Hydroplaning and submarine debris flows

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[1] Examination of submarine clastic deposits along the continental margins reveals the remnants of holocenic or older debris flows with run-out distances up to hundreds of kilometers. Laboratory experiments on subaqueous debris flows, where typically one tenth of a cubic meter of material is dropped down a flume, also show high velocities and long run-out distances compared to subaerial debris flows. Moreover, they show the tendency of the head of the flow to run out ahead of the rest of the body. The experiments reveal the possible clue to the mechanism of long run-out. This mechanism, called hydroplaning, begins as the dynamic pressure at the front of the debris flow becomes of the order of the pressure exerted by the weight of the sediment. In such conditions a layer of water can intrude under the sediment with a lubrication effect and a decrease in the resistance forces between the sediment and the seabed. A physical-mathematical model of hydroplaning is presented and investigated numerically. The model is applied to both laboratory- and field-scale debris flows. Agreement with laboratory experiments makes us confident in the extrapolation of our model to natural flows and shows that long run-out distances can be naturally attained.


1. Introduction

[2] Submarine landslides and debris flows occurring respectively in compacted and unconsolidated clays are relatively common events (on a geological timescale) in many areas around the continental margins. After release, they transform into a finely comminuted mixture of clay and water, with properties of a non-Newtonian liquid. Together with turbidity currents, which are much more diluted mass flows, submarine slides and debris flows represent the most effective process of sediment transport from the shallow continental margin into the deep ocean. These mass flows can often display very long run-out distances of more than 150 km, even on very gentle slopes, i.e., less than 1° [Embley, 1976; Hampton et al., 1996; McAdoo et al., 2000; Vorren et al., 1998; Hampton, 1972].

[3] The study of sediment turnover along continental margins is a central issue in marine geology, and of great interest not only for pure, but also for applied research. In fact, the reshaping of continental margins due to mass flows is an effective and fast process on the geological timescale. The release of large slides and debris flows can be a serious threat to offshore oil installations and pipelines, and represents a potentially destructive means to coastal settlements. Large tsunami waves can be generated by such events. For example, during the Storegga slide which occurred about 8150 years ago in the Norwegian Sea, a volume ≈2300–2500 km³ was transformed into a huge debris flow in the distal region [Bryn et al., 2002]. Calculations by Harbitz [1992] show that the flowing masses generated surface elevations of about 5 m in the source area.

[4] Data for subaerial debris flows show that the run-out distances are much smaller than for their subaqueous counterpart, while simple arguments would lead to the opposite conclusion. In fact, subaerial debris flows usually occur along steeper slopes and experience much less drag forces compared to their subaqueous counterparts. Moreover, the subaerial flows are subject to the full gravity force, while in the submarine environment the buoyancy effect approximately halves the effective gravity.

[5] The debris flows have mostly been described as a laminar viscoplastic flow by the so-called Bingham rheology model [Johnson, 1970; Huang and García, 1998, 1999; Hampton et al., 1996]. In this model, no deformation takes place until the stress in the material exceeds a yield stress. Deformation is then driven by the excess of stress (with respect to the yield value) applied to the material [Huang and García, 1998]. The viscoplastic concept seems valid for clay-rich subaqueous flows, but the very long run-out distances on gentle slopes require small yield stresses, probably incompatible with values deduced from core samples on the Norwegian continental
margin [Elverhoi et al., 2002]. Experiments at St. Anthony Falls Laboratory (SAFL) at the University of Minnesota (USA), may suggest another explanation. The experiments revealed that if the head of a small-scale muddy debris flow exceeds a critical velocity, which for laboratory flows is of the order of or less than 1 m/s, a thin wedge of lubricating water is trapped between the debris and the bed, effectively reducing bed friction and increasing mobility, a phenomenon known as hydroplaning [Mohrig et al., 1998, 1999].

[6] Hydroplaning of natural debris flows has never been directly observed. Rather, marine geologists can study the final product of this process which does not, however, lead to unique interpretation of the processes. Indications that hydroplaning might have occurred include long run-out distances, outrunner blocks, limited or no erosion, and thinning of the deposits. Of these, the most compelling piece of evidence is probably represented by the outrunner blocks, which are isolated blocks travelling hundreds of meters up to tens of kilometers from the front of the debris flow, apparently with reduced friction.

[7] To explain such phenomena, some alternatives to hydroplaning have been put forward. One possibility is a reduction of basal friction due to natural wetting of the sediments without dynamical effects of water [Hampton, 1970]. This mechanism, which is velocity independent, seems to be the best explanation for certain subaerial debris flows characterized by long run-out, deposition fragmentation and detachment of blocks associated with a temporarily wet bed [Hampton, 1970]. In contrast with the latter picture, experiments for subaqueous debris flows show the existence of a critical velocity before hydroplaning can start. Thus hydroplaning is a distinct mechanism of sediment transport. In addition, case studies have been reported where out runner blocks have travelled without apparent friction with a soft sediment at the bottom, probably indicating that the solid material was flowing at sufficient height over the seabed [Longva et al., 2003].

[8] It has also been suggested that the increase in water pore pressure associated with the overloading of a previously water-rich deposit could lead to large run-outs of certain subaqueous debris flows [Gee et al., 1999]. Although this model might explain high sediment mobility in special cases, it does not provide a general view applicable to the majority of subaqueous slides. The reason is that it requires two separate phases consisting in the placement of fine-graded sediment first, followed by the loading by an impermeable debris flow. It is unlikely that such a particular mechanism may be of general occurrence.

[9] In the present paper we argue that hydroplaning can provide an important means of sediment transport for a quite specific class of materials, namely, strong glacigenic muds which are common in the glaciated margins (e.g., in Norway or Canada). Other mechanisms of sediment transport might be more efficient than hydroplaning if the sediment does not remain rather compact. Talling et al. [2002] have suggested potential dilution of weak debris flows even at velocities as low as 2 m/s with rapid, complete transformation into a turbidity current. For debris flows of intermediate strength, wetting of the sediments and a consequent decrease in the yield stress might also be important. Except for experimental debris flows, at the present stage one cannot demonstrate the superiority of hydroplaning with respect to other means of sediment transport. However, as indicated by the experiments, hydroplaning gives such a dramatically improved run-out and velocity that the possible occurrence in the field must be carefully checked on the basis of the present knowledge.

[10] Analytical and numerical calculations based on the laws of fluid mechanics are important in supporting or ruling out the possibility that hydroplaning occurs in nature. The present paper follows up a series of contributions devoted to the theoretical aspects of submarine debris flows and hydroplaning. The most complete theoretical study of nonhydroplaning debris flows in the framework of Herschel-Bulkley rheology (of which Bingham rheology is a special case) was presented by Huang and García [1998] and a computer program was developed by Imran et al. [2001]. Simple estimates of the forces at work during hydroplaning were proposed by Mohrig et al. [1998]. More thorough analytical studies based on an adjusted version of the lubrication theory described by Batchelor [1967] were presented by Harbitz et al. [2003], who proposed a complete analytical model for a hydroplaning glide block. The model shows that hydroplaning at large scale is theoretically possible, and that a glide block can flow under different equilibrium conditions. The model provides insight to the physical conditions of water in the lubricating layer.

