Understanding the high mobility of subaqueous debris flows

Fabio V. De Blasio, Anders Elverhøi, Lars E. Engvik, Dieter Issler, Peter Gauer & Carl Harbitz

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Submarine mass wasting in the form of glacial mudflows, river-laden debris flows, rock avalanches, sandy debris flows, outrunner blocks, or turbidity currents, reveal an extraordinary mobility, demonstrated by the very long runout distance between the source area and the final deposit, even on very gentle gradients. Laboratory experiments reveal that the dynamical behaviour of artificial debris flows depends dramatically on the clay-sand ratio in the experimental slurry. Artificial debris flows with high clay content, which are possibly a realistic replica of mudflows in glacially-influenced areas, tend to form a thin water layer underneath the head which acts as a natural lubricant. In contrast, lubrication cannot be easily invoked for sand-rich gravity flows. Experiments show that sandy debris flows lack cohesion, and that sand settles quickly during the rapid disaggregating phase. In the present work we review the field data, experimental results gained with debris flows of various compositions, and the status of theoretical studies and numerical simulations of submarine debris flows. When dealing with debris flows that remain compact, such as clay-rich debris flows and outrunner blocks, both experiments and simulations indicate the importance of water lubrication for mobility. On the other hand, sandy debris flows are far more complicated owing to the increased importance of water penetration, disintegration, and turbulence, and these difficulties are reflected in greater intricacy of experiments and computer simulations. Thus, the problem of whether sandy debris flows may be highly mobile in the natural setting still remains elusive.

Fabio V. De Blasio, Anders Elverhøi, Dieter Issler, Department of Geosciences, University of Oslo, P.O. Box 1047 Blindern, NO-0316 Oslo, Norway, and International Centre for Geohazards, c/o Norwegian Geotechnical Institute, P.O. Box 3930 Ullevål Stadion, NO-0806 Oslo, Norway; Lars Engvik, Faculty of Technology, Sør Trøndelag University College, NO-7005 Trondheim, Norway; Peter Gauer and Carl Harbitz, Norwegian Geotechnical Institute, P.O. Box 3930 Ullevål Stadion, NO- 0806 Oslo, Norway, and International Centre for Geohazards, c/o Norwegian Geotechnical Institute, P.O. Box 3930 Ullevål Stadion, NO-0806 Oslo, Norway.

Introduction

Deep-sea deposits from submarine landslides, debris flows, and turbidity currents have been recovered hundreds of kilometres from inland, even along very gentle seabed slopes (Bugge et al. 1988). This high mobility may be partly explained considering the large sediment volumes associated with submarine mass-wasting. An example is the Storegga landslide on the Norwegian margin, which occurred about 9,000 years ago and ran on an average slope angle of only one degree. With its 3000 km^3 of volume or more, this landslide dwarfs the largest subaerial landslides. Because large landslides usually reach longer horizontal runout than the small ones (the "volume effect"), it seems natural that subaqueous landslides should be more mobile. However, subaqueous landslides appear to be more mobile even when compared to subaerial landslides of the same volume. A better parameter quantifying the flow capacity of a landslide is the runout ratio, defined as the height of fall in the gravity field divided by the runout. Each landslide can so be identified with a point on a plane where the volume is reported in the abscissa scale and the runout ratio in the ordinate.

Figure 1 illustrates the results for a series of subaerial and subaqueous gravity flows: subaerial and subaqueous rock avalanches, subaqueous debris flows, and outrunner blocks. Similar data plots have been previously explored by Locat & Lee (2002) and Elverhøi et al. (2002), see also Edgers & Karlsrud (1982). Subaqueous debris flows fall below subaerial events, showing that the former are more mobile. On the other hand, subaqueous and subaerial rock avalanches have comparable mobility. Outrunner blocks, appearing at the bottom left corner in Figure 1, are small portions of debris detached from the front of slowing-down submarine landslides. With a runout ratio of the order 0.005 or less, outrunner blocks are amongst the most mobile gravity flows. A vertical drop of only five meters can boost the block to horizontal travel of one kilometre!
Evidently, the fact that subaqueous landslides tend to be large cannot be the sole explanation for their mobility. The comparison with subaerial landslides becomes

