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HYDRAULIC MODELING OF UNSTEADY DEBRIS-FLOW SURGES WITH SOLID-FLUID INTERACTIONS

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ABSTRACT

Interactions of solid and fluid constituents produce the unique style of motion that typifies debris flows. To simulate this motion, a new hydraulic model represents debris flows as deforming masses of granular solids variably liquefied by viscous pore fluid. The momentum equation of the model describes how internal and boundary forces change as coarse-grained surge heads dominated by grain-contact friction grade into muddy debris-flow bodies more strongly influenced by fluid viscosity and pressure. Scaling analysis reveals that pore-pressure variations can cause flow resistance in surge heads to surpass that in debris-flow bodies by orders of magnitude. Numerical solutions of the coupled momentum and continuity equations provide good predictions of unsteady, nonuniform motion of experimental debris flows from initiation through deposition.

INTRODUCTION

Attempts to explain or predict the behavior of debris flows must acknowledge that debris-flow motion is seldom steady or uniform. Several mathematical models of unsteady, nonuniform debris-flow motion have been developed, and all employ depth-averaged ("hydraulic") equations of motion similar to the Saint-Venant equations used in water-flood routing. Hydraulic models differ chiefly in how they represent resisting forces, initial conditions, and boundary conditions. Most models use resistance formulations for homogenous Newtonian fluids, Bingham viscoplastics, or Bagnold grain flows, and use upstream input hydrographs to represent debris-flow initiation and no-slip boundaries to represent interaction with the bed. In contrast, observations indicate that most debris flows initiate when static, water-laden sediment begins sliding frictionally, liquefies, and continues downstream as a heterogeneous surge with a high-friction flow front and more-fluid body (Costa and Williams, 1984; Johnson, 1984). Debris flows commonly stop when the flow front loses momentum and impounds more-fluid debris

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behind it. Realistic models should represent these phenomena as integral parts of the debris-flow process.

This paper describes a realistic hydraulic model that melds principles of soil mechanics and fluid mechanics to simulate debris-flow motion from initiation through deposition. The model is motivated by experimental data, which show that during slope failure, pore-fluid pressures in the body of debris flows can rise to levels nearly sufficient to liquefy the sediment; pore pressures then remain elevated until post-depositional sediment consolidation occurs (Iverson *et al.*, 1997; Iverson, 1997; Major, 1996; Major *et al.*, 1997; Reid *et al.*, 1997). In contrast, the flow front lacks positive pore pressures and resists motion as the low-friction flow body pushes it from behind. The hydraulic model includes the effects of grain-grain and grain-fluid interactions in equations that describe downslope motion of a deforming mass of Coulomb frictional material variably liquefied by viscous pore fluid. In the limit of complete liquefaction, the model reduces to the viscous debris-flow model of Hunt (1994). In the limit of vanishing fluid viscosity and pressure, the model reduces to the granular surge model of Savage and Hutter (1989, 1991). In the initial-condition limit of zero flow velocity, the model reduces to a limit-equilibrium model of slope failure. In the limit of deposition, the model demonstrates how internal and basal Coulomb friction concentrated in the flow front can halt downstream debris-flow motion. Use of the model to predict flow speeds and depths measured in large-scale flume experiments entails no calibration of parameters, yet numerical predictions of the model nearly match experimental data.

CONSTRAINTS FROM EXPERIMENTAL DATA

Several investigators have reported data on the distribution of pore fluid and pore-fluid pressure in debris flows. Takahashi (1991, p. 10) summarized measurements beneath a debris flow at Yake Dake, Japan. The data indicated low pore-fluid pressure at the bouldery debris-flow front but high pore-fluid pressure in the muddy debris behind the front. In some places the fluid pressure apparently reached levels sufficient to liquefy the sediment. Coleman (1995), building upon work of Davies (1990), photographed propagating surges in miniature experimental debris flows composed of coal grains and dilute wall-paper paste. The contrasting albedo of the black grains and white pore fluid revealed that the fronts of surges remained unsaturated while the debris behind the fronts became thoroughly mixed and saturated with pore fluid.