[11] The main purpose of the present paper is to present a more complete theoretical model for a deformable hydroplaning debris flow, which also provides the shape of the final deposit. We shall mainly be concerned with muddy flows, although some of the concepts expressed here could be more generally applicable to more sandy sediments.

[12] The extrapolation from the laboratory scale, where hydroplaning can be observed from start to stop under strictly controlled conditions, to the field-scale flows, where only the final product of much larger events can be observed, is by no means obvious. In order to provide a quantitative description of hydroplaning at the field scale, one has to rely on physical-mathematical modeling. The model presented in this paper describes the motion of a submarine flow taking into account the possibility of water intrusion underneath the debris flow. The Navier-Stokes and continuity equations, supplemented with appropriate boundary conditions between the different phases (water, mud and ground) are written in a Lagrangian fashion as a function of time. The model is two-dimensional, but vertical integration in both the mud and water layers makes it classifiable as a “1-D-depth-integrated” model. The physical model is solved numerically. As hydroplaning does not start immediately after the mud has been mobilized, and because not all of the sedimentary body hydroplanes, we have to describe a nonhydroplaning sediment as well. We shall treat a nonhydroplaning sediment as a Bingham fluid. We build our numerical approach on the framework of the BING model, developed at the University of Minnesota by Imran et al. [2001] to describe the flow of a nonhydroplaning debris flow. The hydroplaning model is applied to laboratory mudflows (the experiments are reported by Mohrig et al. [1998]) and large-scale debris flows. In the following, we loosely refer
to the flowing debris as “mud” because of the ability of clay-rich sediments to hydroplane.

2. A Model for Hydroplaning

2.1. Onset of Hydroplaning

The failure of muddy masses along the continental slope is probably triggered by overloading, earthquakes, increase of pore water pressure, release of gas hydrates or other phenomena. We assume that at failure, a homogeneous liquefied mass of sediment slips on a fixed ground. The leading forces responsible for the flow are proportional to the gravity acceleration parallel to the bed and to the pressure gradient parallel to the bed and resulting from variations in the height of the moving mass. Opposing these forces are the internal resistance of the material, the friction with the bed, and the drag force due to the interaction with ambient water. After release, the sediment accelerates until the drag force (which is proportional to the square of the velocity) and the bed friction become high enough as to balance the gravity force. However, before this velocity is reached, hydroplaning might occur, as discussed below.

For a body moving in an ideal (i.e., nonviscous) fluid, the relative fluid velocity at the front and at the rear points is zero. Because of the Bernoulli principle, this point acquires a higher pressure (stagnation pressure) with respect to the rest of the body given by

\[ P \approx \frac{1}{2} \rho_w u^2, \]

where \( \rho_w \) is the liquid density and \( u \) is the velocity. A real (viscous) liquid would also manifest a pressure increase at the front, but because of boundary layer effects the corresponding high-pressure point at the rear end can be missing. In general, the pressure field around the body changes in a complicated manner, but the maximum pressure differences around the body’s surface are still of the order of the stagnation pressure. Zones of low and high pressure may develop, depending on the velocity field around the body. At the front of the debris flow the fluid is halted relative to the mud, and the highest pressure is generally reached at the stagnation point close to the bottom front of a nonhydroplaning debris flow. If the pressure becomes larger than weight, some water may intrude under the debris flow. The condition reads

\[ (\rho_d - \rho_w)g \cos \theta DWL < PWL, \]

where \( \rho_d \) is the mud density, \( g \) is the acceleration of gravity, \( \theta \) is the slope angle, \( W, L \) and \( D \) are the respective width, length and depth of the sediment affected by the water intrusion. From the above equations a critical velocity \( V_{crit} \) can be found approximately as

\[ V_{crit} \approx F_{rcrit} \left[ \frac{g \cos \theta (\rho_d - \rho_w)D}{\rho_w} \right]^{1/2}, \]

where \( F_{rcrit} \) is the critical densimetric Froude number. Experimentally, it is found that hydroplaning starts when the Froude number approaches a critical value substantially smaller than unity, typically around \( F_{rcrit} \approx 0.3 \) [Mohrig et al., 1998]. When the critical velocity is reached, a wedge of water penetrates underneath. Because water cannot be expelled immediately from the front because of inertial and viscous forces, it will penetrate further, as indicated by the experiments. In the case of a real, three-dimensional debris flow, some water can penetrate sideways or mix into the sediment. Because of the intrinsic two-dimensionality of the model, this possibility will not be considered in the present work.

The previous estimate for the critical velocity (with Froude number \( F_{rcrit} = 0.3 \)) gives values between about 0.3 m/s and 0.6 m/s for heights between 5 cm and 20 cm characteristic of the lab scale, while for natural debris flows with heights between 10 m and 100 m the critical velocities are estimated in the range between 4 m/s and 18 m/s. The measured velocities of natural debris flows deduced from the time of cable breaking, can be much larger than the values above [Hooke and Ewing, 1999]. We thus expect that the kinematic condition for the onset of hydroplaning can be satisfied in natural debris flows.

As the velocity increases from zero to the critical value, a force arising from irregularities in pressure distribution might affect irregularly the sediment and tend to deform and ultimately to disrupt it. If the sediment is sufficiently cohesive (e.g., pure clay), the debris flow might avoid fragmentation and remain one single body (fractures and faults might be formed in the body, however, and water might be incorporated). For debris flows with sandy composition the cohesion between grains is much smaller. Some parts of a sandy debris flow might be disrupted if the stagnation pressure and shear stresses are much larger than the cohesion of the material. In addition, for a sandy body it is more difficult to maintain a pressure difference between the top and the bottom, because water more easily can penetrate inside, owing to the larger water permeability. In fact, laboratory experiments show that sandy bodies do not hydroplane. In the next section we present a more detailed picture of the front of a debris flow.

2.2. Physical State of a Debris Flow During Hydroplaning

In order to investigate the state of a debris flow head during hydroplaning, we studied numerically the pressure and velocity field distributions around an object flowing against resting water and located a fixed height from the bottom. An object with the simplest possible shape, namely, a rectangle with smoothed corners, was first examined. The distance from the bed was fixed, and the angle of attack was zero. Obviously, the detailed shape of the object does affect the results of the computation, and in a real debris flow one should expect more irregular shapes and orientations. However, the simple heuristic approach followed here may reveal the possible average patterns of the pressure and water flow around the object. Using a finite element algorithm, the Navier-Stokes equations were solved, and pressure and velocities were calculated for a stationary flow small scale. In Figure 1 the pressure is shown with colors and the velocity field relative to the moving body with arrows. When the object is very close to the ground a pressure gradient results, both along the vertical and the horizontal directions. The vertical gradient results from the reduced pressure on the top and increased pressure at the bottom (for clarity the hydrostatic pressure term is
absent from the figure). When the pressure difference becomes equal to the submerged weight, water is intruded underneath and hydroplaning starts. The horizontal gradient under the body is also an important parameter. As shown by Harbitz et al. [2003] and as discussed further in the present paper, this gradient is responsible for pushing water further under the sediment. From the present calculations we find that if the water layer thickness is much smaller than the sediment thickness, the horizontal pressure gradient is roughly constant and of the order of the stagnation pressure divided by the sediment length.

The experiments show that a small-scale debris flow has an elongated head with the front slightly lifted up. Such details, apparently insignificant, might greatly enhance hydroplaning. Therefore a more realistic shape as suggested by debris flows experiments was also considered.

