasts may be preserved in the resulting deposit (Fig. 4).

No systematic documentation was made of clast fabric in the laboratory flows or in their deposits, but some consistent trends have been inferred from the observation of randomly selected, large clasts within one of the flows. Conspicuously, the large "gliding" clasts in the sandy flow maintained an *imbricate* orientation through the entire period of their movement along the top of the inertia flow, with their a-axes dipping upflow, often at an angle in excess of $15-20°$ (relative to the bottom of the flume). A similar orientation, though with generally lower dip angles and often with some transient rotation during movement, was displayed by the pebbles transported within the inertia-flow layer (Fig. 3).

The $a(p)a(i)$ orientation of the "gliding" pebbles was not, however, detected in all cases in the resulting deposit, where the clasts observed during transport or equivalent clasts often displayed an $a(t)b(i)$ fabric. No obvious variation in the fabric was observed between the flume centre and walls,

Fig. 4. Gravelly sand deposit of an inferred high-density turbidity current, containing local concentrations of coarse gravel (example from a Miocene fan delta, Tabernas Basin, southern Spain). Note the a(p)a(i) orientation of the large clasts emplaced at the top of the gravel accumulation. Palaeoflow is from the right to left. The vertical dimension of this photograph is about 2 m.

and we thus preclude the possibility that the observed orientation of the clasts was merely due to their interference with the wall.

The a(p)a(i) fabric is characteristic of highly concentrated grain dispersions subject to intense laminar shear (Rees, 1968, 1983; Allen, 1982, pp. 212–217), such as grain-flows or traction carpets. Hence, this type of orientation displayed by the "gliding" pebbles must have been due to their transport essentially *within* the top of the underlying inertia-flow layer (i.e., with the centre of mass of the pebble submerged just below the interface). Accordingly, we further infer that the a(t)b(i) orientation probably typifies those "gliding" pebbles which were dragged by the turbulent flow *on* the top of the underlying inertia flow, with the centre of mass above the interface, or which locally assumed the "rolling" orientation only after their deposition at the top of the frozen inertia-flow layer. We find it unlikely that the a(t)b(i) fabric of these isolated pebbles is due to a collision mechanism of the type postulated by Rees (1968) and further discussed by Taira and Scholle (1979).

Because of the surge-type character of the flows and the limitations associated with small-scale

flume experiments (see discussion by Middleton, 1967), no attempt was made in the present study to reproduce the internal depositional structures that have been reported from natural high-density turbidites (Aalto, 1976; Lowe, 1982; Massari, 1984).

In this paper, only the hydrodynamic/transport phenomena are considered, so it should be valid to extrapolate qualitatively the results of these small-scale experiments to large-scale natural flows. Although the high experimental slope and the manner in which the turbidity current was produced, by sudden release of a uniform high-density suspension, were different from common natural conditions, we believe that the described mechanism of the emplacement of large "floating" clasts is also important in natural high-density turbidity currents.

Discussion and implications

Transport of floating megaclasts

Turbidity currents are known to be fully capable of transporting clasts of even cobble and boulder sizes (e.g., Komar, 1970; Winn and Dott,

1978, 1979; Hein, 1982). Transportation of large clasts is enhanced by the confinement of flows, as in canyons or fan channels, where velocities, thicknesses and carrying capacities of the flows are maximised. When accelerating, such flows can suspend even their largest clasts, albeit probably rather briefly, due to the strong upward component of turbulent shear stresses. However, the potentially high suspension competence (turbulent lift) developed at an early, accelerating stage of a turbidity current does not automatically account for the occurrence of large, outsized clasts found floating ("suspended") in the resulting deposit. Such megaclasts would normally be deposited at the base of the turbidite from the bed-load, unless the depositing flow has a distinctly *layered* character which prevents settling of the megaclasts to the base of the flow.