Large-scale experiments at the USGS debris-flow flume yield detailed, replicable data on pore-fluid pressures in debris flows composed of ~10 m^3 of sand, gravel, and water, with only a few percent silt and clay (Iverson *et al.*,1997; Iverson, 1997; Major, 1996; Major *et al.*, 1997; Reid *et al.*, 1997). Simultaneous, high-speed measurements of flow depth, basal total normal stress and basal pore-fluid pressure indicate that these quantities vary rapidly as debris flows move past a fixed channel cross section. Total normal stress increases in proportion to the flow depth (except during brief intervals when significant velocity normal to the bed occurs), but basal fluid pressure increases only *after* surge fronts have passed (Figure 1). Basal fluid pressures behind surge fronts attain levels nearly sufficient to liquefy the sediment. Realistic models of debris flows should account for longitudinal variations of flow resistance that result from nonuniform distributions of pore pressure, as shown in Figure 1.

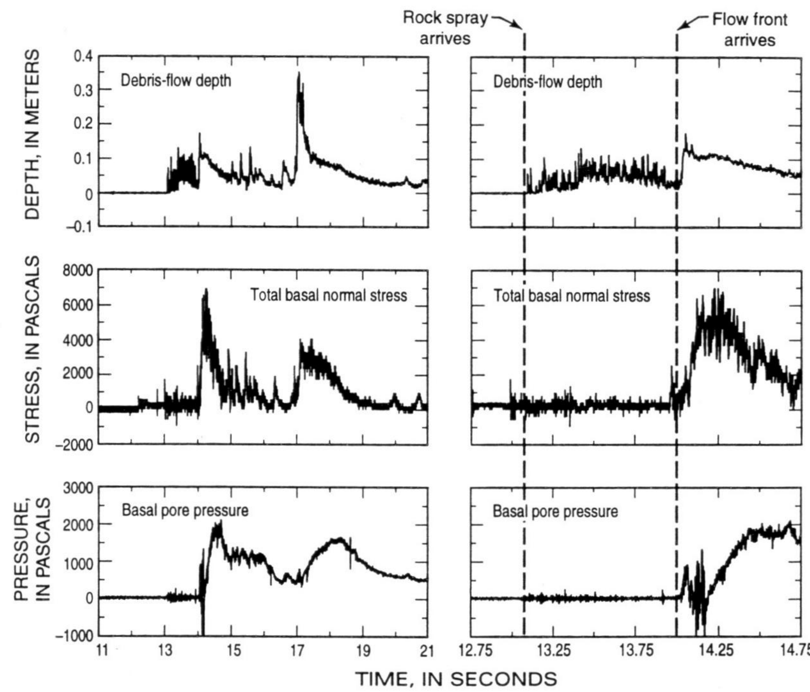

Figure 1. Flow depth and basal stresses recorded as an experimental debris flow of 9 m^3 of water-saturated loamy sand and gravel passed a cross section 67 m downslope form the gate at the head of the USGS debris-flow flume, 31 August, 1994. Left diagram shows data for the entire flow duration. Right diagram shows expanded data for the two-second interval when the flow front passed the cross section.

HYDRAULIC MODEL

The basic equations of the new hydraulic model are derived from continuum mixture theory by depth averaging and summing the pertinent two-phase flow equations, and by making other appropriate assumptions (Iverson, 1997). Key steps in the derivation assume that the debris has a fixed total mass and bulk density, that motion occurs in only the downslope direction, that the velocity of pore fluid differs little from that of adjacent solid grains, and that flow resistance results from a combination of internal grain-contact friction, bed friction, and pore-fluid viscosity. The viscosity and density of the fluid phase in many debris flows exceed those of clear water because the fluid effectively consists of water plus suspended sediment finer than sand (Iverson, 1997).

Coarser sediment grains can interact through both sustained frictional contacts and brief, inelastic collisions, but collisions can be neglected if the Bagnold number and Savage number characterizing the flow are suitably small (Iverson, 1997). For large values of the Bagnold and Savage numbers, effects of grain collisions should be included by evaluating interactions between viscous, frictional, and collisional effects and not by a naive summation of these influences (*cf.* Iverson and Denlinger, 1987; Johnson and Jackson, 1987). Here, as a starting point, I neglect grain collisions and focus on interactions between grain friction and viscous fluid flow.

