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Hydroplaning of subaqueous debris flows

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ABSTRACT

We report laboratory experiments that demonstrate that the fronts of subaqueous debris flows can hydroplane on thin layers of water. The hydroplaning dramatically reduces the bed drag, thus increasing head velocity. These high velocities promote sediment suspension and turbidity-current formation. Hydroplaning causes the fronts of debris flows to accelerate away from their bodies to the point of completely detaching from the bodies, producing surging. Instigation of hydroplaning is controlled by the balance of gravity and inertia forces at the debris front and is suitably characterized by the densimetric Froude number. The laboratory flows constrain hydroplaning to cases where the calculated densimetric Froude number is greater than 0.4. The presence of a basal lubricating layer of water underneath hydroplaning debris flows and slides offers a possible explanation for the long run-out distances of many subaqueous flows and slides on very low slopes.

INTRODUCTION

Debris flows are gravity-driven flows of mixtures of solids and fluids for which the volume concentration of each phase is of the same order of magnitude. These high-density flows occur in the relatively steep terrains of both subaerial and subaqueous environments and constitute an important mechanism for sediment transport. The subaqueous debris flows are also of interest because of their role in producing turbidity currents (Hampton, 1972; Mulder and Cochonat, 1996) and because the deposits of sand-rich flows are proposed to make up important hydrocarbon reservoirs (e.g., Shanmugam, 1996). Hazards associated with their occurrence include tsunamis and destruction of underwater infrastructure (Bea, 1971; Jiang and LeBlond, 1992; Hampton and Locat, 1996). Unfortunately, natural debris flows are difficult to study directly because they are relatively unpredictable, infrequent, and short-lived. These difficulties can be overcome by observing flows in a controlled laboratory environment. Laboratory facilities are often used to study subaerial debris flows (Johnson, 1970; Phillips and Davies, 1991; Major and Pierson, 1992; Iverson and LaHusen, 1993), but only rarely to study subaqueous flows (Hampton, 1972). With this in mind a new facility was built for the parallel study of subaerial and subaqueous debris flows.

Quantitative models of subaqueous debris flow have been adapted from those describing their better-studied subaerial cousins. Very few modifications have been made to these models to account for the difference in ambient fluid; these include (1) reducing the gravitational driving force due to buoyancy of the surrounding water (Karlsrud and Edgers, 1982; Jiang and LeBlond, 1992), and (2) incorporating the resistant shear stress applied to a debris flow along its upper interface by the surrounding water (Norem et al., 1990; Hampton and Locat, 1996). The effects of such alterations, which would act to reduce debris flow mobility, seem inconsistent with the observation that the deposits of subaqueous debris flows appear to have been more mobile than their subaerial counterparts. For example, subaerial debris flows typically cease to flow on slopes between 1° and 9° (Suwa and Okuda, 1983; Costa, 1984; Whipple and Dunne, 1992), whereas deposits of many submarine flows have been emplaced from 10 km to several hundreds of kilometers down slopes of less than 1° (Embley, 1982; Shor and Piper, 1989; Aksu and Hiscott, 1992; Masson et al., 1993; Laberg and Vorren, 1995). In addition, the contiguous deposit of a subaqueous flow may be fringed by discrete parcels of debris that appear to have separated from the toe of the main body and slid forward (Prior et al., 1984; Johns et al., 1986; Lipman et al., 1988), whereas "forerunner deposits" are not reported from subaerial flows. We have observed processes affecting debris-flow mobility that can account for both of these differences.

The laboratory experiments we report here suggest that differences in the runout and connectedness of subaerial and subaqueous debris-flow deposits can be explained by basal lubrication of the subaqueous flows. The most dramatic expression of this lubrication is the dynamic hydroplaning (Horne and Dreher, 1963; Horne and Joyner, 1965; Browne, 1975; Browne and Whicker, 1983) of most of the fronts of subaqueous debris flows. This hydroplaning occurs when a debris flow cannot displace ambient fluid from its contact area with the bed fast enough, leading to a layer of fluid separating its body from the bed. Examples of wedge-shaped layers of water penetrating underneath the fronts of subaqueous debris flows are shown in Figure 1. Experimental results presented herein document the conditions necessary for, as well as some of the consequences of, the hydroplaning of debris flows.

EXPERIMENTAL SETUP

The laboratory facility consists of a glass-walled tank 10 m long, 3 m high, and 0.6 m wide (Fig. 2). A 0.2-m-wide channel with smooth

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transparent walls is suspended within the tank. Subaqueous or subaerial flows are run down this inner channel, depending on whether the tank is filled with water. The suspended channel is articulated so that a slope break and/or gap in its bed may be introduced about 6 m down-dip from the channel head. Each channel segment can be adjusted to tilt between 0° and 20°. Opening a gap between the segments vents the debris flows downward while turbidity currents generated from the debris flows jump across the gap and continue down the channel.

Each slurry is prepared by incrementally adding approximately 5 kg batches of sediment to a tumbling mixture in a portable concrete mixer. Starting with the finest grain sizes, progressively coarser material is added until, after roughly 30 min, the mixture is complete. This debris can be introduced at the start of the channel either from a conveyor belt or from a head tank. With the head-tank configuration a hand-held mixer is used to continue stirring the slurry until release time. A typical volume of slurry released during a single run is 0.16 m^3 and completely drains from the head tank in roughly 60 s. The instantaneous volume discharge of slurry into the channel is measured by either a load cell or a sonic profiler recording either weight or volume change through time, respectively.