On the basis of our laboratory observations, we suggest that outsized clasts in high-density turbidity currents can glide and be subsequently deposited at some height *within* the flow, at the boundary between a highly concentrated, non-turbulent "carpet" (inertia-flow layer) and the overlying, faster-moving turbulent layer (Fig. 5). The megaclast, in such instances, is essentially dragged along the interface between two fluid layers (sediment–water mixtures) whose distinctly different rheological properties result from layering of the sediment concentration within the flow: the upper, lower-viscosity layer being a Newtonian fluid and the lower, high-viscosity layer representing some kind of non-Newtonian fluid, most likely a pseudoplastic or (if with yield strength) quasi-plastic substance. This mode of megaclast transport, though somewhat analogous to the phenomenon of water-skiing, shows also some similarities to bimodal, pebbly sand beds subject to unidirectional turbulent flow.

Everts (1973), for example, experimentally demonstrated that large grains can be transported over a bed of smaller grains even if the latter is
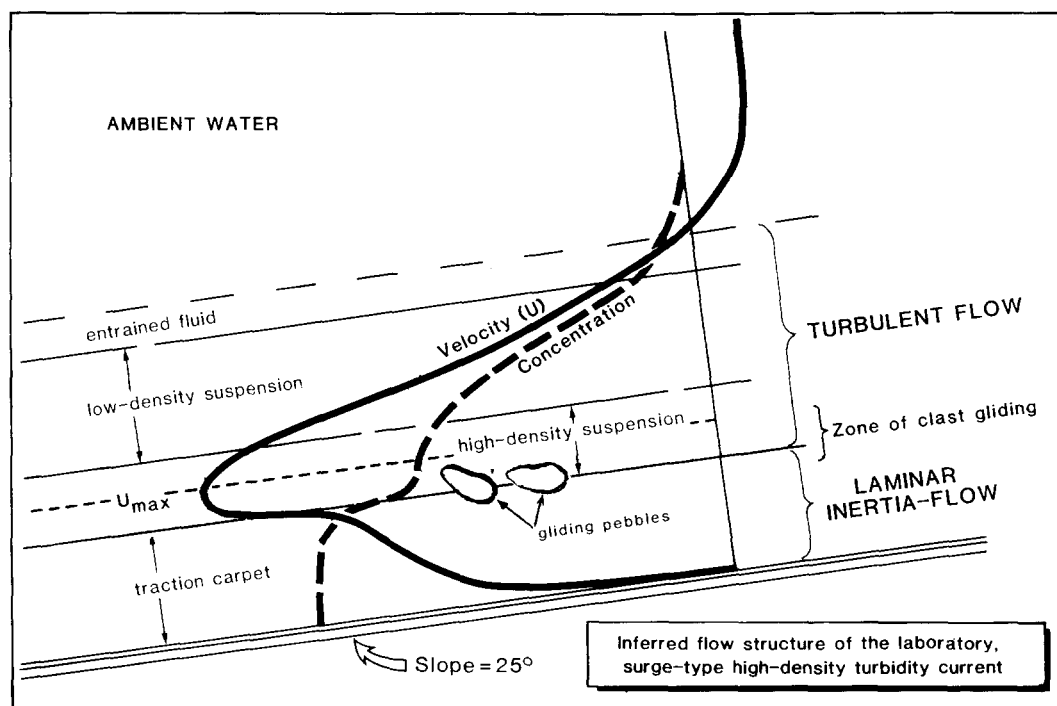

Fig. 5. Inferred two-dimensional distribution (schematic) of the flow velocity and sediment concentration in the experimental, surge-type, high-density turbidity current discussed in the present study. Note the postulated zone in which large clasts were "gliding" within the observed flow.

stable, with the applied shear stresses being less than those necessary to initiate movement of the smaller grains. Thus, in some circumstances, large clasts may be more easily transported than small grains (see also Allen, 1983). In the present case of a "moving bed", the resistance induced by a moving, viscous sandy substratum must have been considerably less than that produced by a stable bed of similar material, thus making the mobility potential of the large clasts much greater.

