9 Physical Geomorphology of Debris Flows

John E. Costa[1]

Introduction

debris avalanche = fast debris flow

Hazard Justification

Debris flows claim hundreds of lives and cause millions of dollars of property damage throughout the world each year. In Japan alone, an average 90 lives are lost annually from debris flows (Takahashi 1981). In 1970 a debris avalanche (a rapidly moving form of debris flow) triggered by an earthquake, completely destroyed the city of Yungay, Peru, killing an estimated 17,000 people and burying the whole city under 5 m of mud and debris (Plafker and Erickson 1978). Some countries with chronic losses from debris flows include Japan (Okuda et al. 1980); United States (Committee on Methodologies for Predicting Mudflow Areas, 1982; Scott 1972; Cummans 1981; Scott 1971; Flaccus 1958; Williams and Guy 1973; Woolley 1946; Morton and Campbell 1974); Indonesia (Scrivenor 1929); Tanzania (Temple and Rapp 1972); Scandinavia (Rapp and Strömquist 1976); Costa Rica (Waldron 1967); China (Li and Luo 1981; Chinese Society of Hydraulic Engineering 1980); Brazil (Jones 1973); Ireland (Prior et al. 1968); Romania (Balteanu 1976); India (Starkel 1972); Bangladesh (Wasson 1978); New Zealand (Selby 1967; Pierson 1980a, b); and the Soviet Union (Gol'din and Lyubashevskiy 1966; Niyazov and Degovets 1975; Gagoshidze 1969).

Terms

Debris flows are a gravity-induced mass movement intermediate between landsliding and waterflooding, with mechanical characteristics different from either of these processes (Johnson 1970). A debris flow is a form of rapid mass movement of a body of granular solids, water, and air (Varnes 1978). Flow properties vary with water and clay content, and sediment size and sorting. In this report, debris flows are broadly interpreted and include mudflows (debris containing mostly sand, silt, and clay-sized particles), lahars (volcanic mudflows), tillflows, or debris in active or stagnant ice, reworked and transported downslope by gravity (flowtills of Hartshorn 1959); and debris avalanches [a variety of very rapid to extremely rapid debris flows (Varnes 1978)]. Pyroclastic flows from volcanic eruptions, grain flows [rapid mass movements involving non-cohesive materials where grain-to-grain interaction is the dominant particle-support mechanism (Middleton and Hampton 1976; Lowe 1976)] and sturzstroms [very rapidly moving and relatively dry landslides derived from large fallen rock masses which disintegrate in the initial stages of movement (Hsu 1975)] are not included.

1 Department of Geography, University of Denver, USA
Present address: U.S. Geological Survey, MS 413, Denver Federal Center, Lakewood, CO 80225, USA

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Origins and Types of Debris Flows

Debris flows originate when poorly sorted rock and soil debris are mobilized from hillslopes and channels by the addition of moisture. Prerequisite conditions for most debris flows include an abundant source of unconsolidated fine-grained rock and soil debris, steep slopes, a large but intermittent source of moisture, and sparse vegetation. The most common moisture sources are rainfall, snowmelt, and to a lesser extent, glacial outburst floods and rapid drainage of volcanic crater lakes. These environmental conditions are typically found in mountainous areas in arid, semiarid, arctic and humid regions. The importance of debris flows in arid and semiarid regions has been recognized by Blackwelder (1928); Hooke (1967); Beaty (1974); in humid-temperate regions by Williams and Guy (1973); Selby (1974); Pierson (1977, 1980a); Johnson and Rahn (1970); and in arctic regions by Winder (1965); Broscoe and Thomson (1969); Rapp and Nyberg (1981); and Lawson (1982).

Progressively smaller and steeper basins have the potential to transport an increasingly larger percentage of eroded material by mass-wasting processes such as debris flows. This is because 1. rainstorms drop proportionally larger volumes of water on smaller basins; 2. smaller basins are usually the highest, where snowpacks accumulate and can melt rapidly in the spring, or are the highest parts of drainage basins draining volcanic slopes; and 3. hillsides in smaller basins have steep slopes (commonly exceeding 30° in mountainous regions), resulting in greater instability of surficial materials. Sufficiently intense precipitation or snowmelt saturates permeable surficial deposits. This increases pore-water pressure and increases the likelihood of slope failures. The most studied mudflows originating from thaw of winter snowpack occur at Wrightwood in the San Gabriel Mountains of Southern California (Morton and Campbell 1974; Johnson 1970; Sharp and Nobles 1953).