Figure 2 shows the result of the calculation. The high pressure at the bottom will be efficient in both deforming the sediment at the front and pumping water underneath. Note that the front is relatively thin, which explains the small value observed for the critical Froude number in the experiments. Both vertical and horizontal pressure gradients are important elements in the physics of hydroplaning.

The motion of an object through an ambient fluid at rest can produce a complex pattern because of the irregularity in the pressure distribution along the surface of the body, resulting in a finite torque which depends on the shape of the body and which will rotate the head of the debris flow from the bottom up. In general, the results indicate that the head is lifted first. The simulations show that when the thickness of the water layer is approximately 1 cm, the effective vertical pressure and weight per unit area are comparable (in magnitude). A more detailed study of the forces and velocity field for different geometries would certainly be interesting. Similar calculations performed with a different computational scheme (finite volume) show comparable values for the pressure gradients.

We notice that the pressure gradient distribution under the debris flow depends on the inclination of the bottom of the debris flow with respect to the seabed. A constant thickness of the water layer results in a high-pressure gradient at the front and a more or less constant pressure under the debris flow, while a progressively decreasing water layer thickness like in Figure 2 gives a more regular pressure gradient.

Lacking any direct data on the flow of large-scale debris flow, one may choose a hypothetical shape based on geometrical rescaling of experimental small-scale debris flows. An estimate of the velocities involved to sustain hydroplaning in the large-scale situations can be obtained with the results from Figure 2 as a basis. We consider a debris flow with height of 10 m and correspondingly 1 m above the ground, and assume that the pressure scales with the square of the characteristic velocity. From this we find that the lift force becomes comparable to the weight of the sediment, so that the water layer can be sustained, when the velocity exceeds 10 m/s.

2.3. Constitutive Equation for the Flowing Mud

The viscoplastic mudflow is here modelled as a Bingham fluid. In such a fluid, there is a finite yield stress beyond which the fluid shear depends linearly on the stress. This kind of fluid represents a good model for muddy debris flows [Johnson, 1970], while it might give an oversimplified view for more sandy materials owing to grain-grain interactions that imply Coulomb friction and a dependence of the rheological parameters on the normal stress. Generalizations such as the so-called Herschel-Bulkley rheology have also been proposed for debris flows. We shall here describe the more general Herschel-Bulkley rheology, but in the application we always use a simpler Bingham concept (corresponding to an index \( n = 1 \) in the following equations). In Herschel-Bulkley rheology the stress is related to the velocity by

\[
\tau = \tau_y + \mu \left( \frac{\partial u}{\partial y} \right)^n \quad (\text{if } |\tau| \geq \tau_y),
\]

Figure 1. The pressure [kPa] around a rectangular object with height 10 cm moving 5 mm above the bed with a velocity of 1 m/s. The velocity of water relative to the object is indicated by arrows. See color version of this figure at back of this issue.

Figure 2. The pressure [kPa] and water velocity field around an object shaped as a debris flow head moving with a velocity of 1 m/s. The shape of subaqueous debris flows seen in experiments has been used. See color version of this figure at back of this issue.
where $\mu_w$ is the dynamic viscosity. The shear rate $\gamma$ is expressed as [Huang and García, 1998; Imran et al., 2001]

$$\frac{\gamma}{\gamma_r} = \begin{cases} 0, & \text{if } |\tau| < \tau_y, \\ \text{sign}(\gamma) \left( \frac{|\tau|}{\tau_y} - 1 \right)^{1/n}, & \text{if } |\tau| \geq \tau_y, \end{cases} \quad (5)$$

where $\gamma_r \equiv (\tau_y / \mu_w)^{1/n}$ is the reference strain rate.

[24] For a steady gravity mudflow down an inclined plane the shear stress increases linearly with depth because of the weight of the overburden mass. In Herschel-Bulkley rheology the flow can be characterized by an unsheared or plug region and a shear layer, where the shear stress does not exceed the yield strength. Below the plug layer there is a shear layer where the stress is sufficient to provide a shear flow.

[25] In the above, we have introduced a model for sediment rheology as a continuous and deformable medium, tacitly assuming that the material remains compact during the flow. Experiments show that not all sediment compositions lead to hydroplaning. In order to hydroplane, the sediment must have a critical amount of clay versus sand. In fact, empty spaces present between grains make the sediment permeable. For a sandy sediment, two consequences might be important. First, some of the water trapped below the hydroplaning sediment can penetrate inside the sediment because of the pressure gradient. Second, hydroplaning itself might not start if water can percolate inside the sediment from the front within the flowing times. A simple estimate of permeability is provided by appropriate modification of the Kozeny-Carman relation for a porous medium [Mitchell, 1992]

$$k = C d^2 \frac{P'}{\mu_w} \frac{e^3}{1 + e} S^3, \quad (6)$$

where $d$ is the particle diameter, $P'$ is the pressure gradient, $e$ is the void ratio (the fraction of voids to solid), $S$ is the degree of saturation and $C$ is a composite form factor accounting for the packing of the granular medium. Although this formula has only qualitative value, the dependence on particle size (contained also in the factor $e$) describes well the experimental data. For compacted, fully saturated clays ($S = 1$), the value of $k$ is of the order

$$k \approx 3 \times 10^{-6} \text{cms}^{-1}.$$ 

while for sand $10^5$ times larger values are possible. These numbers show qualitatively that a sandy debris flow will be less effective in maintaining differences in water pressure. This is clearly observable in laboratory mass flows. In addition, muds can have high cohesion because of electrostatic grain-grain forces, and this contributes substantially to decrease the permeability. However, some amount of water might intrude in regions with many tension cracks formed by high stresses. In the following, we consider only strong sediments, where the sand content is sufficiently small for the mud to remain compact during the flow.

2.4. Governing Hydrodynamical Equations

[26] We consider a two-dimensional unsteady laminar mudflow down an impermeable rigid slope with inclination $\theta$ as shown in Figure 3. The depth of the mud perpendicular to the slope is denoted $D$ and the depth of the lubricating layer is denoted $D_w$. A Cartesian coordinate system $(x, y)$ is defined with the $x$ axis down slope and the $y$ axis upward perpendicular to the bed. The longitudinal and transverse velocities are denoted $u$ and $v$, respectively. The separation of the mudflow into plug and shear layers with respective thicknesses $D_p$ and $D_s$ is not explicitly shown in the figure. No mass flux between the water layer and the mud is assumed.

[27] We assume that the flow depth is small relative to the characteristic flow length and that the flow depth changes relatively slowly in the longitudinal direction. Thus the $y$ components of the acceleration and shear stress have been neglected. The mud flow may then be approximated by the following equations [Huang and García, 1998]:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} = \frac{1}{\rho_d} \left( - \frac{\partial P}{\partial x} + \frac{\partial \tau_{yx}}{\partial y} \right) + g_x, \quad (7a)$$

$$0 = - \frac{1}{\rho_d} \frac{\partial P}{\partial y} - g_y, \quad (7b)$$

where $\rho_d$ is the mud density, $g_x = g \sin \theta$ and $g_y = g \cos \theta$ are the relevant components of the acceleration due to gravity and $P$ is the pressure. Equation (7) represents conservation of mass, while equations (8) and (9) represent momentum conservation in the $x$ and $y$ direction, respectively.