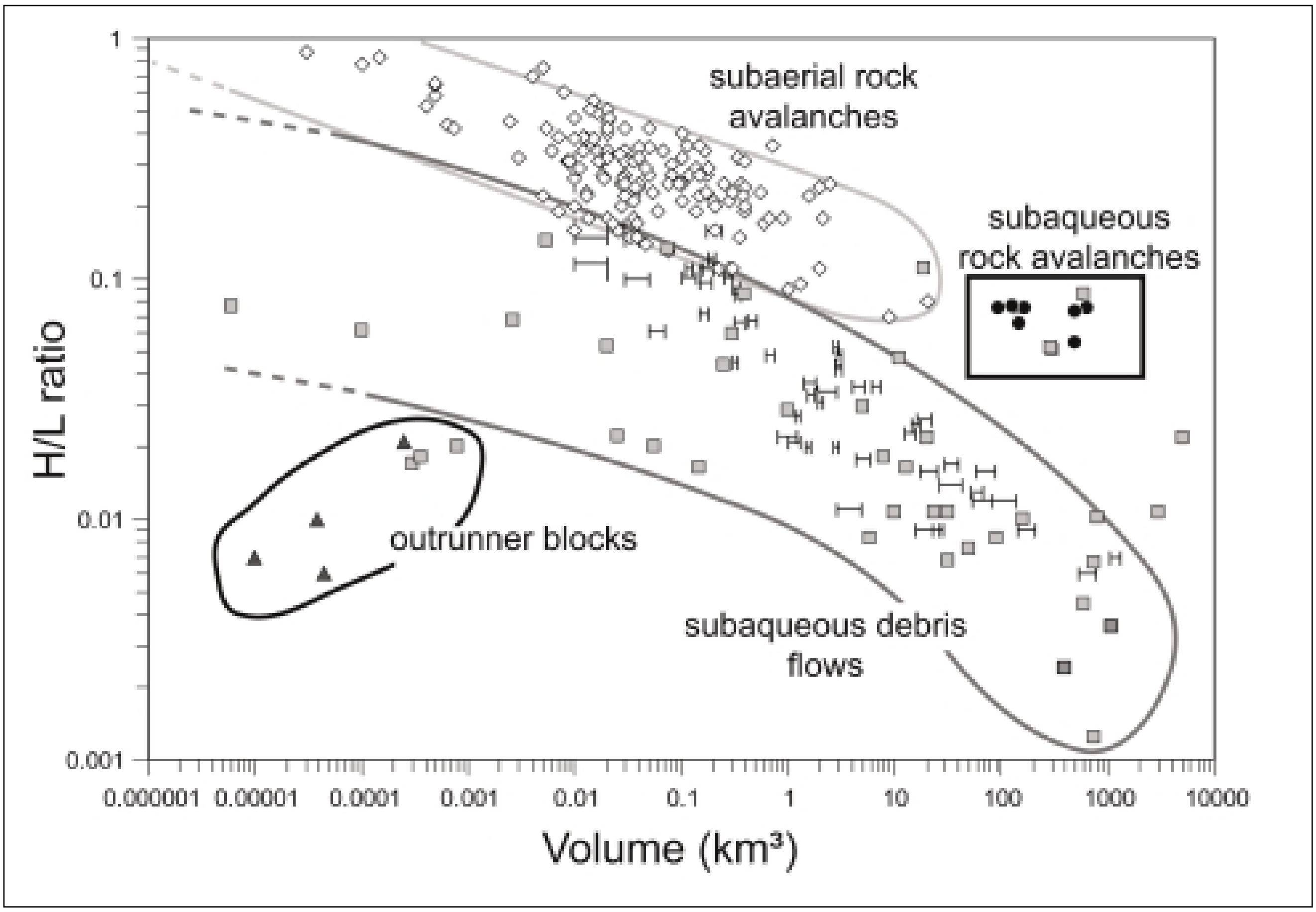

Fig 1: Diagram showing the distribution of subaqueous and subaerial landslide deposits, where the abscissa reports the volume of the landslide, and the ordinate scale is the ratio H/L between the height fall of the landslide, H, and the horizontal runout, L. Data for subaqueous debris flows (grey squares) are mainly from Edgers and Karlsrud (1982), Elverhøi et al. (2002) and Masson et al. (1988). Outrunner block data (black triangles) are from Nissen et al. (1999), Prior et al. (1982a; 1982b), Kuijpers et al. (2001), and Longva et al. (1996). Data for subaerial rock avalanches (empty squares) derive from various sources. Subaqueous rock avalanches (black dots) are taken from Masson et al. (1988). The segments (with length indicating the uncertainty in the volume) are the debris flow lobes of the Storegga slide (Haflidason et al. 2005). For very small volumes (0.01 to 0.2 km^3) the Storegga data are in reality more numerous than suggested by figure 1, as greater uncertainty make many data fall on the same segment (Haflidason et al. 2005). We note in passing that the data from the Storegga landslide area indicate a distribution $f \propto 1/V$ of landslide frequency distribution f as a function of the volume (Issler et al. 2005).

even more striking considering that the effective gravity is diminished in water due to Archimedean buoyancy, and that the drag resistance in water is about one thousand times larger than in air. This indicates that there must be a further explanation for the high mobility of subaqueous landslides, and in the present paper we argue that this explanation is probably complex. Before we tackle this problem in more detail, we notice that the ocean bottom is more regular than on land, and is usually devoid of steep valleys and slope breaks. Such topographic effect may certainly contribute to subaqueous landslide mobility. Still, the difference in mobility is enormous as can be evinced from the figure, and suggests that the explanation must rely on the interaction of the landslide with water.

In the present work we review a number of field data which show the high mobility of subaqueous landslides, and relate it to the experimental activity we conducted at the St. Anthony Falls Laboratories in Minneapolis, and to numerical modelling. Although most results (but not all) have been published elsewhere, we believe it will be interesting to tie together the information from all these different sources. Data, experiments, and theory seen in conjunction indicate the problems to tackle in future investigations.

Field data

In this section we gather information on the long runout of submarine gravity mass flows and make a comparison to similar subaerial events.