The motion of a debris flow is recorded by multiple video cameras, which can be arranged to capture the velocity of the debris head, as well as the bed-normal velocity profile near the channel side wall and the cross-channel velocity profile on the surface of the debris flow. Flow depth is also recorded by the cameras. Spatially averaged values of the front velocity v and flow depth are listed in Table 1. For all subaerial flows the instantaneous front velocities and depths are very close to constant along the channel, indicating approximately steady flow conditions in these cases. Nearly steady values for v, head depth h_h, and average depth h_a are also observed for subaqueous runs 5, 6, 9, 11, 14, 15, and 17 (Table 1). In the remaining subaqueous runs, 12 and 13, marginal hydroplaning results in v and h_h varying somewhat with position and time. Sediment eroded from a subaqueous debris flow and worked into an overriding turbidity current is collected using rakes of siphons.

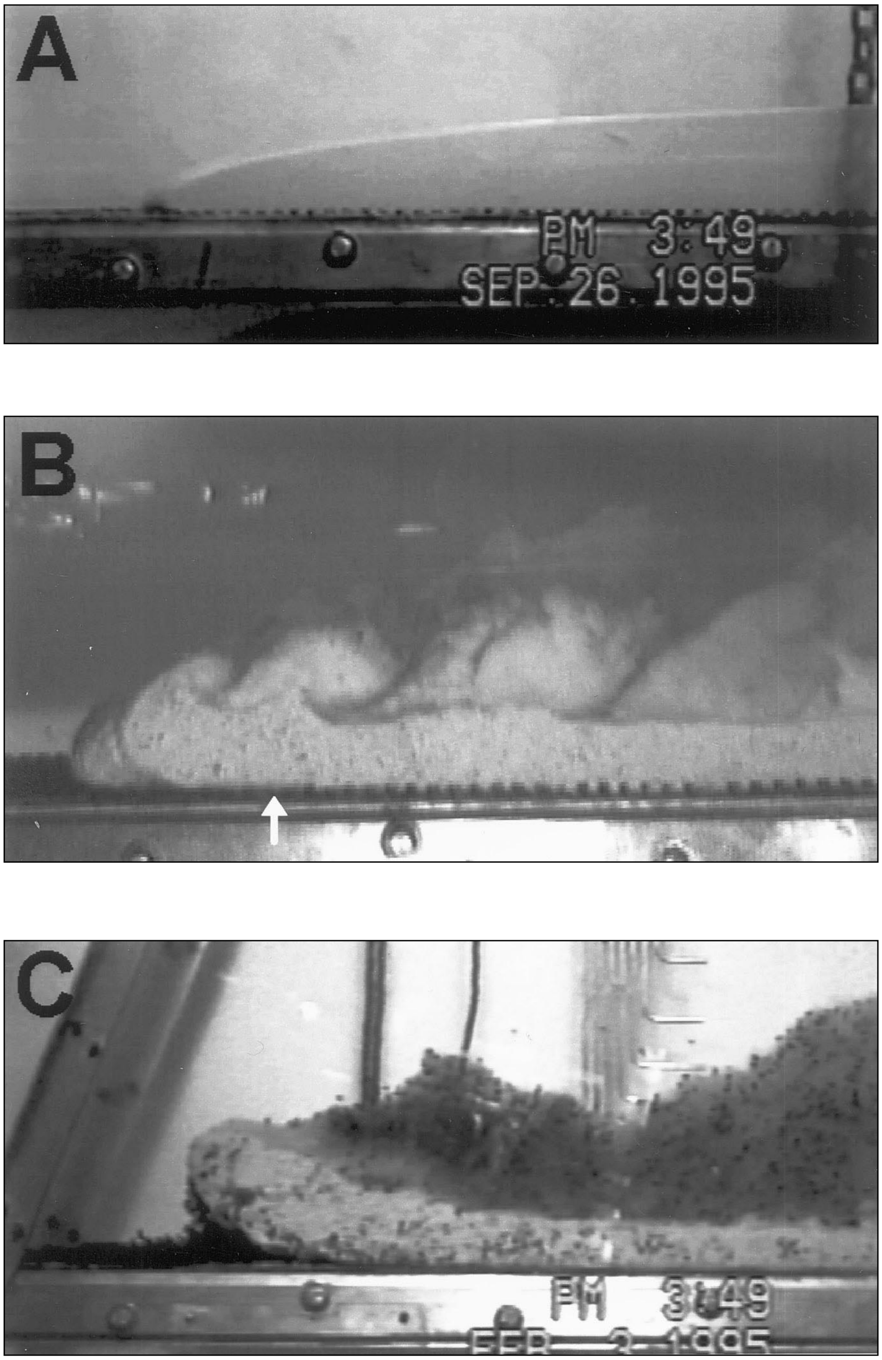

Figure 1. Fronts of laboratory debris flows. (A) Subaerial debris flow. (B) Subaqueous debris flow and overlying suspended sediment cloud. The geometry of the wedge of water underneath the debris head is easily resolvable because it prevents the light-colored debris from filling the grooves of the roughened dark-colored bed. The approximate position where the debris attaches to the bed is marked by the vertical arrow. (C) Snout of a subaqueous flow lifting off of the channel bed, a dramatic manifestation of hydroplaning. The entire debris mass in this picture is sliding downslope as a flexible block on a thin layer of water. The horizontal distances in A, B, and C are 61, 50, and 68 cm, respectively.

PROPERTIES OF THE DEBRIS

Reported compositions of debris-flow deposits from the continental slope and rise have been used to constrain the grain size, water content, and mineralogy of slurries run down the channel. All results presented here are for a single slurry type (Table 1). This slurry is the coarsest grained and has the lowest water content, 16.5 wt%, of all