[28] A hydroplaning mudflow moves on top of a thin water layer. The water flow is governed by equation (7) together with the momentum equations:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + \frac{1}{\mu_w} \left( - \frac{\partial P}{\partial x} + \frac{\partial \tau_{yx}}{\partial y} \right) = g_x, \quad (10)$$

$$0 = - \frac{1}{\mu_w} \frac{\partial P}{\partial y} - g_y, \quad (11)$$

which correspond to equations (8) and (9) with $\rho_d = \rho_w$. 

The bed conditions yield

\[ u(y = 0) = 0; \quad v(y = 0) = 0. \]  

Furthermore, the vertical velocities at the top and the bottom of the mud layer are given by the kinematic boundary conditions

\[ v(y = D + D_w) = \frac{\partial (D + D_u)}{\partial t} + u_p \frac{\partial (D + D_u)}{\partial x}, \]  

and

\[ v(y = D_u) = \frac{\partial D_u}{\partial t} + u_p \frac{\partial D_u}{\partial x} , \]  

respectively, where \( u_p = u(D + D_u) \) is the velocity along the bed at the top of the mud and \( u_w = u(D_u) \) is the velocity at the interface between the water and the mud.

The pressure is divided into hydrostatic and dynamical parts. In this section we will discuss only the hydrostatic pressure. The dynamic pressure is important for hydropelaning and is discussed in section 2.6. The hydrostatic pressure at the top of the mud depends only on the height of the overlying water, i.e.,

\[ P(y = D_u) = \rho_w g H - \rho_w g_y (D + D_u) , \]  

where \( H \) is the water depth. The pressure \( P \) in the region \( D_u < y < D_u + D \) is obtained by integrating equation (9) with respect to \( y \) from \( y = D_u \) to \( y = D_u + D \) using the boundary condition (15). This yields the (hydrostatic) pressure in the mud layer

\[ P(D_u \leq y < D_u + D) = \rho_w g H - \rho_w g_y (D + D_u) \]  

\[ + \rho_y g_y (D + D_u - y) . \]  

The hydrostatic pressure in the water layer is obtained by further integration using equation (11), which gives

\[ P(0 \leq y < D_u) = \rho_w g H - \rho_w g_y (D + D_u) + \rho_y g_y (D_u - y) . \]  

Using the relation \( \frac{dD}{dx} = \sin \theta \) to eliminate the derivative of \( H \) with respect to \( x \), the respective hydrostatic pressure gradients relevant for the motion along the bed are then

\[ \frac{\partial P}{\partial x} = \rho_w g_x + (\rho_d - \rho_w) g_y \frac{d}{dx} [D + D_u] , D_u \leq y \leq D + D_u \]  

for the mud layer, and

\[ \frac{\partial P}{\partial x} = \rho_w g_x + (\rho_d - \rho_w) g_y \frac{dD}{dx} , 0 \leq y < D_u \]  

for the water layer. Note that the hydrostatic pressure gradient in the mud depends on the variation in the total height of water plus mud, while in the water layer it is independent of the water thickness.

### 2.5. Nonhydropelaning Mudflow

For the nonhydropelaning sediment we adopt the so-called BING model developed by Imran et al. [2001]. The mudflow is divided into an unsheared layer called the plug layer with thickness \( D_u \) and velocity \( u_p \), independent of \( y \), overriding a shear layer with thickness \( D_s \), where the velocity decreases gradually from \( u_p \) at \( y = D_u \) to \( u_w \) at \( y = D_u + D_s \). The equations for the fluid flow are derived by integrating equations (7)–(9). Details for the derivation are given by Imran et al. [2001] and Huang and García [1998]. We are interested in the equations given in the reference frame following the average velocity of the mud \( \bar{V} \), defined by

\[ \pi D = u_D D_g + \int_0^{D_g} u(y) dy . \]  

Mass conservation is then fulfilled by

\[ \frac{dD}{dt} = -D \frac{\partial \pi}{\partial x} . \]  

Equations for the plug velocity and the average velocity are expressed by

\[ \frac{du_D}{dt} = (\bar{u} - u_p) \frac{\partial u_D}{\partial x} - \Delta \rho g \frac{\partial}{\partial x} (D + D_u) - \frac{\delta \tau_0 + \tau_1}{D_p \rho_d} + \Delta \rho g_x , \]  

and

\[ \frac{d\pi}{dt} = \frac{1}{D} \frac{\partial}{\partial x} \left( \bar{u}^2 D - \frac{1}{1 - \alpha_1} u_p u_D + \frac{\alpha_1 - \alpha_2}{1 - \alpha_1} u_p^2 D \right) \]  

\[ - g_x \rho \frac{\partial}{\partial x} D - \frac{\tau_b + \tau_1}{\rho_d} + g_y \Delta \rho , \]  

respectively. The shape parameters within Herschel-Bulkley rheology are given by

\[ \alpha_1 = \frac{1}{1/n + 2} , \quad \alpha_2 = 1 - \frac{2}{n + 2} + \frac{1}{2/n + 3} , \]  

and reduce to \( \alpha_1 = 2/3 \) and \( \alpha_2 = 8/15 \) for Bingham fluids \((n = 1)\). Furthermore, the shear stress at the bottom is given by

\[ \sigma_b = \text{sign}(u_p) \tau_b \left[ 1 + \frac{n + 1}{n} \frac{u_p}{\gamma D_s} \right]^{\alpha} . \]  

At the top of the mud the shear stress is proportional to the frictional drag force

\[ \tau(y = D + D_u) = -\tau_1 . \]  

It is found to be important only for the hydropelaning flow and will be discussed further in section 2.6.

### 2.6. Hydroplaning Plug Model

When hydropelaning occurs the stress in the mud is expected to reduce considerably. If the shear stress in the hydropelaning debris flow does not exceed the yield strength, the flow may be treated as a hydropelaning plug; that is, the shear layer thickness becomes essentially zero. The hydroy
The hydroplaning plug is characterized by a constant velocity for the whole cross section of the debris flow

\[ u_p = u_w = \bar{u}, \]

where \( \bar{u} \) is the average velocity of the mud. The correct order of magnitude for a hydroplaning mudflow may be obtained by \( \tau = \mu_w u_p / D_w \). Data from experiments in small, confined settings suggest values of about \( u_p = 0.5 \text{ m/s}, D_w = 0.5 \text{ cm} \) and if we assume a viscosity \( \mu_w = 0.01 \text{ Pa s} \) for the water slurry, one finds \( \tau = 0.01 \text{ Pa} \ll \tau_s \approx 10–100 \text{ Pa} \). Concerning the field scale, assuming a debris flow velocity of 10 m/s and a water layer thickness of 10 cm we find \( \tau / \tau_s \approx 10^{-4} \) for \( \tau_s = 5 \text{ kPa} \). We conclude that a small amount of water is sufficient to depress the shear stress at the base of the mud under values substantially lower than the yield stress. In the numerical calculations of large-scale debris flows we will thus assume that the shear layer is absent during hydroplaning. This permits some simplification of the equations and especially avoids dealing with the very thin shear layers, which might cause numerical instability. The momentum equation for the hydroplaning plug is obtained by integrating equation (8) over the whole mud layer, i.e., from \( D_w \) to \( D_w + D \). The integration of the left-hand side is trivial since the velocity \( u = \bar{u} \), \( \mu_w \frac{\partial u}{\partial x} = 0 \). The pressure gradient given by equation (18) and the gravity term are independent of \( y \). The stress at the bottom of the mudflow is equal to the stress at the top of the water layer, \( \tau_w \), and \( \tau_s \) is the stress at the top of the mud layer. The hydroplaning plug model is obtained by

\[
\frac{dh}{dt} = -\rho g \frac{\partial}{\partial x} (D + D_w) - \tau_s - \frac{\tau_s}{D_w} + \Delta \rho g_s + f_D.
\]