Because of sparse rainfall data in most mountain areas, the intensity or duration of precipitation required to mobilize side-slope materials is poorly known. Campbell (1975) determined that a 6.4 mm/hr rainfall intensity in areas where the total seasonal antecedent rainfall has reached 254 mm are the threshold conditions for the initiation of soil slips and debris flows in the Santa Monica Mountains in Southern California. In the San Francisco Bay region, numerous mass movements, including debris flows, occur during storms in which more than 150–200 mm of rain falls in areas where 250–280 mm of rain has already fallen during a rainy season (Nilsen et al. 1976). — threshold

On a basis of 73 observations of rainfall intensity and duration, and resulting slope failures (debris flow-type in which the initial failure is a slide or slump which rapidly disintegrates into a flow) from all over the world, Caine (1980) has defined a limiting threshold for such slope failures. The limiting curve has the form threshold

$$I = 14.82\ D^{-0.39}$$

where I is rainfall intensity in mm/hr and D is duration of rainfall in hours. This relationship is best defined for rainfall durations between 10 minutes and 10 days.

Lahars are volcanic debris flows which originate on the slopes of volcanoes (Neall 1976). They are common occurrences in historic and prehistoric times, and have destroyed more property than any other process associated with volcanoes and have killed thousands of people (Macdonald 1972). The debris of a lahar can be either hot or cold. Lahars can form as a result of (a) rainfall; (b) rapid melting of snow and glaciers on steep side slopes or in craters during eruptions; (c) rapid draining of crater lakes by expulsion or failure of the rim which entrain large amounts of unconsolidated volcanic debris from steep volcanic side slopes; (d) pyroclastic flows that incorporate water from mixing and melting of eroded snow during downslope movement; and (e) movement of water-saturated material down the slope of the volcano set off by earthquakes.

Lahars may also form when volcanic landslide debris temporarily dams streams. Water can overtop the dam, erode it, and sweep the entire deposit downstream as a lahar (Aramaki 1956). Other lahars occur from the sudden failure of the retaining embankment of a crater lake by factors unrelated to contemporaneous volcanic activity. In 1953 part of the natural dam impounding Crater Lake on Mount Ruapehu, New Zealand failed and a large lahar formed which swept away a railroad bridge just a few minutes before a passenger train arrived; 151 lives were lost (O'Shea 1954).

Tillflows are debris flows that originate in sediments on the surface of glaciers and flow laterally onto adjacent lower surfaces as the glacier melts. Flows are initiated by 1. slumping of sediments covering glacial ice; 2. back-wasting of slopes composed of sediment and stagnant glacial ice; and 3. ablation of debris-laden ice (Lawson 1982). The resulting deposits are called "flowtills" (Hartshorn 1958), or "sediment flow deposits" (Lawson 1982). These deposits have particle size distributions and other physical properties similar to adjacent ground moraines. Hartshorn (1958) reports that flowtills contain fewer boulders and more voids than surrounding glacial deposits. Boulton (1968) observed flowtills covering extensive areas of glacier snouts, and deposits up to 5 m thick on and adjacent to Vestspitsbergen glaciers. Debris flows can also occur from erosion and mobilization of sediments by catastrophic drainage of meltwater stored behind and beneath glaciers (Jackson 1979).

Failure Mechanisms

slope/ topo

Most debris flows begin as slope failures on steep (greater than 15°–20°) side slopes from a relatively quick influx of large amounts of water. The mass movements usually originate at the head of swales (small first-order drainages), but about one-third originate on flat and convex (ridgespurs) side slopes (Smith and Hart 1982). The initial failure can be a slide, slump, or topple. The exact mechanism by which slope failures become debris flows is uncertain, but the rapid transformation of planar or rotational slides and slumps into flows either through dilatancy (defined below) and incorporation of additional water, or by liquefaction, has been favored by many investigators (e.g., Campbell 1975; Hampton 1972; Pomeroy 1980; Starkel 1972; Rapp and Strömquist 1976; Temple and Rapp 1972; Johnson and Rahn 1970; Pierson 1977).

process

When a relatively competent, rigid slab or block of soil becomes saturated above a failure surface, pore pressure increases and shear strength decreases. Overlying soils may or may not be saturated.

pore press, process

Pore pressure in soils increases when the rate of deep percolation is slower than the rate of infiltration from melting snow or rainfall. The failure of a mass of debris under high pore pressures and diminished shear strength can cause soil particles to lose coherency and to rework the soil mass thoroughly enough to cause remolding. This causes the debris to change by spontaneous liquefaction from a rigid slab into a viscous fluid, and flow (Terzaghi and Peck 1967, p. 108; Youd 1973).