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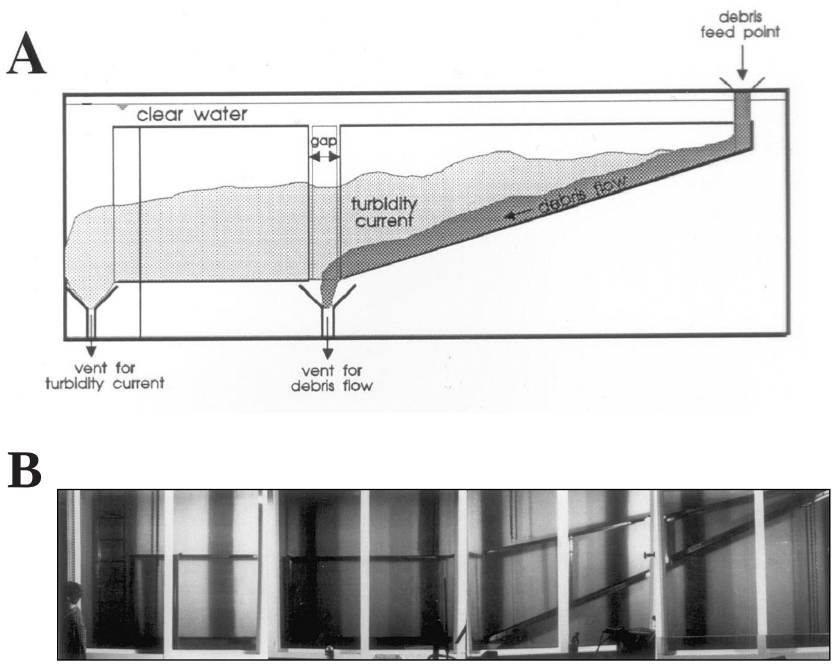

Figure 2. (A) Illustration of the experimental facility. The outer tank is 3 m high, 10 m long, and 0.6 m wide. (B) Photograph of the inner suspended channel.

TABLE 1. DATA FOR REPORTED EXPERIMENTAL DEBRIS FLOWS

Run	Θ (°)	ρ_d (kg/m^3)	Q_d (m^3/s)	h_a (m)	h_h (m)	v (m/s)	$\boldsymbol{Fr_d}$	v_* (m/s)
1(a)	20.0	2.09×10^3	1.3×10^{-3}	5.0×10^{-2}	*	0.132	4.7×10^{-3}	0.13
3(a)	20.0	2.09×10^3	1.3×10^{-3}	4.7×10^{-2}	*	0.152	5.5×10^{-3}	0.14
4(a)	16.0	2.10×10^3	1.7×10^{-3}	4.8×10^{-2}	*	0.223	7.9×10^{-3}	0.17
7(a)	15.5	2.11×10^3	12.5×10^{-3}	10.1×10^{-2}	*	0.618	1.5×10^{-2}	0.61
8(a)	15.5	2.10×10^3	12.1×10^{-3}	11.0×10^{-2}	*	0.541	1.3×10^{-2}	0.54
10(a)	4.9	2.06×10^3	3.69×10^{-3}	8.3×10^{-2}	*	0.196	5.3×10^{-3}	0.22
16(a)	6.0	2.06×10^3	4.93×10^{-3}	9.1×10^{-2}	*	0.262	6.7×10^{-3}	0.27
2(w)	20.0	2.10×10^3	1.4×10^{-3}	3.7×10^{-2}	*	0.179	0.29	0.19
5(w)	16.0	2.10×10^3	3.6×10^{-3}	4.8×10^{-2}	9.3×10^{-2}	0.861	1.2	0.37
6(w)	16.0	2.10×10^3	2.6×10^{-3}	4.1×10^{-2}	7.5×10^{-2}	0.776	1.2	0.31
9(w)	15.5	2.11×10^3	10.3×10^{-3}	5.5×10^{-2}	15.4×10^{-2}	1.24	1.6	0.92
11(w)	4.9	2.06×10^3	5.31×10^{-3}	5.3×10^{-2}	7.3×10^{-2}	0.743	1.0	0.49
12(w)	4.9	2.06×10^3	3.82×10^{-3}	5.8×10^{-2}	6.3×10^{-2}	0.589	0.76	0.32
13(w)	4.9	2.06×10^3	2.46×10^{-3}	5.2×10^{-2}	4.0×10^{-2}	0.326	0.44	0.23
14(w)	6.0	2.06×10^3	6.89×10^{-3}	5.4×10^{-2}	7.2×10^{-2}	0.776	1.0	0.63
15(w)	1.0	2.06×10^3	2×10^{-3}	6.7×10^{-2}	*	0.139	0.17	0.15
17(w)	6.0	2.05×10^3	3.56×10^{-3}	4.4×10^{-2}	7.1×10^{-2}	0.732	1.1	0.40

Notes: (a)—subaerial, (w)—subaqueous. Values for slurry density ρ_d, downslope velocity of the front of the debris flow v, average thickness of the debris flow h_a, channel inclination Θ, and input discharge of debris Q_d were collected for every run. For each run where a "squashed" front to the debris flow developed via hydrodynamically induced deformation, the thickness of the debris head h_h was also measured. Each value of h_a is the average depth for the first 6 m of debris flow in the channel and Q_d is averaged over the interval of time associated with the debris front traversing this distance. Each value for v and h_h is an average of instantaneous measures collected along the first 6 m length of the channel. The densimetric Froude number $\boldsymbol{Fr_d}$ and the characteristic average velocity v_* are calculated using reported values of the relevant parameters. For subaerial and subaqueous cases the density of the ambient fluid ρ is taken as 1.2 kg/m^3 and 1.0×10^3 kg/m^3, respectively.

we have studied to date. Resulting debris flows have a bulk density of 2.08 (±0.03) × 10^3 kg/m^3 and consist of a unimodal distribution of particle sizes whereby dry weight d_5, d_{10}, d_{16}, d_{50}, d_{84}, d_{90}, and d_{95} are 7, 12, 18, 57, 350, 500, and 700 μm, respectively. All grains are quartz; there are no clay minerals.

The hydraulic conductivity K of the sediment mixture is estimated here using an empirical method recommended by Masch and Denny (1966; Freeze and Cherry, 1979). Based on measures from samples prepared in the laboratory, K is defined as a function of both the median grain size and the particle sorting. For the grain-size distribution reported above, K is 3×10^{-5} m/s and the associated hydraulic permeability k is 3×10^{-12} m^2.