Glacial clay-rich mudflows

Mud-rich debris flows (up to 70 % clay and silt) are a rather common occurrence along glacial margins, where they are associated with glacial deposits due to former ice streams. Examples come from East Canada, East Greenland, the Norwegian and the Svalbard-Barents Sea margins (Aksu & Hiscott 1989, 1992; King et al. 1998; Dowdeswell et al. 1997; Solheim et al. 1998; Vorren et al. 1998; Elverhøi et al. 2002; Taylor et al. 2002). Well-studied cases include the debris flows along the Svalbard-Barents Sea margins (Bear Island fan). Volumes of each individual debris flow range from less than 1 to about 50 km^3 of sediments (Vorren et al. 1998). The depth difference is at most 1,500 m, and with a maximum runout of 150 km or more, this implies a runout ratio smaller than 0.01. The mobility of the Bear Island debris flow is even more remarkable considering that the mudflow did not probably dilute in a suspension but rather remained more or less intact, as evident from the well-preserved lobes (Vogt et al. 1993).

One of the subaerial mudflows with the longest runout is the Nevado del Ruiz lahar, with a runout ratio of about 0.05 and an estimated volume of one hundredth of a cubic kilometre. The high mobility, which was responsible for many casualties and serious destruction, was largely due to high water content in the pyroclastic material. Despite the factors favouring the flow of the Nevado del Ruiz lahar, the mobility of the subaqueous lobes in the Bear Island fan has actually been far greater.

Thick deposits of glacial clays may also achieve high shear strength (hundreds of kPa or more) due to compaction by the overlying layers. Measured shear strengths for the slide material in the Storegga area should be high enough to virtually hamper landslide flow. Nonetheless, the Storegga landslide lobes flowed smoothly as testified by low runout ratios, equal to 0.007 for the lobe which failed first (Haflidason et al. 2005). Probably a combination of landslide disintegration and water penetration into the comminuted material played a key role in lessening the material strength. This is a poorly understood process that we termed shear wetting (De Blasio et al. 2005). In figure 1, subaqueous debris flows (not necessarily from high-latitude glacial clays) are indicated with grey squares, whereas the Storegga lobes are shown with horizontal segment lines, with the length indicating the uncertainty in the volume.

Sand-rich debris flows or turbidite deposits

Oil and gas fields often occur in reservoir rocks made up of extended sand bodies (Stow & Mayall 2000). In the Norwegian Sea, the sea-way between mainland Norway and East Greenland developed into the deep-marine Vøring and Møre basins (Brekke et al. 2001). Sand was transported from Scandinavia and Greenland during the Cretaceous and Paleogene, and deposited in deep-sea fans. The long distance between the source and the final deposition site is bewildering if one considers the extremely flat sea bottoms for the marine basins. Most of the deep water sand deposits have been interpreted as turbidites, i.e., deposits deriving from low-density turbidity currents. These deposits are often associated with depositional elements such as channels, levees, or graded sequences (Posamentier 2003). Alternatively, turbidity currents and deep-water sand depo-

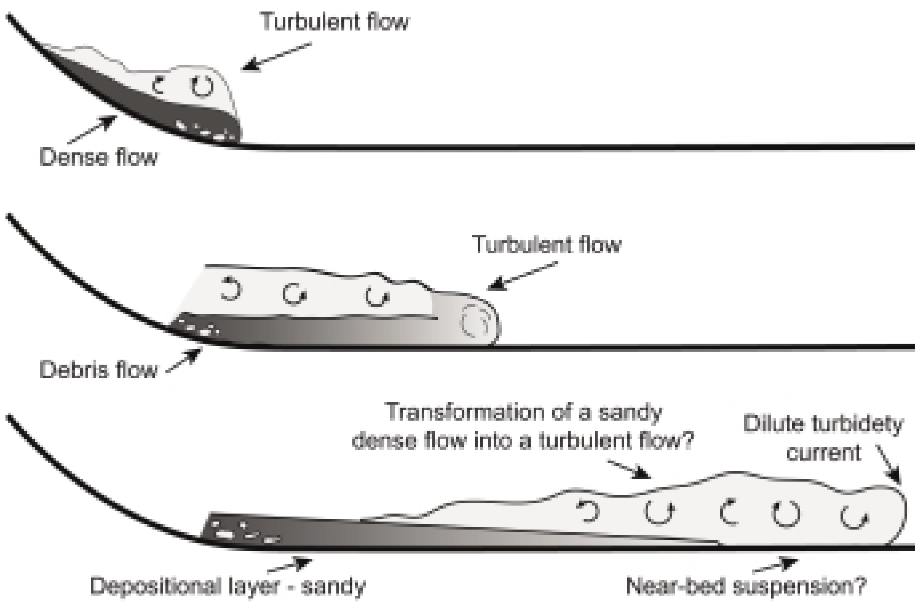

Fig 2: Diagram illustrating how a debris flow can generate a dilute turbidity current at its top and a denser basal flow during its flow down a submarine slope. Modified from Tinterri et al. (2003).

sits may form either from the disintegration of submarine landslides or from the mobilization of large mud deposits (e.g. Middelton & Hampton 1973; Hampton et al. 1996; Shanmugam 1996). According to this concept, a slide flowing downslope deposits initially coarse-grained materials and then gradually finer and finer particles (e.g Tinterri et al. 2003). Fines become suspended in a dilute low-density turbidity current, while the coarser sandy material settles to enrich a mobile basal, high-density layer (figure 2).