Isolated large clasts on a sandy substratum affect the flow by generating proximal, lateral and distal flow separation. The higher the "obstacle effect" (form resistance) of such clasts, the stronger the energy transfer into the separation cells, and the stronger the drag force component. Johansson (1976) concluded, from his experimental work, that isolated pebbles on a flat sandy bed, subject to turbulent flow, cause a strongly contracting flow above them and create conditions of high drag force and low lift force. This effect contrasts with pebble pavements having a more complete clast-size distribution, whereby the flow above the individual pebbles is less contracted and the lift, therefore, is greater than drag. In the former case, the stream lines converge above an isolated pebble (where the highest shear stress occurs) and flow energy is transferred into the separation cell upflow of the pebble where the roller flow impinges on the pebble and its sandy substratum. The outsized clast protruding above the substratum thus attains a remarkably high level of potential energy. In our experimental flows, the actual effectiveness of this downflow "pushing" force (drag) must have been increased considerably due to the relatively low bed resistance of the moving substratum (shearing inertia-flow layer).

The upflow imbrication of the outsized clasts on a sandy substratum reflects their natural tendency to produce a minimum flow disturbance (see experiments by Johansson, 1976). This pertains also to the $a(p)a(i)$ orientation of the clasts. An elongate or elliptical clast assumes such a position to minimise resistance to the surrounding or overpassing flow (Unrug, 1957; Johansson, 1965). As discussed by Johansson (1976), the flow separation results in weak, fluctuating shear stresses in the wake-vortex zone which act on the

downflow end of the pebble. Notably, this lift component would also be important, in our case, in helping the pebble to "glide", by lifting its distal end and thus making the drag force component more effective. Moreover, high flow velocity would prevent sand from being whirled into the wake separation wedge under the expanding flow and from forming features like wake ridges (Johansson, 1976), hence further minimising the substratum resistance to the gliding pebble. The reduction or cessation of these effects as the flow decelerates would then cause the pebble to lose even more rapidly its high gliding impetus, and whatever the actual combination of physical factors, the "gliding" pebble would tend to be deposited from the decelerating flow.

The chance of a gliding clast overpassing the inertia-flow layer becomes negligible once the flow has become non-uniform (i.e., when a velocity gradient has developed between the head and body of the flow). Consequently, such clasts will tend to be trapped and deposited within the body of the flow, most likely in a "floating" position, particularly once the inertia-flow layer begins to freeze from its base upwards.

The freezing of the inertia-flow layer does not preclude the possibility of a gliding clast migrating to an even *higher* level within the depositing flow. Due to continued sediment settling from the overlying turbulent flow, the inertia-flow carpet will collapse and freeze, but often a new carpet may form (Lowe, 1982). The freezing process is likely to involve only the lower part of the carpet, whereas its uppermost part, with much higher rates of shear strain (Figs. 3 and 5), then gives rise to the new, succeeding carpet. The rising rheological interface may thus force the gliding clast to a higher level within the flow, until the clast eventually loses its gliding impetus. The clast then comes to rest at a level above the base of the resulting deposit.

Clusters of floating megaclasts (Figs. 1 and 4) may be due to the tendency of a large clast to form an obstruction and trap for other clasts gliding along the same surface. Such a random clustering would then mainly depend on the actual abundance of "gliding" clasts per unit area of the rheological interface in the flow.

Other possible occurrences

The development of a rheological interface, promoting the "gliding" of megaclasts within the flow, may not be exclusively related to the formation of a basal inertia-flow layer (bed-load traction carpet). In the light of our laboratory and field evidence, it is tempting also to speculate on other possible occurrences of gliding clasts, and three such cases are postulated and discussed below.

(1) *Debris-flow with entrained turbidity current.* Turbidity currents are often entrained on top of subaqueous debris-flows, cohesive or cohesionless (*sensu* Nemec and Steel, 1984), particularly when the debris-flow actually generates the turbidity current, by reverse shear at the flow-water interface in the head region (Hampton, 1972). With the underlying debris-flow behaving in a fully viscous manner, some megaclasts could be driven along the interface between the two flows in essentially the same fashion as discussed above (cf. examples A, C and E in Fig. 1). Moreover, the top of the self-sustaining debris-flow layer would normally be partly reworked, before or after freezing, by the traction phase of the overpassing turbidity current. Some large clasts may then be derived from the top of the debris-flow body, and they may even overpass the latter and be deposited more distally, as scattered "outrunners", within the sandy turbidite deposited farther downslope (see Fig. 1E and Clifton, 1984).