With the caveats described above, one-dimensional equations for conservation of mass and linear momentum in debris flows may be written as (*cf.* Iverson, 1997)

$$\frac{\partial h}{\partial t} + \frac{\partial (hv)}{\partial x} = 0 \tag{1a}$$

$$\begin{aligned} \frac{\partial v}{\partial t} + v\frac{\partial v}{\partial x} = {} & g[\sin\theta - (\cos\theta - \frac{p_{bed}}{\rho g h})\tan\phi_{bed}] \\ & - \upsilon_f (m+2)\frac{\mu}{\rho}\frac{v}{h^2} \\ & - k_{a/p}\left[g\cos\theta\frac{\partial h}{\partial x} - \frac{1}{\rho}\frac{\partial p_{bed}}{\partial x}\right] - \frac{1}{\rho}\frac{\partial p_{bed}}{\partial x} \end{aligned} \tag{1b}$$

in which the dependent variables are h, the flow depth measured normal to the bed, and v, the depth-averaged downslope flow velocity (Figure 2). The independent variables are x, the downslope distance parallel to the bed, and t, time. The model parameters are g, the magnitude of gravitational acceleration, θ, the angle of inclination of the bed slope, ρ, the bulk density of the debris-flow mixture, p_{bed}, the pore-fluid pressure at the base of the flow, ϕ_{bed}, the friction angle of the debris-flow sediment on the channel bed, υ_f, the fluid volume fraction, μ, the fluid viscosity, m, a descriptor of the vertical velocity profile shape, and $k_{a/p}$, the lateral grain-pressure coefficient. The appropriate expression for $k_{a/p}$ was first presented without derivation by Savage and Hutter (1989):

$$k_{a/p} = 2\frac{1 \mp [1 - \cos^2\phi_{int}(1 + \tan^2\phi_{bed})]^{1/2}}{\cos^2\phi_{int}} - 1 \tag{1c}$$

and a derivation was provided by Iverson (1997). In (1c) ϕ_{int} is the internal friction angle of the granular solids, and the "-" in "$\mp$" applies for k_a (active, extending flow case), whereas the "+" sign applies for k_p (passive, compressing flow case). Equation (1c) results from applying classical Rankine earth-pressure theory (*e.g.*, Lambe and Whitman, 1979) to situations in which a Coulomb mass simultaneously fails internally and along the bed. Table 1 summarizes typical values of k_a, k_p, and other model parameters.

Table 1. Examples of typical values of parameters in the debris-flow model given by (1a,b,c).

Type or Section of Debris Flow	θ (deg.)	ρ (kg/m^3)	μ (Pa-s)	υ_f --	p_{bed} (Pa)	m --	ϕ_{int} (deg.)	ϕ_{bed} (deg.)	k_a --	k_p --
bouldery flow front	0-45	2000	0.001	0	0	--	45	25	0.50	5.5
	0-45	2000	0.001	0	0	--	40	30	0.82	4.0
	0-45	2000	0.001	0	0	--	35	30	1.01	2.9
	0-45	2000	0.001	0	0	--	30	30	1.67	1.67
flow body with watery pore fluid	0-45	2100	0 001	0.3	10,000h	1	40	30	0.82	4.0
	0-45	2100	0.001	0.3	10,000h	1	35	30	1.01	2.9
	0-45	2100	0.001	0.3	5000h	1	35	30	1.01	2.9
flow body with muddy Pore fluid	0-45	2200	0.01	0.3	10,000h	1	40	30	0.82	4.0
	0-45	2200	0.01	0.3	10,000h	1	35	30	1.01	2.9
	0-45	2200	0.01	0.3	5000h	1	35	30	1.01	2.9

Significance of Right-Hand Side Terms

Terms in (1a) and on the left-hand side of (1b) have well-known meanings explained in standard references (*e.g.*, Henderson, 1966), but terms on the right-hand side (RHS) of (1b) merit explanation. The first RHS term expresses the balance of forces used in static stability analyses of infinite slopes (Iverson, 1992); that is, it expresses the sum of the gravitational driving force (per unit mass) $g\sin\theta$ and the resistance due to Coulomb bed friction in a mass of uniform thickness. Bed friction is modified by the effects of basal pore-fluid pressure, p_{bed}. If $p_{bed} = \rho g h \cos\theta$, pore pressure balances the total normal force on the bed and liquefies the overlying sediment mass, and flow resistance due to Coulomb bed friction is zero.