A detailed analysis of the relative roles of liquid viscosity, grain collisions, and grain friction as mechanisms of internal shear resistance and momentum transport in the debris flows (e.g., Iverson and LaHusen, 1993) is both beyond the scope of this report and not particularly relevant to the issues at hand because debris-flow rheology does not seem to play an important role in governing the conditions for hydroplaning. For the sake of comparison it is helpful to provide some quantitative description of debris-slurry rheology based on the simple assumption of a homogenous, viscoplastic (Bingham) continuum, having a finite yield strength τ_y and flowing as a linearly viscous (Newtonian) fluid once τ_y is exceeded. Assuming this rheology, estimates for τ_y and the viscosity μ_s have been obtained through complementary studies of flows in half-pipe channels, following methods suggested by Johnson (1970). The values for τ_y and μ_s are 29 Pa and 13 Pa s, respectively. The deforming portion of every debris flow (Table 1) had the structure of laminar flow.

The hydroplaning of a debris flow or slide requires the medium to be sufficiently impermeable for load-supporting pressures to build up and be sustained in the relatively thin layer of water penetrating underneath the debris. We characterize the degree of error associated with approximating the reported debris as being impermeable with two simple calculations; the first compares the time scale of pore-pressure decay within a debris flow to the characteristic run time, and the second determines the representative distance of penetration of ambient water into a typical flow. If the calculated time scale for diffusive decay of the load-supporting pressure up through a hydroplaning debris flow is very long relative to the duration of the flow, then the debris can with reasonable accuracy be estimated as effectively impermeable over this duration. This means that most of the ambient fluid in front of a debris flow must be deflected from its path in order for it to advance. Some fraction of this ambient fluid will, however, infiltrate the debris because it is not perfectly impermeable. Determining how much of the ambient fluid is incorporated into a debris flow during its runout is of great interest because dilution of the debris can change significantly the physics of the flow. In the extreme case, dilution of a subaqueous debris flow is proposed to cause its wholesale transformation to a turbidity current in which the solids are suspended by fluid turbulence (Morgenstern, 1967; Allen, 1971). The amount of infiltration is calculated assuming Darcian, porous-media flow.

The time scale for diffusive decay of any excess pressure beneath an advancing debris-flow

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front is $\delta^2\mu/kE$, where E is the stiffness of the granular mixture, k is the permeability of the granular mixture, μ is the viscosity of the pore fluid, and δ is a characteristic length that for this analysis is taken to be the average flow depth (Iverson and LaHusen, 1993; Iverson, 1997). To calculate this value for our experiments we use a representative average flow depth of 0.05 m (Table 1) and $k = 3 \times 10^{-12}$ m^2. We did not collect measurements from which E can be directly estimated, so we have selected the largest value for this parameter from the range of typical values for debris-flow mixtures of 10^3–10^5 Pa reported by Iverson (1997). The use of $E = 1 \times 10^5$ Pa is intended to yield a minimum or conservative estimate for the time scale. Before calculating this scale we need an estimate of μ. The viscosity of the pore fluid will deviate from that for clear water (1×10^{-3} Pa s) because of the presence of fine grains in the fluid. Following Iverson (1997), an empirical equation developed by Thomas (1965) is used here to estimate the effective fluid viscosity. This equation is $\mu/\mu^* = 1 + 2.5\varepsilon_f + 10.05\varepsilon_f^2 + 0.00273\exp(16.6\varepsilon_f)$, where μ^* is the viscosity of clear water and ε_f is the volume fraction of slurry occupied by fine grains. Following the rationale put forth by Iverson (1997) and Major (1996), all silt- and clay-size grains are considered here to act as part of the fluid. The complete slurry is 34% water, 33% clay and silt, and 33% sand by volume, so in a normalized mixture of water and fines, the two phases make up 51% and 49% of the volume, respectively. Taking $\varepsilon_f = 0.49$ yields a value for μ/μ^* of 14.5, which gives a value for μ of 1.45×10^{-2} Pa s. From the reported values for all of the relevant parameters the calculated time scale for diffusive decay of any excess pressure beneath an advancing debris-flow front is 121 s. This conservative value is still significantly greater than the time, about 10 s (Table 1), it takes a hydroplaning debris flow to travel the length of the flume, thus explaining why high pressures can be sustained beneath the advancing front.

The rate of water infiltration into the front of a hydroplaning debris flow can be estimated in the following way. A subaqueous debris flow advancing with speed v into quiescent water generates at its leading edge a theoretical stagnation pressure $P_f = 1/2\rho v^2$, where ρ is the density of the ambient fluid and v denotes the front velocity. This pressure provides an appropriate scale for the dynamic pressure in the fluid layer over which the debris front hydroplanes. The part of the pressure differential between the top and bottom of the head of the debris flow that deviates from hydrostatic can thus be scaled as $dP/dz \sim -(1/2)\rho(v^2/h)$, where h is a representative thickness for the debris flow. The effect of hydroplaning would then be to drive water up into the debris flow. As can be seen in Table 1, typical values for v and h for the experimental subaqueous debris flows undergoing hydroplaning are 0.75 m/s and 0.05 m, respectively. The intergranular flow that this would drive can be estimated with the use of Darcy's law, where q denotes the velocity of upward flow of water: $q = -(K/\gamma)(dP/dz) \sim (1/2)Kv^2/gh$, where $\gamma = \rho g$ and g is the acceleration of gravity. Using the above estimates for v and h and the previously quoted estimate for K, it is found that $q \sim 2 \times 10^{-5}$ m/s. The front of a typical debris flow travels down the channel over a period on the order of 10 s, implying that during the course of the event the thickness within the head that could be penetrated by ambient water from below is of the order of 0.2 mm. Because the pressure differential driving the penetrating flow is greatest in the bed-normal direction, this distance is a relative maximum estimate for that portion of the front of a debris flow affected to some undetermined degree by dilution. The very small value for this distance corroborates the treatment of these debris flows as being impervious to water.