The thickness \( D \) of the mudflow is determined by mass conservation and is given by equation (21). For the hydroplaning flow, drag forces are found to be very important. These forces have been accounted for by the frictional drag

\[
\tau_f = -\frac{1}{2} \rho_w C_f u_p^2.
\]

and the pressure drag term

\[
f_p = -\frac{1}{2} \rho_w C_p \frac{\partial D}{\partial x} u_p^2.
\]

The part proportional to \( C_f \) is the (specific) drag force for a flat plate (the upper face), while the part proportional to \( C_p \) accounts for the finite thickness of the sediment. A space derivative results because the drag force must be included only for the elements of the sediment that are not shadowed by other sediment situated in front.

\[ [35] \text{The pressure coefficient } C_p \text{ can be estimated by measuring the drag force on objects of given thickness. In our study, the value of the coefficients was fitted from standard literature [e.g., Newman, 1977]. The drag forces turn out to be important for the front part of the debris flow, which travels at a speed twice as high or more than the rear part. In the model, we do not account for added mass, namely, the possible acceleration of seawater by the moving body. This should be a good approximation for flat debris flows considered in the present study.} \]

\[ [34] \text{The magnitude of the stress at the top can be estimated for the small-scale simulations where, we applied a frictional drag coefficient } C_f = 0.005. \text{ The maximum velocities obtained were } 0.75 \text{ m/s one obtains } \tau_s \approx 1 \text{ Pa}, \text{i.e., much less than the yield stress in the laboratory experiments, which is typically } 50 \text{ Pa. For the field-scale simulations we applied slightly less values for the drag coefficient } C_f = 0.003. \text{ The maximum velocity obtained was } 70 \text{ m/s, which correspond to a stress } 7 \text{ kPa. This is less than the smallest yield stress in the calculations } 10 \text{ kPa.} \]

\[ 27. \text{ Water Layer} \]

\[ [35] \text{To determine the water flow properties we need to parameterize the velocity profile. The parameterization is constrained in the stationary situation where the water flow reduces to the so-called Couette flow. This suggests a parabolic form of the water profile:} \]

\[
\frac{\partial u}{\partial y} = ay + b, \quad u = \frac{1}{2} ay^2 + by. \tag{29}
\]

where \( a \) and \( b \) are constants. Rather than using these constants in our numerical model we characterize the water flow by the velocity \( u_w \) at the top of the water layer, which is related to \( a \) and \( b \) by

\[
u_w = \frac{1}{2} a D_w^2 + b D_w \tag{30}
\]

and the water discharge which is defined by

\[
f_w = \int_0^{D_w} dy \langle y \rangle = \frac{1}{6} a D_w^3 + \frac{1}{2} b D_w^2. \tag{31}
\]

In order to partially account for the nonlinear terms representing water inertia on the left hand side of equation (10), we keep the parabolic form for the water profile, equation (29), but express the parameters \( a \) and \( b \) in terms of \( f_w \) and \( u_w \), which are calculated from the general nonlinear case, whence the velocity profile, equation (29) is expressed as

\[
u(y) = \left( \frac{3 u_w}{D_w^2} - 6 \frac{f_w}{D_w^2} \right) y^2 + \left( 6 \frac{f_w}{D_w^2} - 2 \frac{u_w}{D_w} \right) y. \tag{32}
\]

With these relations we express the shear stresses \( \tau_b \) and \( \tau_w \) in terms of \( u_w \) and \( f_w \) for the steady flow

\[
\tau_b = \mu_w \left| \frac{\partial u}{\partial y} \right|_{y=0} = \mu_w \left( 6 \frac{f_w}{D_w} - 2 \frac{u_w}{D_w} \right), \tag{33}
\]

\[
\tau_w = \mu_w \left| \frac{\partial u}{\partial y} \right|_{y=D_w} = \mu_w \left( 4 \frac{u_w}{D_w} - 6 \frac{f_w}{D_w} \right). \tag{34}
\]

[36] A more general model of the velocity profile fully accounting for the nonlinear terms has been discussed by Harbitz et al. [2003], where a hydroplaning flat plate implying a linear increase in water depth with distance is
assumed to close the equations and calculate the pressure gradient. In the present study, irregular changes in the water profile are accounted for and this requires a calculation of the water depth as a function of both time and position. With this approximation equations (33) and (34) have to be used, with \( u_w \) and \( f_w \) to be calculated at every time step and for each element. The velocity profile, equation (32) can then be used to account for the nonlinear terms in the Navier-Stokes equations.

(37) The continuity equation for the water flow is obtained by integrating equation (7) with respect to \( y \) over the water layer from 0 to \( D_w \). This gives

\[
\int_0^{D_w} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dy = \frac{\partial}{\partial x} \int_0^{D_w} u(y) dy - u_w \frac{\partial D_w}{\partial x} + v(D_w)
\]

(35)

where \( v(D_w) \) is obtained from equation (14). We identify the water discharge \( f_w \). This gives

\[
\frac{\partial D_w}{\partial t} + \frac{\partial f_w}{\partial x} = 0
\]

(36)

To obtain the thickness of the water layer in a reference frame following the average velocity of the mudflow we apply the derivative

\[
\frac{d}{dt} = \frac{\partial}{\partial t} + \mathbf{v} \cdot \nabla
\]

(37)

The thickness of the water layer is then obtained by

\[
\frac{d D_w}{dt} = \bar{u} \frac{\partial D_w}{\partial x} - \frac{\partial f_w}{\partial x}.
\]

(38)

(38) Momentum conservation in the water layer is fulfilled by equation (10). Note that there is no net contribution from the gravitational forces acting on the water layer. The accelerating forces on the water layer is caused by the weight of the mud together with the induced stress. Integrating the equation over the water layer, i.e., \( y = 0 \) to \( y = D_w \) yields

\[
\int_0^{D_w} \left( \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} \right) dy = \int_0^{D_w} \left( - \frac{1}{\rho_w} \frac{\partial P}{\partial x} + \frac{\tau_{xy}}{\rho_w} \right) dy.
\]

(39)

The first term on the left-hand side of the equation can be rewritten as

\[
\int_0^{D_w} \frac{\partial u}{\partial t} dy = \frac{\partial f_w}{\partial t} - u_w \frac{\partial D_w}{\partial t},
\]

(40)

while the two other terms give

\[
\int_0^{D_w} \left( u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} \right) dy
\]

(41)

We define the momentum flux in the water layer by

\[
J_w = \int_0^{D_w} dy \ u^2(y).
\]

(42)

Using \( J_w \) and \( f_w \) and inserting the expression for \( v(D_w) \) from equation (14), we get

\[
\int_0^{D_w} dy \left( \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} \right) = \frac{\partial f_w}{\partial t} + \frac{\partial J_w}{\partial x}.
\]