Long-runout subaqueous rock avalanches

In the runout ratio-volume plot in Figure 1 the black circles represent the scarce data for volcanic rock avalanches of high volume (see also Locat & Lee 2002 and Elverhøi et al. 2002). This region, bounded by a rectangle in figure 1 ("subaqueous rock avalanches"), falls on the prolongation of the interpolated line for subaerial rock avalanches, suggesting a similar behaviour of subaerial and subaqueous rock avalanches. However, other cases of subaqueous rock avalanches of comparable volume may exhibit far higher mobility, such as the Canary rock avalanche (volume 400 km^3 and runout ratio 0.0025, Masson et al. 1996) and the Saharan debris flow (volume 1100 km^3 and runout ratio 0.0036, Gee et al. 1999). These data points fall inside the tongue populated by subaqueous debris flows, well below the rectangle. The question is how a rock avalanche formed by material of all sizes (clasts from centimetre-sized to 300 m slabs have been found in the deposits of the Canary debris flow) can acquire a mobility comparable to subaqueous mudflows, which are more uniform in size and contain large amounts of clay. The question may be relevant also for the interpretation of sand-rich debris flows, as discussed in a later section.

Outrunner blocks

Outrunner blocks are intact blocks occasionally observed at the front of submarine landslides. Outrunner blocks probably originated during a rapid deceleration of the landslide front, which forces intact pieces to persist in their inertial motion. The best documented accounts come from Nigeria (Nissen et al. 1999), Kitimat in Canada (Prior et al. 1982a, 1982b, 1987; Johns et al. 1986), the Faeroe islands (Kuijpers et al. 2002; Nielsen et al. 2004) and Finneidfjord in Norway (Longva et al. 1996; Ilstad et al. 2004). Their inordinate mobility is testified for example by the 25 km runout ahead of the landslide front in the Faeroe basin (less than one degree average slope) or by the 1.2 km runout in Finneidfjord (less than half a degree slope). Deep glide tracks were generated during the passage of the largest blocks (in the Faeroe islands basin, Kitimat, and Nigeria) whereas glide tracks are very tiny along the trajectory of the small blocks of Finneidfjord (Ilstad et al. 2004a). Outrunner blocks are small (10^5–10^6 m^3 at most) and hence they fall on the bottom-left area of the plane shown in figure 1.

From the field to the laboratory

Most of present knowledge on submarine landsliding relies upon the examination of the final deposits, whereas little information can be usually obtained on the dynamics. For this reason, laboratory experiments are essential to complement the static data. Flow velocities, shear rates, or pressures can be easily measured in the laboratory, and the behaviour of the landslide material interacting with water can be directly observed. Moreover, laboratory measurements provide a platform for calibrating the numerical models prior to application to more complex field cases (Gauer at al. 2006). In the following, we show that experiments provide a first elucidation of the processes of high mobility for clay-rich debris flows and outrunner blocks. This also demonstrates the validity of the experimental methods when appropriate caveats are considered in the interpretation of the scaled results.

Experimental results and interpretation of field data based on them

Experimental set-up

Experiments were set up at the St. Anthony Falls Laboratory (University of Minnesota) in Minneapolis using slurries composed of silica sand and kaolinite clay. The composition varied between high clay content (28.7 % clay, 36.3 % sand and 35 % water by weight) and low clay content (5 % clay, 60 % sand and 35 % water by weight), (Ilstad et al. 2004b, c). In one type of experiments, the slurry is confined to a 0.2 m wide, 9.5 m long and 3 m high channel, whereas in a second type the slurry is released in a laterally unconfined pool. In the first set of experiments, a high-speed video camera recorded the velocity in the debris flow by direct tracking of marker particles floating near the transparent wall. Transducers at the bed measured the total load and the pore water pressure. The experimental set up is detailed in Ilstad et al. (2004a, b, c).

Clay-rich debris flows

Clay-rich slurries (clay content over 20 %, sand less than 36%) did not disintegrate significantly during the experiments. Eroded material from the surface of the debris flow formed a clay suspension (Hampton 1972) that can be interpreted as the experimental counterpart of a turbidity current. In such cohesive debris flows, the front is found to flow rigidly and at high speed. This