HYDROPLANING OF DEBRIS FLOWS

The deflection of ambient fluid by a debris flow from its path changes the momentum of this ambient fluid, thereby imparting to the fluid a reaction force that in turn acts on the debris-flow surface. For subaerial flows the density of air relative to debris is very small (<1:1000), so stresses applied to the flow surface by the ambient fluid are usually negligible compared to basal friction and are typically excluded from descriptions of the governing force balance (e.g., Liu and Mei, 1989; Savage and Hutter, 1991). Under water the density of fluid relative to debris is close to 1:2, and stresses along the upper boundary are probably significant. For example, analysis of the submarine Sokkelvik slide (Norem et al., 1990; Hampton and Locat, 1996) indicates that shear resistance along its upper surface equaled about 30% of the shear resistance at its base. Here we are interested not in these shear stresses, but rather in the normal stresses acting upslope on the typically blunt-shaped fronts of debris flows. The hydrodynamic pressures developing at the fronts of experimental subaqueous flows are almost always great enough to strongly modify their shape (Figs. 1 and 3) and to at least partially support their loads.

The fronts of all the subaerial debris flows are similar (Fig. 3A). This shape is also shared by the slowest moving subaqueous flow (run 15, Table 1). The overhang seen in the leading edges of these flows is caused by frictional resistance at the bed (Middleton, 1966; Simpson, 1972). Overhanging shifts the foremost point on each flow front some finite distance above the bed and affects the movement of nearby ambient fluid. Fluid beneath the position of the foremost point cannot move up and over the debris flow, but is rather pushed along in front of or penetrates underneath the flow (Middleton, 1966; Allen, 1971). The foremost point thus becomes a stagnation point at which the dynamic pressure is estimated by $P_f = 1/2\rho v^2$, as noted previously.

The hydroplaning of a debris flow is caused by hydrodynamic pressures deforming its wetted perimeter from the stagnation point down to the bed (Fig. 3). This deformation enables a discrete wedge-shaped layer of water to penetrate underneath the debris, having the effect of peeling back the debris from the bed. The lower-viscosity fluid is easily sheared and is thus associated with a dramatic reduction in the bed-resistance force acting on that section of the debris flow detached by a layer of fluid thick enough to submerse roughness elements on the bed. Development of such a layer of fluid under the front of a debris flow is likely limited by whether the developing hydrodynamic pressure P_f is able to support the average submerged debris-load $P_d = (\rho_d - \rho)gh_a \cos\Theta$, where ρ_d is bulk density of the debris and Θ denotes the angle of bed inclination. The simple balance between P_f and P_d, cast in the form of the dimensionless Froude number, has previously been used with great success in scaling the effects of the resultant pressure force on a moving body. Examples include uplift forces on hydraulic structures and NASA's empirical hydroplaning parameter for pneumatic tires (Horne and Dreher, 1963; Horne and Joyner, 1965). The work on tires is particularly relevant to our study. Measured fluid pressures in the water separating tires from contact with the hard bed are very close to those calculated with the expression for P_f above; complete hydroplaning occurs as the hydrodynamic pressure approaches the total vertical load on the tire bottom (Horne and Joyner, 1965). Following these examples we propose to describe the necessary condition for and the extent of debris-flow hydroplaning with a densimetric Froude number, $\boldsymbol{Fr_d}$, that specifically accounts for the buoyancy of the ambient fluid:

$$\boldsymbol{Fr_d} = \frac{v}{\sqrt{\left(\frac{\rho_d}{\rho} - 1\right) g h_a \cos\Theta}} \qquad (1)$$

Values of $\boldsymbol{Fr_d}$ for all studied debris flows are listed in Table 1. For subaerial flows the parameter is between 10^{-3} and 10^{-1}, and the subaqueous cases have $\boldsymbol{Fr_d}$ values between 10^{-1} and 10. The difference of two orders of magnitude in the range of values for $\boldsymbol{Fr_d}$ comes from the large difference in density of the two ambient fluids. A basal layer of ambient fluid associated with hy-

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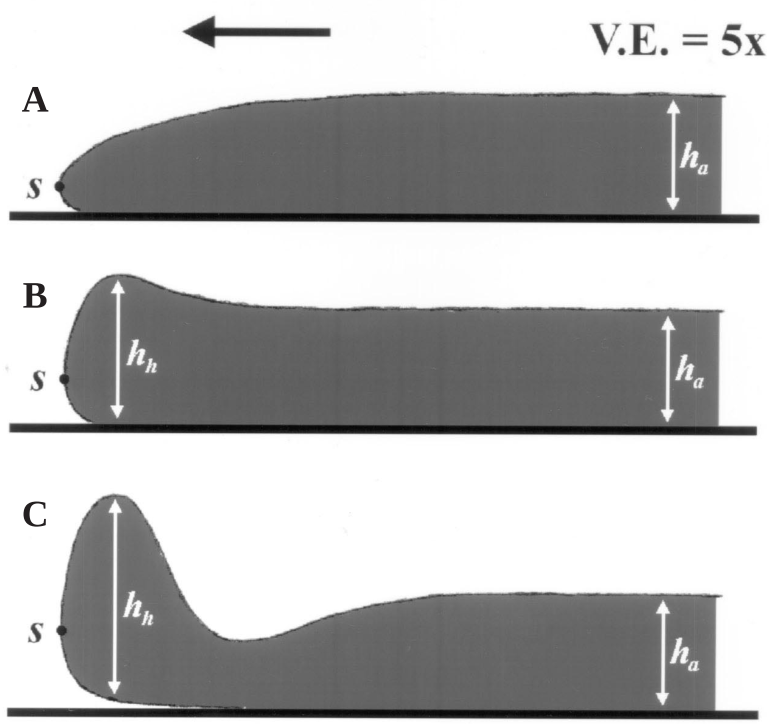