(2) *Freefall clasts on viscous debris-flow.* Isolated pebbles are occasionally found floating in sandy grain-flow (cohesionless debris-flow) deposits, and their diameter may even exceed the bed thickness (Nemec et al., 1984; see also uppermost beds in Fig. 1C). It is theoretically implausible that such an outsized clast could be supported in a cohesionless, sandy/granule flow merely by the dispersive pressure which presumably maintained the surrounding, much finer-grained massflow. The magnitude of dispersive stress is directly dependent on the grain size of the shearing sediment flow (Bagnold, 1956), and hence is relatively low in a sandy grain-flow (see also discussion by Middleton and Southard, 1978). Rather, we infer that the outsized clasts were derived by freefall

from local steep slopes and probably "glided", or even partly rolled, downslope on the top surface of the viscous (fully shearing) inertia-flow (Nemec et al., 1984; Postma, 1984). Grain-flows are known to be associated with unstable, relatively steep slopes (Middleton and Southard, 1978), hence are likely to be accompanied by free-falling large clasts (if available). The high freefall momentum of such clasts would provide the actual impetus to glide downslope. When losing its momentum, the clast assumes the velocity of the inertia-flow below and begins to sink downwards. The clast may settle all the way to the base of the flow, where it will be deposited, or it may settle part way and be incorporated into the freezing flow (see Fig. 1C).

However, in natural examples of either of the two cases considered above, the outsized clasts emplaced by the "gliding" mechanism may be difficult to discriminate from those originally carried in the rigid plug of the debris-flow or subsequently settled from the plug, once it was subjected to shear. Although an association with a rigid plug is commonly invoked by field geologists to explain outsized clasts and is physically feasible, we emphasise the necessity always to evaluate the interpretation in the light of other sedimentary evidence regarding the actual behaviour of the depositing debris-flow. For example, if grain fabric, well developed grading or other features (see Enos, 1977) in the deposit suggest that the flow was fully shearing, it is unlikely that a floating megaclast represents an original component of a presumed rigid plug. In other words, it may be inappropriate to infer a plug-type flow merely from the presence of an isolated, floating megaclast.

(3) *Bipartite turbulent suspension.* A fully turbulent flow also may create a rheological interface along which floating outsized clasts probably can be deposited. Even in a fully turbulent sediment-water mixture, the suspended load may develop marked vertical heterogeneity. Increasing flow unsteadiness, due to deceleration, causes the suspended load to become progressively concentrated towards the bed. Such partitioning is particularly marked in the coarser fractions (Spells, 1955; Smith, 1958; Newitt et al., 1962). The resulting vertical gradient in apparent viscosity in such

a "layered" suspension may be sufficiently abrupt to create a rheological interface, and the latter may rise through the flow. The underlying, slower, higher-viscosity suspension is likely to display a pseudoplastic, rather than Newtonian behaviour (if the viscosity increases with decreasing shear stress), and will eventually become pseudolaminar once its turbulence has been suppressed. Its freezing, however, may be immediately followed by suspension-phase deposition, with no intervening tractional phase. Such a basal inertia-flow layer does *not* directly underlie any true tractional phase of the flow, although this phase may eventually appear at a later stage of the deposition. In such a case, the term "traction carpet" (*sensu stricto*) may be inappropriate, as noted by Massari (1984).

In fact, many turbidites appear to have recorded the development of an inertia-flow phase directly from the suspension phase, with no evidence of an intervening tractional phase (Massari, 1984; see also examples B and H in Fig. 1). Somewhat analogous behaviour, with deposition from a more or less distinctly bipartite suspension displaying both inertia-flow and limited tractional features, probably characterises some rapid, turbulent "hyperconcentrated" subaerial flows (see Fig. 1F and other examples described by Smith, 1986; also Nemec and Muszyński, 1982). Interestingly, both the subaqueous and the subaerial deposits often contain floating outsized clasts (Fig. 1B, F and H), scattered along what we infer to have been a rheological interface within the depositing flow. The outsized clasts may occur at more than one horizon, suggesting a progressively rising interface. Such megaclasts, if truly "gliding" also in these instances, would be driven solely by the downflow velocity component of the turbulent suspension phase, with no distinct traction phase involved.