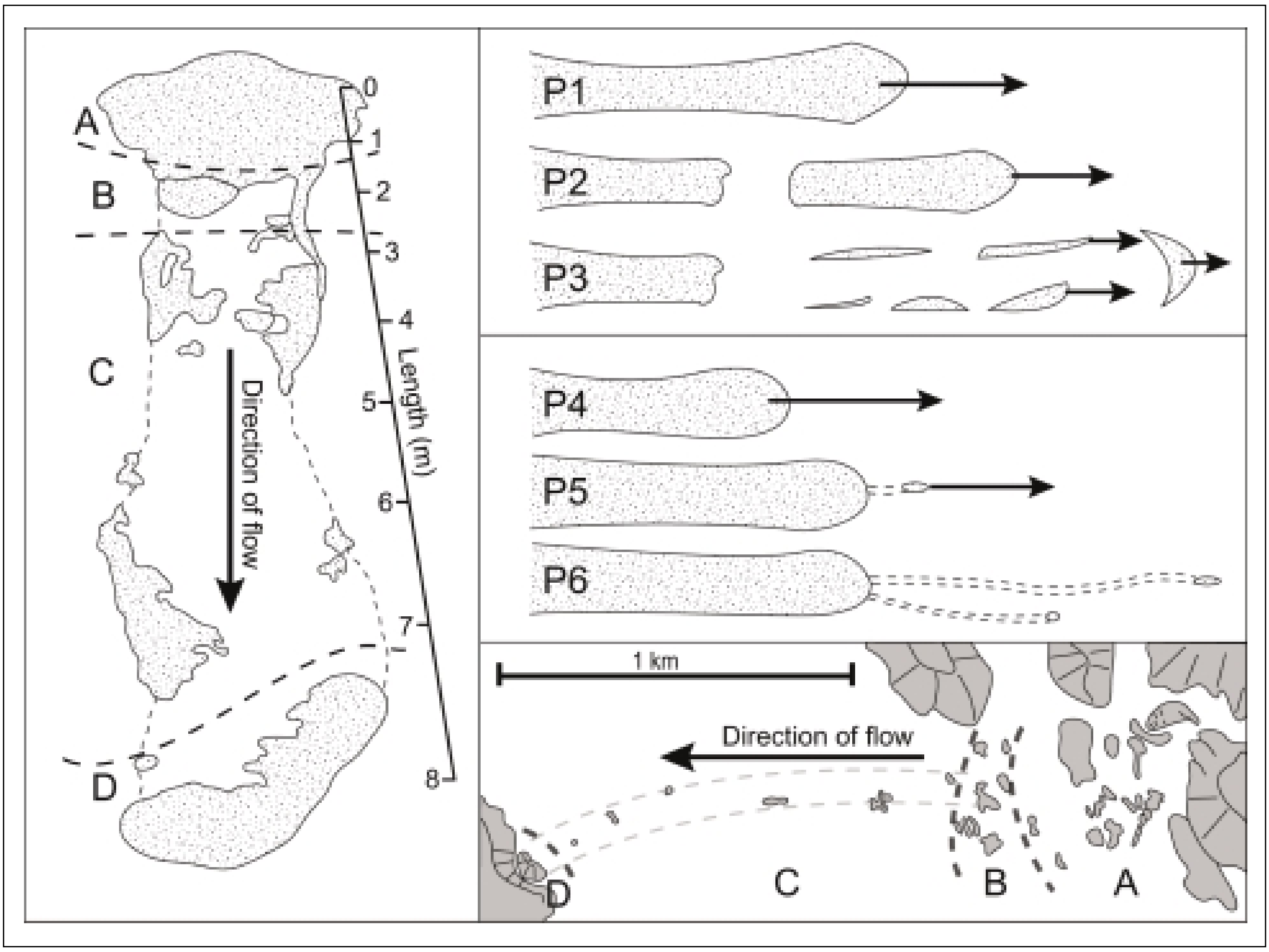

Fig 3: Scheme showing two possible scenarios for the formation of outrunner blocks. The figure to the left (modified from Ilstad et al. 2004a) shows the deposit from an unconfined experiment with cohesive slurry. A thick layer is deposited close to the gate (A), then fragmentation commences (B) forming a corridor where blocks are left behind at the sides (C), and a final block crossing the entire width of the flow path is dumped (D). The deposition mechanism is shown schematically at the top-right figure. As a consequence of front lubrication, pulling and stretching occurs at the front (P1). When the tensile strength is exceeded, the whole front portion detaches forming a large block (P2) that further fragments (P3). Erosion of the bed is negligible during the process. Bottom-right: schematic view of the block in Finneidfjord showing remarkable similarity with the experiments (modified from Ilstad et al. 2004a). From P4 to P6: model for the large blocks in Nigeria, Kitimat, and the Faeroes. P4: in contrast to the Finneidfjord case, the body of the landslide is probably not hydroplaning, and thus stretching does not occur. P5: the debris flow stops en masse. *Some blocks persist in their motion if the inertial force exceeds the cohesive force. P6. Blocks form deep glide tracks and may continue flowing for several kilometres beyond the landslide front.*

occurs because beyond a threshold velocity dictated by Froude criteria, a water layer of a few millimeters thickness intrudes at the base of the debris flow front. The water layer lubricates the front of the debris flow, boosting the front and causing a marked stretching (Mohrig et al. 1998; Ilstad et al. 2004a, c) which may lead to detachment of frontal parts of the debris flow.