The transformation from a solid, rigid mass to a viscous fluid can also occur through dilatancy. Dilatancy is an increase in the bulk volume of a soil mass which occurs during deformation accompanying slope failures. It is caused by a change from close-packed structure to open-packed structure, accompanied by an increase in the pore volume. With the incorporation of additional moisture and remolding, the solid mass can become a flowing, viscous fluid. Term

Just where and when this transformation from solid mass to viscous fluid occurs during downslope movement will vary from site to site. Investigation of failure scars, landforms, and deposits in the channelway downslope may indicate the first appearance downslope of distinctive debris flow deposits and landforms. The change from a solid to a viscous fluid also changes the resistance to downslope movement from sliding friction to the viscosity of the flow. This allows newly formed debris flows to accelerate quickly and attain high velocities on steep slopes.

Channel deposits can also be mobilized by runoff and act as sources for debris flows. Beaty (1963, p. 525) reports that debris on canyon floors in the White Mountains was virtually the sole source of sediments for extensive debris flows in 1952. Scott (1971) noted that channel deposits are a major source of debris flow sediments in the San Gabriel Mountains, California. In Glenwood Springs, Colorado, rainfall on steep (25°–40°) side slopes saturates, weakens, and erodes unconsolidated material as small landslides and debris avalanches (Mears 1977). The debris comes to rest at the foot of landslide chutes, and may accumulate to depths of 10 m or more. Stormwater runoff from upper basin areas meets these debris dams, infiltrates the material, and mobilizes the sediment down the channels as debris flows. This process of avalanching, damming, and debris flow formation may be a common mechanism for debris flow genesis in other mountainous areas as well. BULKING

Characteristics of Flowing Debris

There have been relatively few observations of debris flows by trained professionals (Sharp and Nobles 1953; Morton and Campbell 1974; Pierson 1980a; Broscoe and Thomson 1969; Curry 1966; Vinogradov 1969; Wasson 1978; among others), but some quantitative data have been collected during some of the flows. Table 1 summarizes the results of many of the reported analyses and measurements of sampled flows.

Debris flows usually follow pre-existing drainageways, but can move down hillslopes and across unobstructed fan surfaces in almost any direction because flows tend to build their own channels as levees form at the lateral boundaries of the flow. Observed debris flows resemble wet concrete that generally moves downvalley in a

Table 1. Physical properties of sampled debris flows

Location	Velocity (m/s)	Slope (%)	Bulk density (g/cm^3)	Newtonian viscosity (Poise)	%Clay	Depth (m)	Solids (% Wt.)	Shear strength (dn/cm^2)	Reference
Rio Reventado, Costa Rica	2.9–10	4.6–17.4	1.13–1.98	–	1–10	8 –12	20–79	–	Waldron 1967
Hunshui Gully, China	10 –13	–	2.0 –2.3	15– 20	3.6 (<0.005 mm)	3–5	80–85	294–490	Li and Luo 1981
Bullock Creek, New Zealand	2.5– 5.0	10.5	1.95–2.13	2,100– 8,100	4	1.0	77–84	–	Pierson 1981
Pine Creek,[a] Mt. St. Helens, Wa.	10 –31.1	7 –32	1.97–2.03	200– 3,200	–	0.13– 1.5	–	3,900–11,300	Fink et al., 1981
Wrightwood Canyon, Ca. (1969 flow)	0.6– 3.8	9 –31	1.62–2.13	100–60,000	–	1.0	59–86	–	Morton and Campell 1974
Wrightwood Canyon, Ca. (1941 flow)	1.2– 4.4	9 –31	2.4	2,100– 6,000	<5	1.2	79–85	–	Sharp and Nobles, 1953
Lesser Almatinka River, U.S.S.R	4.3–11.1	10 –18	2.0	–	–	2 –10.4	58	–	Niyazov and Degovets 1975
Matanuska Glacier,[b] Alaska	0.001–1.3	2 –47	1.8 –2.6	–	≦3	0.01– 2.0	67–89	$<0.4\times10^6$ to 1.5×10^6	Lawson 1982
Nojiri River, Japan	12.7–13.0	5.8– 9.2	1.81–1.95	–	–	2.3 – 2.4	–	–	Watanabe and Ikeya 1981
Mayflower Gulch, Colorado	2.5	27	2.53	30,000	1.1 (<0.004 mm)	1.5	91	–	Curry 1966
Dragon Creek,[a] Arizona	7.0	5.9	2.0	27,800	–	5.8	80	221	Cooley, et al. 1977