(43)

(39) The right-hand side of equation (39) gives

\[
\frac{1}{\rho_w} \int_0^{D_w} \left( - \frac{\partial P}{\partial x} + \frac{\tau_{xy}}{\rho_w} \right) dy = - g_y \rho_d - \frac{\rho_d - \rho_w}{\rho_w} g_y D_w \frac{\partial D_w}{\partial x} + \frac{\tau_w - \tau_b}{\rho_w}.
\]

(44)

where we have used the pressure given by equation (19), which is independent of \( y \). Approximate expressions for the stresses \( \tau_b \) and \( \tau_w \) at the bottom and top of the water layer are given by equations (33) and (34). From this we obtain the following equation of motion for the water flow:

\[
\frac{\partial f_w}{\partial t} + \frac{\partial J_w}{\partial x} = - \frac{\rho_d - \rho_w}{\rho_w} g_y D_w \frac{\partial D_w}{\partial x} + \frac{\tau_w - \tau_b}{\rho_w}.
\]

(45)

The quantities \( J_w \) as well as \( \tau_w \) and \( \tau_b \) are calculated using the velocity profile given by equation (32). The momentum flux is expressed in terms of \( u_w \) and \( f_w \) as

\[
J_w = \int_0^{D_w} u^2(y) dy = \frac{1}{20} a^2 D_w^5 + \frac{1}{4} a b u_w^4 + \frac{1}{3} b^2 D_w
\]

(46)

We insert the expression for \( J_w \) given by equation (46) and to obtain the water discharge as a function of time we use the Lagrangian derivative defined by equation (37). This gives

\[
\frac{df_w}{dt} = \left( \frac{6}{5} \rho - \frac{12}{5} f_w \right) \frac{\partial f_w}{\partial x} + \left( \frac{6}{5} f_w^2 \frac{2}{5} D_w^2 \frac{2}{15} \right) \frac{\partial D_w}{\partial x}
\]

(47)
The stresses on the water layer are obtained by equations (33) and (34). This equation is solved together with the continuity equation for water, equation (38). Note that in equation (47) we have written explicitly the gradient of the stagnation pressure. This quantity is not easy to calculate. A simple estimate suggests a pressure gradient of the form \( \frac{\partial P_{stag}}{\partial c} \approx \frac{1}{2} p_{stag} r_{wup}^2 / L \), where \( L \) is the length of the sediment. The increase of water thickness from the rear end to the front gives an additional contribution to the pressure gradient (for a steady laminar stationary flow the linearly increasing water thickness this gradient can be calculated analytically [Harbitz et al., 2003]). However, in the present work we use the simple approximation of a constant pressure gradient. The numerical calculations with finite element method presented in Section 2.2 show that this approximation is not unrealistic.

3. Numerical Model

3.1. General

The coupled equations of motion for the mud and the water are solved numerically. The numerical model is an extension of the BING code developed by Imran et al. [2001], and reduces to BING when hydroplaning is absent. The entire flowing mass is divided into vertical elements, which at the beginning of the calculation are equally spaced. \( N + 1 \) nodes limit \( N \) elements of sediment.

The acceleration of a node when hydroplaning is taking place is calculated from equation (26) using a finite difference scheme, analogous to Imran et al. [2001]. When the acceleration of one node is calculated, its new velocity and position after a finite time step can be found. Since adjacent nodes can have different accelerations, their relative velocities and positions will change as a function of time. The conservation of the total amount of mud is ensured by calculating the new height of the mud \( D(i) \) so that the cross area \( A(i) \) of all the elements \( i = 1, N \) is constant during the calculation

\[
D(i) = \frac{A(i)}{X(i+1) - X(i)},
\]

where \( X(i) \) is the position of the \( i \)th node.

When the velocity of the front node exceeds a critical value for hydroplaning given by equation (3), the calculation of water flow is initiated according to equations (47) and (38), and we impose that a certain number of nodes at the front are hydroplaning. The initial value of the water depth is assumed to be maximal at the front, and decreases toward the rear part of the mud.

Concerning the critical Froude number for the onset of hydroplaning, it has been discussed that the critical value observed in the laboratory is about (\( Fr_{crit} = 0.3 \)). The small value of the critical Froude number in the laboratory is probably due to the thinning of the sediments at the tip of the debris flow. The critical Froude number for hydroplaning in a large-scale debris flow is basically not known. In the present calculations for the large-scale we preferred to be conservative, and set a critical Froude number \( Fr_{crit} = 1 \). The water velocity profile and depths are calculated for each node. We assume that hydroplaning takes place at one given node if the water thickness \( D_w(i) \) at the node \( i \) is larger than a fixed value, representing the size of the roughness height present at the bottom of the sea and in the mud. Typical values we use for the minimal water thickness are a few millimeters and a few decimeters for the small and the large scale, respectively. When hydroplaning does not take place, equations (22) and (23) are used, corresponding to the BING model [Imran et al., 2001].

We impose hydroplaning to stop when the thickness of the water layer decreases to the critical thickness. To simulate this situation, the part of the code without hydroplaning is called again. Usually this provokes a rapid arrest of the flow.

3.2. Initial Volume and Configuration

In order to start the simulation, some initial conditions must be assumed for the water and the mud. The initial shape of the mud is assumed parabolic [Imran et al., 2001]. This is certainly not very realistic for large-scale calculations, where the initial shape is more complex. Since the original profile of the initial mass is essentially unknown (even if it can be partially estimated in some cases), most of the calculations reported here will necessarily start from a simplified shape also for the large-scale simulations. In particular, a parabolic shape is described by two arbitrary parameters, the height \( H \) and the length \( L \), while their product \( HL \) can be better constrained by field observations.

To summarize, the sequence of calculations proceeds as follows. The simulation starts with a fixed shape of the deposit which evolves according to the BING model [Imran et al., 2001] implemented with the drag force term. The velocities of the mudflow in the Bing model is obtained by equations (22) and (23). Equation (3) is used to compare the front velocity of the debris flow with the critical velocity for hydroplaning (an average from the first three cells is used). When the front velocity becomes the larger of the two, the program fixes an initial shape for the water profile, typically linear or parabolic. Then equations (21) and (26) are used to calculate the thickness and motion of the mud. These must be coupled with equation (47), which determine the water flux and equation (38) for the evolution of the thickness of water layer. These coupled equations are called sequentially and the result of each calculation is updated at every time step.

4. Results

4.1. Comparison With Laboratory Experiments

Small-scale laboratory experiments provide a unique way to closely study the dynamics of the flow. Experiments of artificial muddy debris flows performed in the last decade at the St. Anthony Falls Laboratory at the University of Minnesota by G. Parker, D. Mohrig, J. Marr, and others, provide a large set of available data [Mohrig et al., 1998, 1999]. They have performed various experiments, where mud is released in confined as well as unconfined tanks. These experiments clearly show the onset of hydroplaning, mud acceleration and head detachment. We compare our numerical model with experiments made by Mohrig et al. [1999] where the mud was released into a transparent channel approximately 10 m long, 3 m deep and 20 cm wide. The channel was segmented with a break in slope, the upper and lower slope angles being 6° and 1°, respectively.
The slope break was located at 5.7 m downslope from the position where the material was released into the tank. Approximately 30 l of debris slurry was released from a large box at the upper end of the facility by opening a gate to form a slot with height 20 mm and width 170 mm. The total time to drain the tank was 3.5 s. In the experiments yield stresses are 49 Pa and 36 Pa and viscosities are 0.035 and 0.023 Pas, respectively. The density of the slurry was 1600 kg/m³ in both cases. Subaerial as well as subaqueous experiments were performed.