The second RHS term in (1b) describes resisting force (per unit mass) due to shearing of the viscous pore fluid. If $\upsilon_f = 1$ and $m = 1$, this term reduces to the standard depth-averaged form for laminar flow of a linearly viscous fluid down an inclined plane (*e.g.*, Hunt, 1994). Generally, $0.2 < \upsilon_f < 0.4$ in debris flows. Values $m \neq 1$ indicate that the vertical velocity profile of the pore fluid is not parabolic, a condition that may result from drag of adjacent granular solids or from turbulence. If the velocity profile is more sharply curved than a parabolic profile, $m > 1$ applies, whereas $m < 1$ applies if the velocity profile is less sharply curved that a parabola.

A more general, empirical form of the second RHS term can be written as $-g(n^2 v^2 h^{-4/3})$, in which n is the Manning flow-resistance coefficient. This form is preferred by many authors owing to its routine use for describing turbulent open-channel flow and its adaptability to various flow-resistance models (*e.g.*, Macedonio and Pareschi, 1992). However, in models that aim to explain debris-flow dynamics, I believe it is undesirable to include a Manning flow-resistance term, because n amalgamates the effects of bed resistance and internal resistance and thereby camouflages key physical issues.

The third RHS term in (1b) represents the most significant feature of the model. The term describes the longitudinal force (per unit mass) due to longitudinal variations in flow depth and pore pressure. Unlike comparable terms in models that use Bingham, Bagnold, or Newtonian rheologies, the third RHS term in (1b) accounts for solid-fluid interactions and allows assessment of dynamic interactions between liquefied debris-flow bodies and high-friction debris-flow fronts. Fluid effects are represented completely by p_{bed}, because hydraulic model assumptions require that pore-fluid pressure varies linearly (but not necessarily hydrostatically) with depth, ranging from 0 at the surface to a maximum of p_{bed}. If $p_{bed}=0$, longitudinal forces are transmitted by grain-contact friction alone, and the third RHS term reduces to $-\cos\theta\,[k_{a/p}g(\partial h/\partial x)]$, which describes lateral forces in a dry, failing Coulomb material. As $p_{bed}\rightarrow\rho g h\cos\theta$, the lateral fluid-pressure force increases and the lateral grain-contact force decreases in proportion. One special case assumes a pore-pressure distribution for steady, uniform, slope-parallel flow (given by $p_{bed}=\rho_f g h\cos\theta$ for debris flows with pore-fluid density ρ_f and thickness h), which is the pore-pressure distribution assumed in Takahashi's (*e.g.*, 1991) well-known debris-flow model. A more instructive special case represents complete liquefaction, given by $p_{bed}=\rho g h\cos\theta$. With complete liquefaction the third RHS term reduces to $-\cos\theta\,[g(\partial h/\partial x)]$, which describes lateral forces in a viscous fluid flow.

The critical importance of the third RHS term in (1b) results from the great change of behavior that occurs as p_{bed} ranges from 0 to $\rho g h\cos\theta$. If $p_{bed}=\rho g h\cos\theta$ and the sediment mass behaves like a liquid, normal stresses are isotropic, equal to the static pressure, and independent of the local style of deformation. If $p_{bed}=0$ and the debris behaves like a Coulomb solid, normal stresses are anisotropic, and the longitudinal normal force depends strongly on whether the sediment mass is locally extending $(\partial v/\partial x>0)$ or compressing $(\partial v/\partial x<0)$ as it deforms and moves downslope. For example, in a typical case with $\phi_{int}=40°$ and $\phi_{bed}=30°$, (1c) indicates that values of the active (extending) and passive (compressing) grain-pressure coefficients are $k_a=0.82$ and $k_p=4.0$, respectively (Table 1). This means that longitudinal forces in regions of extending flow will be 18% less than in a liquid of density ρ, but longitudinal forces in regions of compressing flow will be four times greater than in a liquid. Consequently, the model predicts that strong local gradients in the longitudinal normal force can occur for two reasons: either the style of deformation changes locally from extending to compressing, or the pore pressure varies locally from low to high. Thus, depending on the local deformation style and pore-pressure distribution, the model represented by (1a,b,c) can represent unsteady flow behavior that ranges from that of a granular surge, as modeled by Savage and Hutter (1989, 1991), to that of a Newtonian fluid surge, as modeled by Hunt (1994). Furthermore, the front of a fully developed debris flow may act like a compressing granular solid and support high lateral stresses, while the trailing flow acts more like a fluid. More than any other factor, this simultaneous expression of fluid and solid behavior gives debris flows their unique attributes.