Figure 3. Schematic profiles of the fronts of observed debris flows. V.E. is vertical exaggeration, which is 5× in each drawing. The foremost point or stagnation point *s* is marked on each front. (A) Densimetric Froude number, $Fr_d < 0.3$. This shape is shared by all nonhydroplaning flows, subaerial and subaqueous. (B) $Fr_d \approx 0.35$. Shape of the debris front at the very onset of hydroplaning. Hydrodynamic deformation of the front causes development of a bulbous head. Head depth h_h is the maximum thickness of debris in the head. (C) $Fr_d > 0.4$. Shape of a hydroplaning debris front. The necking in the flow behind the head is due to flow attenuation between the lubricated front and the more bed-attached body. An estimate of the average depth of flow h_a is marked on each schematic profile.

droplaning was observed underneath the fronts of all of the subaqueous flows except runs 15 and 2 (Table 1). No basal layer of fluid is seen separating the fronts of these two subaqueous flows from the bed. Calculated values of $\boldsymbol{Fr}_d$ for the nonhydroplaning flow 2 and the hydroplaning flow 13 are 0.29 and 0.44, respectively. These two cases constrain the onset of hydroplaning in the laboratory experiments to a value for $\boldsymbol{Fr}_d$ between 0.3 and 0.4, supporting the hypothesis that the hydrodynamic pressure P_f must come close to balancing the average debris load P_d in order for hydroplaning to begin. The hydrodynamic pressures necessary to induce hydroplaning are also sufficient to significantly deform the fronts of the laboratory debris flows, leading to a change in their shape (Fig. 3). This hydrodynamically induced deformation produces distinct "squashed" heads on these flows that in the laboratory have a height, h_h, that is as much as three times as great as the average depth of flow h_a (Fig. 3C). For the hydroplaning flows the ratio of h_h to h_a increases as an approximately monotonic function of $\boldsymbol{Fr}_d$ (Table 1).

In the laboratory the maximum distance a layer of water could be seen penetrating upslope underneath a hydroplaning debris flow is about 10 times the average flow depth h_a. The distance seems to increase with $\boldsymbol{Fr}_d$, and debris underlain by this lower viscosity shear layer ceases to flow internally and moves forward as a block. For this particular slurry the most striking consequence of hydroplaning is the increase in front velocities relative to those of nonhydroplaning debris flows. In Figure 4A each hydroplaning case has a front velocity significantly greater than that value associated with a comparable subaerial flow. For a nonhydroplaning flow, one expects the front to advance no faster than the velocity at which debris is supplied to the channel, $v_* = Q_d/(Wh_a)$, where Q_d is the volume discharge of debris into the channel and W is channel width (Table 1). Comparing front velocity v to v_* therefore provides a convenient way to determine whether the front is pulling away from the body, a likely manifestation of the dramatic reduction in bed-resistance force acting on the front of a hydroplaning debris flow. Figure 4B shows that v and v_* are approximately equal for the subaerial cases and the two slowest moving (nonhydroplaning) subaqueous cases, whereas the measured values of v noticeably exceed v_* for the remaining subaqueous flows. The difference in forward velocity between the fast-moving debris riding on a discrete layer of water and the trailing portion of the flow that is more attached to the bed and moving at a slower rate leads to a stretching and attenuation of the flow directly behind the bulbous, hydroplaning, debris head (Fig. 3C). In run 9 (Table 1) the head accelerated away from the body to the point of completely detaching from the body, causing a new head to form. The detached debris mass slid to the end of the channel in front of this newly developed head to the body of debris, suggesting that under the right conditions a single debris flow might shed from its front a succession of forerunning blocks of debris.

TURBIDITY-CURRENT PRODUCTION

All subaqueous flows in the laboratory generated turbidity currents via sediment erosion from the debris flows and suspension into the water column. As reported by Hampton (1972), we find most of the sediment constituting these turbidity currents to be eroded from the fronts or heads of the debris flows. The significant sediment erosion is limited to this area where the velocity gradient, or shear, in the column of water in direct contact with the debris surface is high. Nearly neutrally buoyant beds seeded within the ambient fluid show that shear at the debris-water interface rapidly decreases behind the head of a debris flow because this water is quickly accelerated by the moving debris flow up to a velocity that is close to that of the upper surface of the debris flow. The observed lack of erosion along the upper debris-flow surface is a consequence of this lack of shear. For the experimental flows the transfer of sediment out of a debris flow and into a developing turbidity current is relatively inefficient. Even for the fast-moving debris flows traveling a distance equaling approximately 200 times their own average depth, the fraction of sediment eroded from the debris flow and incorporated in the turbidity current is ≤1% of the total volume of sediment in the system. Whereas the average volume concentration of sediment within the debris flows is 0.66, measured values for the vertically averaged, volume concentration of suspended sediment in the turbidity currents are between 1×10^{-4} and 2×10^{-3}. An important reason for this inefficient transfer is the previously discussed low hydraulic conductivity of these debris