The initial conditions in the experiments where the mud is released through a thin slot, lead to strong shear forces and inertial forces at the beginning that cannot be treated correctly within our numerical model. Note that the inertial forces within the Bing model are only indirectly accounted for through the variation of the thickness of the sediment during the flow. In the simulations we applied an initial parabolic mud profile at rest with front position at zero. The initial length was 2.0 m and the maximum height was 0.14 m. The mud is allowed to freely pass position zero. These initial conditions generally yield slightly higher mud mobility and run-out distances compared to the ones in the experiments. However, here we want to focus mainly on the difference between subaerial and subaqueous mudflows seen in the experiments. The best way to do this is to compare the experiments with calculations that give the same run-out distance in the subaerial case. We therefore applied slightly higher yield strengths in our calculations (60 Pa and 50 Pa) than the ones in the corresponding experiments (49 Pa and 36 Pa), while the densities and viscosities were the same. Figures 4(top) and 5(top) show the final deposit shapes. The dashed lines represent the results from calculations for subaerial flows. These calculations reproduce the deposits in the experiments indicated by stars. For the subaqueous flows we have performed calculations using the Bing model as well as the hydroplaning plug model. We applied a frictional drag coefficient $C_F = 0.005$ and a pressure drag coefficient $C_P = 0.14$ as extrapolated from experiments on cylindrically shaped models at different Reynolds numbers [Newman, 1977], and for the hydroplaning model we assumed a viscosity in the water layer $\mu_w = 0.01$ Pas. The deposit profiles obtained with the Bing model are shown by the dotted lines. The Bingham fluid stabilizes after only 1 m and clearly a nonhydroplaning model fails to reproduce the long run-out distance observed in subaqueous flows (squares). The results obtained by the hydroplaning plug model are shown by solid lines in the figures. We can see that hydroplaning provides a reasonable description of the laboratory data for the final deposit profile. In addition, water intrusion under the head causes the front to accelerate and to almost detach from the rest of the flow, a feature that does not appear in the simulations without hydroplaning.

In Figures 4(bottom) and 5(bottom) the head velocities are shown as a function of the distance. For the
experiments the initial velocity was well above 1 m/s when
the mud entered the tank. Thus the early stage velocities
from the experiments are not comparable with our calcu-
lations. The velocities for the subaerial flows calculated
with the BING model seem to slightly overestimate the true
velocities at late stages. In the subaerial experiments the
velocities were reduced to less than 1 m/s after only a few
meters.

[52] In the subaqueous flow, the pure Bingham visco-
plastic flow does reach velocities comparable to the exper-
imental data at the beginning, but quickly decelerates and
comes to rest. The hydroplaning model, however, is able to
reproduce front velocities as well as deposit thickness.

[53] Without including hydroplaning, according to our
simulations, the sediment would have a longer run-out
distance in air than in water, simply due to the larger drag
force and the reduced effective gravity force in water.
Including hydroplaning in our simulations, we reproduce
the experimental data well and avoid the “air-water para-
dox”. Obviously, we do know that hydroplaning occurs in
the laboratory. The challenge is to extend the previous
findings to the large scale, where hydroplaning has never
been observed, but might offer an explanation for the
observed run-out distances and limited erosion features.

4.2. Hydroplaning in Natural Debris Flows

[54] Having tested the model at the laboratory scale, the
main objective is now to predict the behavior of the hydro-
planing flow at the field scale. Unfortunately, the only
information available at the field scale (at least partially)
is the thickness, the shape and the rheological properties
of the final deposit. No large submarine flows have ever been
studied during the flow. There have been attempts to trigger
artificially the release of a 10^4 m^3 debris flow by explosives,
and follow it with geophones and acoustic sounders [By,
1991]. However, in an attempt performed in Canada, the
expected slide was not released by the explosion. The only
dynamical quantity available for analysis is the speed of the
flow head as recorded in some instances by cable breaking
at the bottom of the ocean, but there is some uncertainty as
to these breakings correspond to the passage of a slide, a
debris flow or (in the distal part) to a turbidity current
[Heezen and Ewing, 1999; Hsu, 1989]. Hydroplaning
blocks as observed in front of some submarine debris flows
and slides have size of a few tens up to a few hundreds of
meters across. They are separated from the mean body by
several hundreds of meters up to kilometers. These blocks
have probably hydroplaned down to the final location, in
some cases apparently without direct contact with the bed.

[55] In passing from the laboratory to the field scale,
some of the physical quantities must be extrapolated by
several orders of magnitude. Table 1 collects the typical
values in the laboratory and field scale, including also glide
blocks, which can be considered of intermediate size (small
field scale).

[56] Note that the linear quantities (length of the hydro-
planing sediment, thickness and width) are about three to
cfive orders of magnitude larger while the rheological
properties (yield stress and viscosity) are about two to four
orders of magnitude greater. The value of the yield stress is
constrained by the thickness of the deposits along the Norwegian margin. Unfortunately, the final deposit thickness has a simple relation with the yield stress only if the debris flow has not hydroplaned. However, there are indications that the debris flows at the Isfjorden fan in the Barents Sea, which provided the present estimates, have not hydroplaned [Marr et al., 2002]. More direct measurements of yield stress for compacted clay are needed.

The viscosity of the mud is a more uncertain parameter. Laboratory tests indicate that the viscosity of remoulded clay is approximately proportional to the yield stress. According to laboratory measurements by Locat and Demers [1988], values of 50 Pas should be used for the large field scale. Following Marr et al. [2001], we increased the viscosity up to 300 Pas in our simulations. The exact value of the viscosity is not very important, however, because the ratio of the yield to viscous forces, known as the Bingham number, is large, $\frac{\tau_y}{\mu v} \approx 1800$, $\tau_y^{0.13} \approx 500–1000$, where a parametrization of the relationship between viscosity and yield stress from the data by Locat and Demers [1988] has been used. We find a negligible effect with a change of the viscosity by two orders of magnitude.

The viscosity of water in the hydroplaning layer will be larger than for pure liquid. General equations stemming from Einstein’s equations for very dilute suspensions [Einstein, 1906] and extended to more concentrated slurries [Coussot, 1997] would provide a reasonable approximation for water viscosity if the solid concentration and composition were known. Unfortunately, this is not the case. In our simulations, we thus followed the work by Mohrig et al. [1998] and for both small and large-scale debris flow set a viscosity about twenty times larger than for pure water, which seems reasonable for laboratory experiments.