Persistence of High Pore Pressure

Owing to the importance of solid-fluid interactions in debris flows, it is significant that the hydraulic model described by (1a,b,c) treats the basal pore-fluid pressure p_{bed} as a parameter rather than a dependent variable. Although values of p_{bed} are constrained by experimental data such as those in Figure 1 and are not arbitrarily adjustable, values of p_{bed} are not predicted by the model. This limitation has great theoretical significance (although pragmatically it affects model performance rather little). Soil consolidation theory might seem to provide an obvious tool for prediction of p_{bed}, and a consolidation model could be appended to (1a,b,c) -- just as Hutchinson (1986) appended a consolidation model to a rigid-body landslide model. However, key elements of consolidation theory -- a compressible sediment-water mixture and a flux of solids normal to the bed -- are incompatible with the depth-averaged hydraulic equations of motion. Although the hydraulic model (1a,b,c) and consolidation models can each be derived from mixture theory (Iverson, 1997), they cannot be derived using a consistent set of assumptions. Therefore, rather than append a consolidation model to (1a,b,c) in *ad hoc* fashion, I adopt a simpler approach and evaluate only the time scales of consolidation and debris-flow motion. Comparison of these time scales confirms that it is commonly appropriate to treat the pore pressure p_{bed} as a parameter that remains constant within individual volume elements for the duration of a debris flow.

The time scale T_c for debris-flow consolidation (or diffusion of pore pressures that exceed hydrostatic values) may be estimated from

$$T_c = \frac{h^2 \mu C}{k} \tag{2}$$

in which C is the compressibility and k is the intrinsic hydraulic permeability of the aggregate debris-flow solids (Iverson, 1997). Values of k for debris-flow sediments commonly range from about 10^{-9} to 10^{-14} m^2; values of C commonly range from about 10^{-4} to 10^{-5} Pa^{-1}; and a positive correlation exists between k and C (Major, 1996; Major *et al.*, 1997; Iverson, 1997). As consolidation proceeds, reductions in k and C tend to compensate for each other and keep the value of T_c roughly constant. For a debris flow with thickness $h = 1$ m and typical values of μ (Table 1), estimates of T_c range from about 10^3 to 10^7 s — about 20 minutes to 100 days. Thus, liquefied parts of debris-flow masses 1 m thick can be expected to maintain high pore pressures and a fluid-like consistency over time scales that typically exceed the duration of debris flows, and thicker debris flows will maintain high pore pressure for longer times. As a first approximation, therefore, pore pressure can be treated as a spatially varying parameter rather than as a dependent variable in flow-dynamics models. Major (1996) and Iverson (1997) provide detailed calculations and data that support this conclusion.

NORMALIZATION AND ANALYSIS OF THE MODEL EQUATIONS

Useful inferences can be drawn from normalized versions of the equations of motion (1a, 1b). To normalize the equations, I employ the dimensionless variables

$$x^* = x/L \qquad t^* = t/\sqrt{L/g} \qquad v^* = v/\sqrt{gL} \qquad h^* = h/H \tag{3a,b,c,d}$$

$$p^* = \frac{p_{bed}}{\rho g H \cos\theta} \qquad \mu^* = \frac{\mu}{\rho g H (H/\sqrt{gL})} \tag{3e,f}$$

Two length scales (L and H) appear in (3a-f) and denote the characteristic length and height of a debris-flow surge (Figure 2). Use of the length scale L in the definitions of dimensionless time t^* and velocity v^* differs from the convention in open-channel hydraulics, in which the length scale H is used. The rationale for using L derives from the observation that debris flows have a finite length (L) and originate on steep slopes, where the typical velocity is that of free fall ($\sqrt{gL}$) rather than surface-wave propagation ($\sqrt{gH}$) (*cf.* Savage and Hutter, 1989). Identification of the two length scales H and L in (3a-f) leads to definition of the debris-flow aspect ratio, $\epsilon = H/L$ (Figure 2), which typically may be regarded as a small parameter, $\epsilon \ll 1$.