Outrunner blocks have been observed in experiments with both confined and unconfined settings. The latter case is perhaps more interesting, in showing features very similar to the Finneidfjord blocks, as demonstrated in Figure 3. In the left panel of Figure 3, the final deposit of an unconfined experiment shows the presence of numerous blocks at the sides and at the end of the flow path. The succession of events reconstructed from the experiments can be summarized as follows (figure 3, upper right panel). The lubricated front accelerates whereas the body of the debris flow is left behind; this results in a necking region between them (P1). The front detaches (P2) and then fragments into a series of blocks located along the edges of the slide track (P3). As mentioned earlier, the final block distribution along the path, the existence of a single, large block at the end, and the apparent absence of friction at the floor, closely resemble the field data from Finneidfjord, shown at the bottom right of Figure 3. We can conclude that despite the difference in size, experiments provide a good small-scale model of the Finneidfjord outrunner blocks. However, differences arise when comparing the experimental data with the larger blocks of Kitimat, Nigeria, or the Faeroes, where blocks typically represent a smaller fraction of the debris flow. Seabed imagery shows that in this case only one (or few) single blocks are emitted from the front of the debris flow, with no evidence for lubrication of the whole front (see sequence P4-P6 in figure 3). It can be shown that if the material of the debris flow is homogeneous, a block of assumed semispherical shape may detach from the front of a slowing debris flow if its

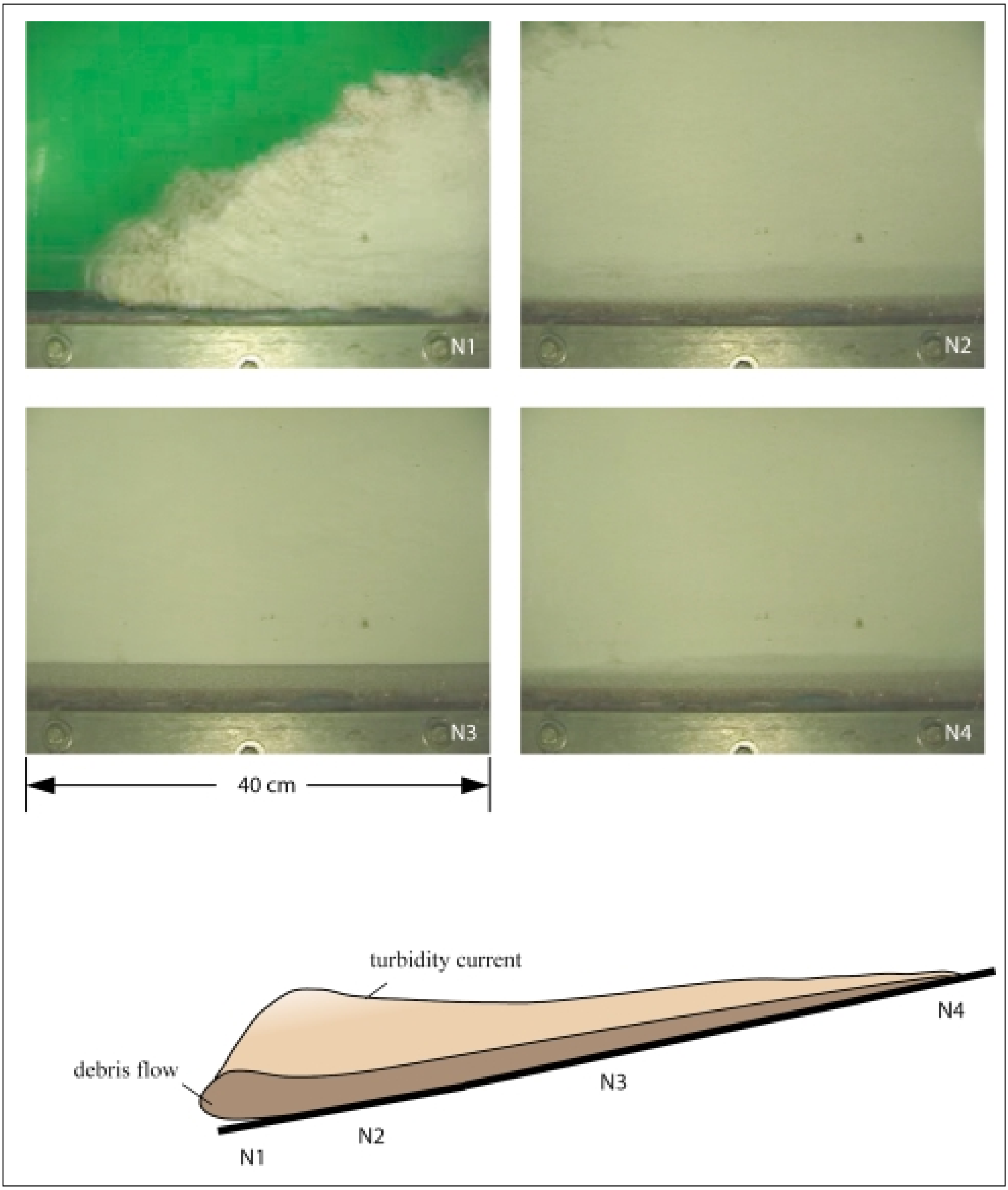

Fig 4: Experiments with a clay percentage of 10%, considered to be a "sand-rich" debris flow. N1: front of debris flow. N2: a sand layer begins to be deposited. N3: a sand wave crossing the screen at high speed. N4: final sand layer.

radius is at least larger than $R>3\gamma/a\rho$ where γ is the cohesion of the material, a is the deceleration of the debris flow, and ρ is its density. Adopting a Bingham model for the material, the deceleration is $a\approx\tau_y/\rho H$ at zero sloping angle, where $\tau_y\approx\gamma$ is the yield stress, H is the debris flow thickness, and it follows that the size of the expelled block cannot be smaller than H. On the other hand, if lubrication occurs, the deceleration a may become substantially smaller, favoring the production of even larger blocks. After the detachment, it is likely that both small and large blocks flow lubricated by a thin water layer (De Blasio et al. 2006).