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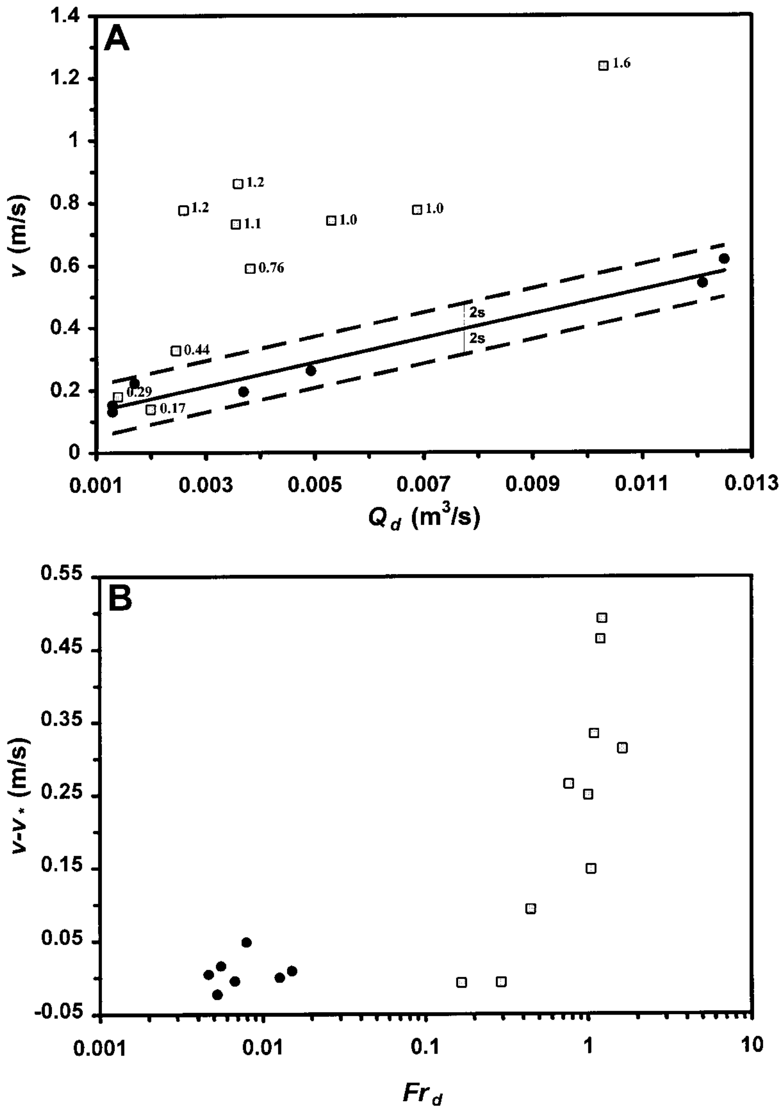

Figure 4. (A) Front velocity *v* of debris flows as a function of their input discharge of debris Q_d. Circles mark subaerial flows and squares mark subaqueous flows. The estimated value of the densimetric Froude number, Fr_d, for each subaqueous case is shown next to its marker. The best-fit line to the subaerial data is plotted, along with twice the assumed constant standard deviation of *v* given Q_d. The estimator of this standard deviation *s* is the square root of the average of the squared residuals. Approximately 95% of all possible front velocities for subaerial flows should fall within the boundaries defined by the two dashed lines. Subaqueous debris flows with front velocities plotting above the upper dashed line are therefore interpreted as moving significantly faster than comparable subaerial flows. (B) Difference between front velocity *v* and a characteristic velocity v_* plotted as a function of Fr_d. A value of $(v - v_*)$ very close to 0 implies that the debris front is fully connected to its body and that there is no hydroplaning. A value significantly greater than 0 implies that the head is accelerating away from its body and that it is hydroplaning. Circles mark subaerial flows and squares mark subaqueous flows.

flows, which suppresses the infiltration of water into debris flows and any possibility of their wholesale dilution and transformation into turbidity currents. Instead, the turbidity currents are generated by the suspension of grains individually plucked from the surfaces of debris-flow heads. These turbidity currents range in depth to as much as roughly six times the average depth of the associated debris flow. Where the debris flow deposited before reaching the end of the flume, the turbidity current peeled off and continued advancing downslope (Fig. 5). At the end of each run a layer of sediment deposited from the turbidity current was seen mantling the top of the debris flow, as well as any exposed section of the channel bed out in front of the flow. All of these turbidites were ≤1 mm in average thickness and enriched in the finer grain sizes relative to the debris that is their source.

Hydroplaning significantly enhances both the suspension of sediment and the resulting turbidity currents by increasing the velocities of the heads of debris flows. The observation that multiple heads are sometimes shed from a single debris flow by hydroplaning has further implications for turbidity-current generation. Each of these heads may be a moving point source for the production of turbidity-current sediment. Depending on the spacing between surging heads, either a single current with a pulsating sediment concentration field or multiple turbidity currents may be generated from one debris flow. In both scenarios the amount of sediment suspended would likely be greater than expected from a single debris flow. We have observed the generation of multiple turbidity-current heads from a single debris flow in the laboratory.

DISCUSSION

Our laboratory results define a minimum value for Fr_d of roughly 0.35 for the onset of hydroplaning. This relationship may be used to estimate critical values of the relevant parameters for larger natural debris flows, subaqueous or subaerial. Direct records of subaqueous gravity-driven currents are rare, but reported thicknesses, densities, and downslope velocities support the suggestion that hydroplaning is an important process in this environment (Karlsrud and Edgers, 1982; Piper et al., 1988; Piper and Savoye, 1993). For example, Table 2 lists calculated values of Fr_d for five submarine landslides analyzed by Edgers and Karlsrud (1981) that either exceed or are very close to the minimum value for hydroplaning as defined in the laboratory. In addition to these calculations there exists at least one additional line of evidence supporting hydroplaning: the reported occurrence of enigmatic isolated blocks of debris that apparently slid up to kilometers beyond the terminus of the contiguous debris deposit (Prior et al., 1984; Johns et al., 1986; Lipman et al., 1988). The separation of blocks of debris from the fronts of subaqueous flows has been observed in the laboratory by Hampton (1972) and by us. Middleton and Hampton (1976, Fig. 8) included a photograph of a detached block of debris that is mov-