Substitution of (3a-f) in (1a,b) yields normalized equations of motion

$$\frac{\partial h^*}{\partial t^*} + \frac{\partial (h^* v^*)}{\partial x^*} = 0 \tag{4a}$$

$$\begin{aligned}\frac{\partial v^*}{\partial t^*} + v^* \frac{\partial v^*}{\partial x^*} = {} & \sin\theta - \left(1 - \frac{p^*}{h^*}\right)\cos\theta \tan\phi_{bed} \\ & - \upsilon_f (m+2)\mu^* \frac{v^*}{h^{*2}} \\ & - \epsilon \cos\theta \frac{\partial}{\partial x^*}\left[k_{a/p}(h^* - p^*) + p^*\right]\end{aligned} \tag{4b}$$

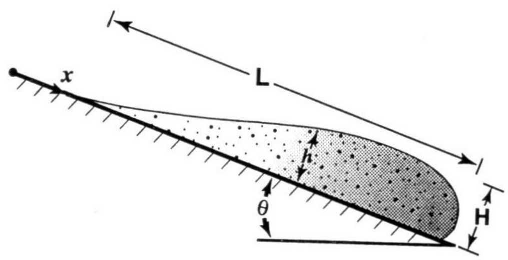

Figure 2. Schematic diagram of surge with definitions of geometric parameters.

in which the relative importance of different terms can be evaluated. Terms of greatest interest in this regard are those on the RHS of (4b). The first RHS term is of order 1, although as $p^* \to h^*$ basal sliding friction diminishes and driving stress dominates the term. The second (viscous) RHS term is of order $m\upsilon_f\mu^*$, and the third RHS term is of order ϵ. If m is less than about 8 (probably true in most all cases), the product $m\upsilon_f$ is of order 1, and the relative importance of the second and third RHS terms depends on the relative magnitudes of the dimensionless viscosity μ^* and the aspect ratio ϵ $(=H/L)$.

Observations and experiments (Takahashi, 1991, p.6; Iverson, 1997) indicate that the aspect ratio ϵ of debris flows commonly ranges from about 0.01 to 0.001 and probably differs little with scale; values of μ^* appear to be less tightly constrained, but typical values can be estimated for a range of debris-flow compositions and sizes (*cf.* Table 1). For example, for a small debris flow in which the pore fluid is water (such as the 10 m^3 flows at the USGS debris-flow flume), $\mu^* \sim 10^{-4}$ is calculated from the representative values $\rho \sim 2000\,\mathrm{kg/m^3}$, $g \sim 10\,\mathrm{m/s^2}$, $\mu \sim 0.001$ Pa-s, $H \sim 0.1$ m, and $L \sim 100$ m. For a much larger field-scale debris flow in which the pore fluid is muddy water, the values $\rho \sim 2000\,\mathrm{kg/m^3}$, $g \sim 10\,\mathrm{m/s^2}$, $\mu \sim 0.01$ Pa-s, $H \sim 10$ m, and $L \sim 10{,}000$ m yield $\mu^* \sim 10^{-6}$. These values of μ^* demonstrate that $\mu^* < \epsilon$, and that μ^*/ϵ generally decreases as debris-flow size increases. Thus in most field-scale debris flows of practical significance, the third (longitudinal stress-gradient) term in (4B) exceeds the viscous term. This is particularly true at the head of debris-flow surges, where significant gradients in flow depth $(\partial h^*/\partial x^*)$ and pore pressure $(\partial p^*/\partial x^*)$ occur. In miniature flows of fine-grained debris (*e.g.*, those in many laboratory experiments) viscous resistance may be more important, and in this respect such experiments may poorly simulate natural events.

NUMERICAL PREDICTION AND COMPARISON WITH DATA

Numerical solutions of the normalized equations of motion (4a,b) use the Lagrangian finite-difference scheme of Savage and Hutter (1989, 1991). Details of initial and boundary conditions and computational implementation are described by Iverson (1997). Here, for brevity, I omit these details and present only comparisons of numerical calculations with data from experiments at the USGS debris-flow flume.

Parameter values used in calculations were determined from independent measurements described by Iverson (1997): $\phi_{bed} = 28°$, $\phi_{int} = 42°$, $\theta = 31°$. The only other parameters that appear in (4a,b) are μ^*, p^*, and ϵ. As a first and simplest approximation, and on the basis of the analyses described in the previous section, the calculations assume $\mu^* = 0$. Thus viscous shear resistance is neglected. Initially the shape of the static sediment heap at the head of the flume establishes $\epsilon \sim 0.3$, but the value of ϵ decreases and tends toward the range $0.01 < \epsilon < 0.001$ after the sediment is released and the surge elongates. Values of p^* are determined by measurements of basal pore-fluid pressures (*e.g.*, Figure 1). As a simple approximation to the rather complicated type of pore-pressure distribution shown in Figure 1, the calculations assume that the leading edge of the debris flow (5% of its length) has $p^* = 0$; over the next 35% of the debris-flow length, the pore pressure increases linearly from 0 to $p^* = 0.9$. The trailing

60% of the debris flow maintains $p^* = 0.9$, a value that denotes pore pressures 90% as large as those required for complete liquefaction.