Clay-poor debris flows

The dynamics is completely different for clay-poor debris flows (clay content less than 10 %, sand more than 46% by weight in our experiments). After release, the front mixes efficiently with ambient water, a process leading to complete disintegration. Sand grains freed from the matrix settle to form a sand layer, whereas clay remains suspended for a long time. Figure 4 shows the results from the experiment with 10 % clay. The four pictures (N1 to N4) have been taken with a fixed camera at successive times. The approximate position of the camera within the debris flow is shown at the bottom of figure 4. The front (N1) is turbulent. Behind the frontal bulge (N2) one can observe sand depositing (the mist is due to the clay suspension).

As sand settles, water is expelled from the lower layers and flows upwards. This helps maintaining the sand in a fluidized state, a process called hindered settling. In accord with hindered settling, we observed the sand layer flowing for some time before stopping completely. Velocities have been measured with precision by direct tracking of black coal slag particles introduced in to the slurry, and for the sand layer they were of the order of 10 cm/s. The otherwise tranquil flow of the sand layer is at times suddenly broken by the arrival of sand waves, travelling faster than the layer itself. One example of a sand wave is shown in picture N3 of figure 4. We have seen that successive waves form a stacked sequence of normally graded layers with sand at the bottom and a clay veneer at the top.

The experiments with low clay content demonstrate high mobility of these flows, even if they do not hydroplane like the clay-rich flows. We found that slurries with little clay actually attained the highest velocities. However, it is difficult to assess whether the high velocity also implies a longer runout, because the debris flows are still moving fast at the edge of the nine-metres long flume. Even more thorny is the problem of extrapolating the experimental results to the field scale. This question is particularly relevant for understanding deep water sand bodies. Because sand settles rapidly in our experiments, there is not much sand transported in suspension. One possibility for sand to be transported over long distances is thus via the fluidized layer. We performed a simple calculation accounting for the downslope flow of a sand layer fluidized by hindered settling. Assuming invariance along the coordinate x parallel to the bed, from conservation of solid particles we have

$$\frac{\partial n}{\partial t} = -\frac{\partial}{\partial y}[nv] \qquad (1)$$

where v is the vertical settling velocity of sand particles, n the number of sand particles per unit volume, and y is the coordinate perpendicular to the bed. The Richardson-Zaki equation gives the settling velocity of sand particles when water expulsion is accounted for (Richardson & Zaki 1954; Rhodes 1998)

$$v = -u_\infty \, \varepsilon^a \qquad (2)$$

where ε is the void ratio ($\varepsilon\approx 1$ for very dilute sand suspensions), u_∞ is the Stokes limit velocity of an isolated particle, and a is a constant related to the viscosity of the mixture. From the relation between particle density and void ratio $\varepsilon=1-n/n_{\max}$ where $n_{\max}$ is the maximum particle density, we find

$$\frac{\partial \varepsilon}{\partial t} = \frac{\partial}{\partial y}[(1-\varepsilon)\varepsilon^\alpha]\,u_\infty \qquad (3)$$

The horizontal velocity of the sand layer can be calculated from the Navier-Stokes equation for a laminar layer

$$u(y)\approx g\sin\beta\int_0^y \frac{D-y'}{2\mu}(1-\varepsilon)\Delta\rho\, dy' \qquad (4)$$

where g is gravity acceleration, β is the slope angle, D is the thickness of the sand layer, and the $\Delta\rho$ is the density difference between sand and water. The viscosity μ is calculated from the equation (Rhodes 1998)

$$\mu=\mu_0\,\varepsilon^{2-a} \;\; (\text{if } \varepsilon>1-\varepsilon_{\max})$$
$$\infty \;\; (\text{if } \varepsilon<1-\varepsilon_{\max})$$

and we call μ_0 the reference viscosity in the absence of sand particles. Reference viscosities are usually much greater than for pure water owing to the clay suspension in the slurry.