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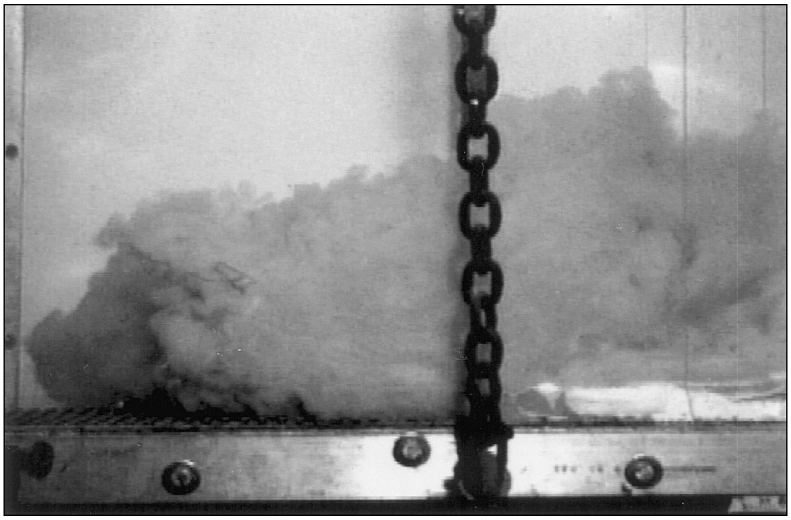

Figure 5. The head of a turbidity current advancing beyond the terminus of the deposit of the debris flow from which it was produced. The front of the debris deposit coincides with the horizontal position of the hanging chain in the photograph. Total horizontal distance in the photograph is 56 cm.

TABLE 2. DATA FOR NATURAL LANDSLIDES

Slide	Θ (°)	ρ_d (kg/m^3)	h_a (m)	v (m/s)	$\boldsymbol{Fr_d}$
Grand Banks (w)	3	1.71×10^3	50.	27.4	1.47
Messina (w)	3	1.71×10^3	20.	6.00	0.51
Orleansville (w)	15	1.71×10^3	20.	19.5	1.68
Orkdalsford (w)	5	1.71×10^3	10.	2.60	0.31
Sandnessjoen (w)	5	1.71×10^3	2.0	0.70	0.19
Blackhawk (a)	0	2.0×10^3	30.	34.	4.8×10^{-2}
Elm (a)	0	2.0×10^3	38.	29.	3.7×10^{-2}
Frank (a)	0	2.0×10^3	46.	49.	5.6×10^{-2}
Sherman (a)	0	2.0×10^3	6.0	51.	0.16

Notes: (a)—subaerial, (w)—subaqueous. The densimetric Froude number, $\boldsymbol{Fr_d}$, is calculated using the parameters above. For subaerial and subaqueous debris slides the density of the ambient fluid ρ is taken as 1.2 kg/m^3 and 1.0×10^3 kg/m^3, respectively. Values for subaqueous slides are from Edgers and Karlsrud (1981). The reported bulk density ρ_d is thought to be typical of large submarine slides, the average slide thickness h_a is estimated based upon the known dimensions of the slide, and Θ is the bed inclination in the vicinity of the initial failure. Downslope velocity v is calculated from the known distance and elapsed time between the first and second breaks of submarine cables. Values for subaerial slides are from Shreve (1966, 1968a). A bulk density ρ_d thought to be typical of the Blackhawk Slide is imposed (Shreve, 1968b). Reported downslope velocity v is estimated by conserving slide energy, the maximum reported deposit thickess is taken as the average slide thickness h_a, and the bed inclination Θ is set to zero. Values of h_a, and Θ were chosen for the subaerial slides to provide conservative estimates of $\boldsymbol{Fr_d}$.

ing out ahead of its body. We believe that this detachment of debris from the fronts of flows is probably a characteristic signature of hydroplaning, where increasing thickness of the lubricating water layer toward the front of a flow causes the head to accelerate away from its body, as discussed in the previous section.

This is not the first time that the high mobility of some debris flows and rock slides has been ascribed to their riding on distinct, easily sheared, lubricating layers. For example, Shreve (1966, 1968a) showed how characteristics of some rockslide deposits are consistent with their having moved at high speeds over relatively thin layers of trapped air, and Voight and Sousa (1994) speculated that a debris flow, upon entering a river channel, may have ridden locally on top of a shearing layer of water. The Shreve hypothesis envisions a one-time trapping of lubricating fluid beneath a slide that is different from the mechanism documented here, for which the underlying fluid remains connected to the body of ambient fluid and can be replenished from the leading edge. This notwithstanding, it is of some value to compute values of $\boldsymbol{Fr_d}$ from Shreve's estimates of v, h, and ρ_d. For the four cases cited in Table 2, minimum estimates of $\boldsymbol{Fr_d}$ are found to be 0.04, 0.05, 0.06, and 0.16. The highest of these values approaches the value here proposed for the onset of hydroplaning. Insofar as the estimates for v and h in particular are subject to considerable inaccuracies, it seems appropriate that all future studies of fast-moving subaerial flows consider the possibility of hydroplaning.

It should be emphasized that equation 1 is a necessary but insufficient condition for hydroplaning. If the heights of the roughness elements protruding from the bed consistently exceed the thickness of the hydroplaning layer that would otherwise result, the layer may be broken up and hydroplaning may be suppressed. The fronts of many subaerial debris flows advancing over a bed partially or completely covered with water may not hydroplane for this reason. A subaqueous sea floor subject to long-term pelagic or hemipelagic sedimentation is likely to offer a relatively smooth surface over which the front may advance, whereas the steep surfaces on which subaerial debris flows occur are more likely to be covered with gravel, cobbles, boulders, and woody debris.

The inference that subaqueous debris flows can hydroplane over a broad range of conditions does not prove that hydroplaning is the cause of the long runout distances often seen in debris-flow deposits. Hydroplaning nevertheless is a viable mechanism deserving of further research. In a more conclusive experimental test, an attempt would be made to reproduce the long runout distances on low bed slopes. The facility that we used lacks the required length. A second way to test the hypothesis involves the use of a numerical model that (1) adequately describes both the flow mechanics of debris flows and the phenomenon of hydroplaning and (2) is capable of reproducing the experimental results obtained to date. We intend to pursue this avenue. The same numerical model should also be useful for studying surging and turbidity current formation.

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