Figure 3 illustrates comparisons between numerical predictions and debris-flow hydrographs measured during a flume experiment. Replicate experiments yielded similar results. Although the data reveal complications such as roll waves (*cf.* Schonfeld *et al.*, 1997), the model predicts the arrival time and overall shape of the debris-flow surge reasonably well.

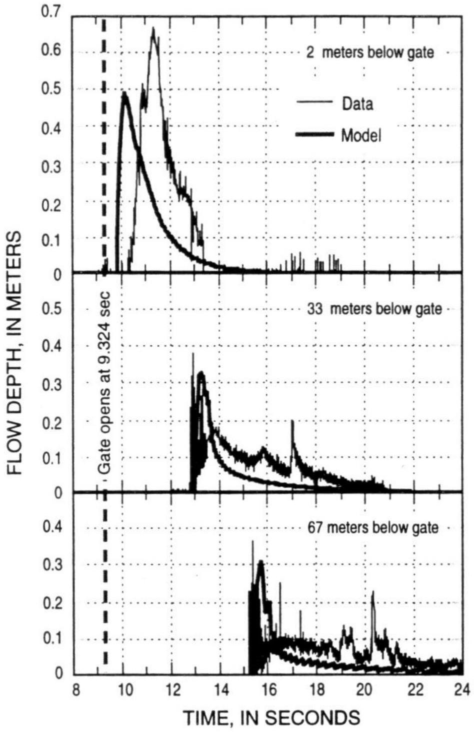

Figure 3. Comparison of model predictions and measurements of debris-flow depths and travel times. Data were obtained at three cross sections during an experiment at the USGS debris-flow flume, 24 July, 1995. The flow consisted of 9.4 m^3 of water-saturated sand and gravel. Detailed records of debris-flow depths at cross sections 33 m and 67 m downslope from the sediment-release gate were acquired with high-precision laser ranging devices, and the less-detailed record 2 m downslope was acquired with an ultrasonic transceiver (after Iverson, 1997).

CONCLUSION

A hydraulic model that emphasizes interactions of pore-fluid pressures and grain-contact friction adequately predicts the behavior of experimental debris flows at the USGS flume. The model equations help illuminate the relative importance of resisting forces that can influence debris-flow motion. The simplest approximation assumes that resistance is that of a rigid Coulomb body of uniform thickness. This resistance is mediated by the effects of basal pore-fluid pressure, and it diminishes as pore pressures approach liquefaction magnitudes. Significant modification of rigid-body resistance results from longitudinal stress gradients due to nonuniform flow depths and pore pressures. A generally smaller modification results from viscous shear resistance. However, where debris-flow sediment is completely liquefied, overall resistance is greatly reduced and viscous shear resistance may dominate. Although complete liquefaction can occur at least locally in debris-flow bodies, no evidence indicates that it occurs in the heads of debris-flow surges. Coarse-grained surge heads may thus dominate flow resistance, and the rheology of finer-grained slurries that trail surge heads may influence debris-flow dynamics relatively little.

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Abstract:
This proceedings, *Debris-Flow Hazards Mitigation: Mechanics, Prediction, and Assessment*, contains papers presented at the First International Conference held in San Francisco, California, August 7-9, 1997. The papers covered a variety of topics ranging from debris-flow mechanics to debris-flow hazards prediction and assessment. In addition to the peer-reviewed papers, this volume includes two invited papers. One presents an overview on the geoscience and geotechnical engineering aspects of debris flow while the other provides an overview of hydroscience and hydrotechnical engineering aspects.

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Cover photos: Four photos of a debris flow over a dam at Kamikamihorizawa Creek of Mount Yakedake, Japan (photographed on August 3, 1976, in sequence of 2 seconds apart). Photos courtesy of Prof. Hiroshi Suwa of Kyoto University.