As previously described, we start a typical simulation with a parabolic shape for the initial deposit. The initial height of the parabola is 300 m and the length is 20 km. The mud moves down an incline of approximately one degree slope and after 100 km a slope break occurs. Figure 6 shows the results obtained with our program (see the figure caption for a complete list of input values). In the beginning, the mud accelerates without hydroplaning, until the front reaches the critical velocity and starts to hydroplane. More water intrudes under the mud, and the front of the material is accelerated further. This causes the front to be almost detached from the rest of the body. The tail of the deposit does not follow accordingly. Note that hydroplaning determines a much longer run-out of the deposit (the final deposit profile obtained with the nonhydroplaning BING model is also reported for comparison). This is a rather general

![Figure 6](image_url)

Figure 6. Results from the hydroplaning model. A deposit with a parabolic initial shape is accelerated in the gravity field on a 1 degree slope, which becomes 0.17 degrees after 100 km. The front accelerates faster than the tail, and the flow stops when the water is expelled. The initial water layer thickness when the front starts hydroplaning is 2 m (at the front) and the mud is considered to hydroplane if the water layer is at least 20 cm thick (which corresponds to a quite smooth bed). The mud viscosity is 300 Pa s, the yield stress is 15 kPa, and the water viscosity is 0.03 Pa s, about 20 times the value for pure water. The corresponding final deposit without hydroplaning (BING model) is also shown.
finding of the present model; hydroplaning usually produces an acceleration of the head with consequent larger run-outs. Figure 7 reports the final deposit profiles from different simulations, with and without hydroplaning. More results have been obtained by reducing the yield stress from 15 kPa to 10 kPa. Note that while the run-out distance of a nonhydroplaning body depends strongly on the yield stress of the material, hydroplaning produces a more material-independent mechanism of sediment transport. The fluctuations in the deposit profiles obtained with hydroplaning are more directly related to the shape and thickness as well as position of the front of the body at onset of hydroplaning.

Figure 7. Result of the hydroplaning model. The same as in the previous plot for different yield stresses. Note that in these simulations the highest yield stress gave the larger run-out. On the average, the run-out decreases with a very large increase of the yield stress, but fluctuations such as the one observed here are possible if the variation in the yield stress is limited.

Figure 8. Calculated velocities as a function of time. The velocity for the front and midpoint of a hydroplaning flow are shown with full, thick line and dotted line, respectively. The critical velocity for hydroplaning is represented by a short-dashed line. The BING calculations are shown (for the front point) by a long dashed line (with drag) and dot-dashed (no drag). The case without drag effects is shown also for the case of BING plus hydroplaning.
high initial velocities because of an excess of height of the sliding material. In fact, the shape and size of the initial deposit and the shape the surface of failure can affect the velocity in the calculations.

Clearly, the triggering mechanism of the debris flow, the shape of the initial deposit and the “feeding mechanisms” of mass into the flow need to be better understood. Note also that without the drag terms unrealistically high velocities are reached. The velocity of a point situated halfway between the head and the tail and the velocity of the head calculated with the BING model, which does not introduce hydroplaning, are also presented. Since the BING model as published by Imran et al. [2001] does not include the drag force, it is interesting to compare the BING with and without drag. The differences are less dramatic than in the case with hydroplaning. In addition, the critical velocity for hydroplaning is shown for a Froude number \( Fr_{cr} = 1 \). This shows that hydroplaning starts within the first couple of minutes.

We finally note that in a hydroplaning sediment, the main forces are the gravity force along the bed and the drag force. When these forces are ideally equal and opposite, the velocity remains constant. From the equation of motion equation (26), it follows that this terminal velocity essentially scales like the square root of the sediment thickness. Since the critical velocity for hydroplaning is also proportional to the square root of the thickness, it follows that the ratio of the typical velocity of hydroplaning debris flow to the critical velocity does not change when going from the small to the large-scale flow. This indicates that the extrapolation from the laboratory to the field-scale debris flows is perhaps less problematic than one would expect. More elaborated distorted scaling relations can be worked out to compare more quantitatively the laboratory and the field scale.

5. Conclusions

In this paper we have presented a model for hydroplaning debris flows. The model gives reasonable estimates for laboratory flows. Applied to the field, it gives run-out distances much greater than predicted by models that do not include hydroplaning, and might thus contribute to explain the large run-out distances observed for several large debris flows. In the model we have neglected some effects which might turn out to be important. In particular, a more comprehensive description of water flow into the lubricating layer, of turbulence and of the forces (drag and lift) on a debris flow will be an important step toward a quantitative model of hydroplaning. The implementation of the model to more sandy flows implies water seepage through the main body and a different rheological relation before hydroplaning takes place, including Coulomb friction between grains. The effect of seeping water might decrease the local yield stress because of increased water pore pressures. Using the more general Herschel-Bulkley rheology rather than a Bingham type may represent an improvement of our model. However, this does not seem to be very important, since hydroplaning does not depend much on the rheology. It is possible that multiphase fluid-dynamical codes might provide a natural framework for simulation of hydroplaning.

We also remark that the model results depend on more variables than its predecessor BING, many of which are not well known. In addition, hydroplaning is a delicate, highly nonlinear effect, where small perturbations can potentially change radically the flow pattern. For example, laboratory experiments of unconfined debris flows show that under the same conditions, hydroplaning may start or not, according to unpredictable triggering mechanisms. Even the same debris flow in an unconfined experiment may develop parts which hydroplane and parts which do not. In view of these problems, some of which are due to limitations of the present model, and some are intrinsic to the phenomenon, we cannot expect the model to have the same predictive power as BING.

Finally, submarine slides and debris flows are complex phenomena. In the areas of interest, structures like faults and cracks, ridges, isolated blocks, or steep headwalls appear. The history of the sliding areas can also be extremely complicated and involve multiple phases, remobilization and retrogressive sliding. Clearly, the present work addresses only a part of the whole problem of the evolution of sliding areas on the continental margins. However, we believe that the most important physical effects of hydroplaning debris flows are included in the model.

### Notation

\( \Delta \rho = 1 - \rho_w/\rho_d \) nondimensional reduced density of the mud.

- \( \gamma \) shear rate.
- \( \gamma_r \) reference strain rate.
- \( \mu_d \) viscosity of the mud.
- \( \mu_w \) viscosity of the water slurry.
- \( \theta \) slope angle of bed.
- \( \rho_d \) density of the mud.
- \( \rho_w \) density of water.
- \( \tau_b \) stress in at the bed.
- \( \tau_t \) stress at the top of the mudflow.
- \( \tau_w \) stress at the top of the water layer.
- \( \psi \) yield stress.
- \( \tau_{xx} \) shear component of the stress tensor.
- \( D \) total thickness of the mud layer.
- \( D_s \) thickness of the shear layer.
- \( D_p \) thickness of the plug layer.
- \( D_w \) thickness of water layer.
- \( f_t \) mass flux in the shear layer along the x direction.
- \( J_x \) momentum flux in the shear layer along the x direction.
- \( P \) pressure.
- \( g \) gravitational acceleration.
- \( g_x \) x component of g.
- \( g_y \) y component of g.
- \( u \) water velocity along the x direction.
- \( v \) water velocity along the y direction.
- \( u_p \) velocity of the plug layer.
- \( u_w \) velocity at the top of the water layer.
- \( x \) position along down slope.
- \( y \) position normal to the slope.

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References


Figure 1. The pressure [kPa] around a rectangular object with height 10 cm moving 5 mm above the bed with a velocity of 1 m/s. The velocity of water relative to the object is indicated by arrows.

Figure 2. The pressure [kPa] and water velocity field around an object shaped as a debris flow head moving with a velocity of 1 m/s. The shape of subaqueous debris flows seen in experiments has been used.