The laminar shear in the lower, progressively thickening layer of the flow may be sufficient to selectively rearrange small clasts, but probably not to affect the largest ones (see fabric pattern in Fig. 1F). Horizontal pseudolamination may also be formed due to the shear (Stauffer, 1967; Carter, 1975). Dispersive stresses would be low and no inverse grading would be developed, except perhaps occasionally near the base, at the earliest

stage. The actual behaviour of this type of inertia-flow layer (or "moving bed") would then be different from the "classical", grainflow-type carpet incorporated in Lowe's (1982) model, and the inertia-flow itself might range from cohesionless to cohesive ("slurry-flow") varieties (see also discussion by Middleton, 1970; and Carter, 1975, p. 168).

We emphasise this mode of deposition from a turbulent sediment–water mixture (see also Lowe, 1982, p. 294) because we believe it is particularly important when interpreting the emplacement of large floating *intraclasts* (Fig. 1G–J).

The presence of floating intraclasts indicates that the flow was erosive at some stage, deriving the clasts through bed erosion and suspending them. The semi-consolidated debris, typically mud clasts, when suddenly introduced into a turbulent flow, will normally become subject to disintegration in the "abrasion-mill" conditions of the flow interior. However, if an initial damping of turbulence occurs at the same stage as erosion, perhaps because of the increased energy expenditure at the base of the eroding flow and possibly due to the introduction of cohesive fines, the suspended intraclasts may survive intact. In other words, a *rapid* transformation in the flow behaviour is probably necessary for the "soft" intraclasts to be preserved. The settling potential of an intraclast buoyed in a highly concentrated, highly viscous, non-turbulent flow layer will be extremely low.

Conclusions

The mechanism for the transport and deposition of large, isolated clasts "floating" in turbidites is a problem that has been difficult to resolve. Observations from experimental, high-density turbidity currents indicate that isolated outsized clasts can be transported at some height within the flow and subsequently be deposited in such a position. The clast is initially suspended by strong turbulent lift and then "glides" along the boundary between a highly concentrated, non-turbulent carpet (inertia-flow layer) and the overlying, faster-moving turbulent-flow layer (Fig. 5).

As the inertia-flow layer freezes upwards and a new one forms, or as the layer simply thickens, the gliding clast probably can be displaced to a progressively higher level within the flow. With increasing flow unsteadiness, the clast eventually becomes trapped and deposited in a "floating" position, once the inertia-flow phase entirely freezes and/or the turbulent shear stresses from above decrease critically. The arrested clast may display an $a(p)a(i)$ or $a(t)b(i)$ orientation, depending on whether it has adjusted to the inertia-flow regime below or the tractional regime above.

The surface along which the clasts glide reflects the sediment concentration layering within the flow and can probably be considered as a rheological interface, as it separates sediment-water mixtures with quite different rheological properties, particularly viscometries. The large clasts tend to be driven along this interface, rather than settling through it to the base of the flow, because their higher form-resistance provides them with a downflow drag-force component that is sufficiently strong to overcome their actual settling potential in such a "layered" flow. Given our limited experimental data, it is unclear as to how sharp the rheological discontinuity within the flow needs to be in terms of the actual density/viscosity gradients in order for this mode of clast transportation and deposition to occur. However, it is probably valid to extrapolate qualitatively our observations from the small-scale experiments to large-scale natural turbidity currents and their deposits (see examples in Fig. 1).

It is also postulated that a similar mechanism for the emplacement of outsized clasts probably operates in other types of sediment flows, subaqueous and subaerial, wherever a distinct rheological interface develops within the flow. This may be particularly relevant to some viscous (fully shearing) subaqueous debris-flows, cohesive or cohesionless, which move with an entrained turbidity current on top.

Finally, we emphasise that the implications of this study pertain to the megaclast-bearing beds deposited from *single* bipartite flows, or from two intimately associated, superimposed *concurrent* flows (as in the example mentioned above). In the field, therefore, the depositional units need to be carefully examined to avoid confusion with amalgamated or surging-flow beds, whose obliterated interfaces may also contain outsized clasts, emplaced by a more "conventional" mechanism (such as bed-load traction).

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