Figure 5 shows the results of the calculations in terms of the distance reached by sand as a function of time. Although we have not carefully tracked the complete path of sand particles during the experiments, a visual estimate for the runout of sand was about one metre. This seems to be well reproduced by the calculations (lower graph in figure 5). When the model is applied at the field scale, predicted distances are of some kilometres. A more systematic study reveals high sensitivity to the slope angle, to the reference viscosity and to the thickness of the fluidized layer. Nevertheless, the results

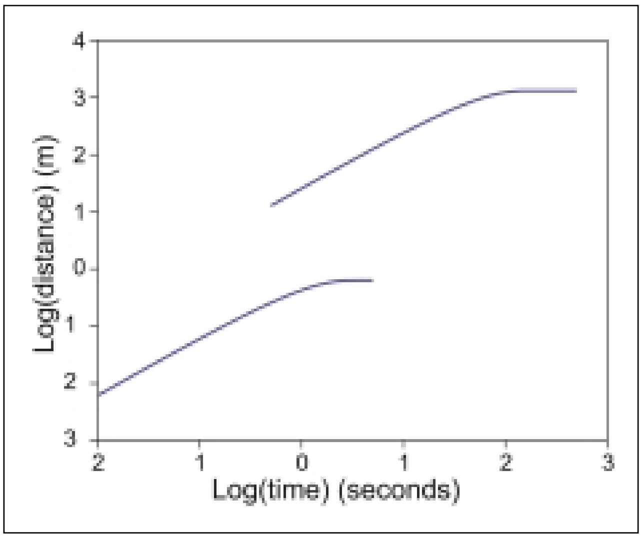

Fig 5. Maximum distance reached by sand in the fluidized layer, before settling out completely. The lower curve refers to the laboratory case, the upper one to a large scale calculation with a thickness of the sand layer of 0.5 m and slope angle 0.1 degrees.

look promising and indicate that a fluidized sand layer may boost sand mobility, even though the hundreds of kilometres typical of some deep-water sand bodies seem to be out of reach. We conclude that some other mechanism must be invoked at least for the cases of the longest runouts.

Another possibility for long runout of sandy material is indicated by the sand waves occasionally descending at high speed (picture N3 in figure 4) and overrunning the rest of the debris flow. Estimated Froude numbers of sand waves are $Fr \approx 5$ and according to Trownbridge (1987) this implies instability towards wave formation in a cohesive flow. Unstable waves and pulses have also been reported for mudflows (Coussot 1997) and in snow avalanches. As discussed earlier, successive pulses resulted in a cyclic sedimentation pattern. We speculate that some of the turbidite sequences commonly regarded as resulting from several turbidity current episodes might in reality be the product of one single debris flow carrying successive sand waves.

In all experiments, the basal dense debris separated from a dilute turbidity current at the top (Hampton 1972; Mohrig & Marr 2003). Experimental turbidity currents are mostly composed of particles elutriated from the original slurry, and the basal sandy flow is depleted in finer particles (Ilstad et al. 2004b, c). Such processes have also been investigated by Postma et al. (1988) and are included in the modelling of a turbidity current by Tinterri et al. (2003). However, it is not easy to extrapolate the behaviour of suspended particles from the laboratory to the field. Sand used in the experiments is not carried aloft owing to low fluid velocities associated with turbulent flow. On the other hand, because of larger velocity fluctuations, a natural turbidity current is more competent in sand transport.

A final observation from subaqueous rock avalanches might be pertinent. As mentioned earlier, some subaqueous rock avalanches (for example, the Canary and the Saharan debris flows) are very mobile, despite the presence of clasts of large dimensions. Because such large clasts have not been transported by turbidity currents, these cases reveal that the transport of large clasts by a dense flow is feasible in the ocean. Therefore, the transfer of sand-sized material (which is much smaller than the largest grains from the Canary landslide) by a dense debris flow must be considered likely, and perhaps common. One possibility to understand the long runout of the Canary debris flow is to assume that the pyroclastic material acted as a cohesive matrix (transport in a fluidized state must be negligible owing to the presence of fragments too large in size). The Canary debris flow might have flowed as a "clay-rich" hydroplaning flow, where the word "clay" here is related to the granulometry rather than to the composition. An outcome of this analysis is that sandy material may be able to flow over long distances if supported by a dense, cohesive clay matrix. A problem with this suggestion, however, is that sand in deep-water sandy bodies is often devoid of clay, and this calls for an explanation as to how clay was expelled before sedimentation took place.

Conclusions

The data shown in figure 1 indicate a high mobility of submarine mass wasting. Unfortunately, the sole examination of landslide deposits does not directly offer information on the dynamics. Dynamical information is indispensable to understand and predict the behaviour of such events, and can be obtained by experiments and numerical simulations. Experiments indicate that the mobility of cohesive debris flows can be explained invoking lubrication by a thin water layer (hydroplaning). A clear indication that hydroplaning does take place in natural setting is offered by outrunner blocks, and in particular by the small Finneidfjord blocks. By showing a remarkable similarity with laboratory data, the records of small outrunner blocks bridge a gap between the laboratory and very large flows, demonstrating that in some cases we can be confident in extrapolating the laboratory data by two orders of magnitude in length. However, not all deposits are clay-rich, and this questions the generality of the hydroplaning model. The experiments themselves show that sandy debris flows do not hydroplane, rather, they rapidly disintegrate, and it is not clear whether sand liberated from the matrix is capable of streaming over long distances.

We can probably state that at present, cohesive submarine debris flows are better understood (albeit still inadequately) as testified by the recent development of numerical models based on Herschel-Bulkley rheologies suitable for mudflows (Imran et al. 2001; De Blasio et al. 2004, 2005; Gauer et al., 2005). The theoretical and experimental study of sandy debris flows and their relation to turbidity currents appears as a promising and important challenge for the next future.

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