[a] Calculated values from deposits
[b] Type I, II, III flows only

series of waves or surges, with periods ranging from a few seconds to several hours. A debris flow on July 8, 1921 in the Malaya Almatinke River, USSR, had 80 surges (Sokolovsleii 1968, p. 402), and 100 to 200 surges occurred along the Xiaojiang River, China (Li and Luo 1981). Surges originate from the temporary damming and breaching of channels by debris, and damming at constrictions in the channel. The front of debris flow surges are usually higher than trailing portions, and contain the largest boulders being moved. The surges are followed by more fluid, watery, turbulent slurries with unusually high suspended sediment concentrations, but fewer boulders. This more fluid phase continues until the next surge arrives or until debris flow activity ceases (Johnson 1970; Pierson 1980a; Sharp and Nobles 1953). process

The velocity of debris flows varies because of the character of the debris – the size, concentration, and sorting of material, and because of channel geometry including shape, slope, width, and sinuosity. Observed velocities range from 0.5 to about 20 m/s. vel. rate

The results of mechanical analyses of many debris flows indicate that a very small portion of debris flow material consist of silt and clay-sized particles (roughly 10–20%), and the percentage of clay can be surprisingly low, generally no more than a few percent. Curry (1966), Sharp and Nobles (1953), and Lawson (1982) all report less than 3% clay in sampled debris flows. Bull (1964) reports up to 76% clay in debris flow deposits from clay-rich source areas in California. low cohesion

Debris flows can have Newtonian viscosities as much as 1 to 8×10^3 poises, (Table 1), compared with 0.01 poise for pure water at 20° C (Campbell 1975; Pierson 1980a). Table 2 lists some viscosities of other non-Newtonian fluids. viscosity

Bulk densities of debris flows vary widely, and are highly dependent on sampling methods. If two samples are collected from the same debris flow, and one sample happens to contain a single unusually large clast, the bulk density values for the two samples will be very different. Because sampling methods are not standardized, some reported bulk densities were made using the entire mixture of large and small particles, while others were made using only the finer-grained components of the flows. Source area sediments are also important in bulk density computations. Debris flows rich in pumice will have unusually low bulk density values.

Table 2. Viscosity and strength of some common fluids (from Weltmann 1960; Reiner 1960)

Fluid	Viscosity (poises)	Strength (dn/cm^2)	Type of fluid
Water (20 °C)	0.01	0	Newtonian
Ketchup	0.83	150	non-Newtonian
Mustard	2.94	390	non-Newtonian
Shaving lather	3.46	210	non-Newtonian
Mayonnaise	6.33	850	non-Newtonian
Margarine	7.77	480	non-Newtonian
Cement	24	480	non-Newtonian
Mortar	34	950	non-Newtonian
Honey	115	580	non-Newtonian

During floods when large amounts of sediment are being moved, bulk densities of streamflows are typically 1.01 to 1.30 g/cm³. Measured bulk densities range from 1.40 g/cm³ and less for very fluid sediment flows in Japan (Okuda et al. 1977) and Costa Rica (Waldron 1967), to 2.53 g/cm³ for a relatively dry debris flow in the Southern Rocky Mountains (Curry 1966). Bulk densities of flows less than 1.8 g/cm³ are probably hyperconcentrated sediment flows, and not true debris flows unless the flows have an unusually large amount of fines, such as mudflows originating in the loess areas of China, or contain montmorillonite clays. Debris flows have a range in volume concentration of solids of 25 to 86%, and solids weight proportion of about 35 to 90%. The water content of debris flows generally ranges from about 10 to 30% or greater by weight. These measured values strongly support the statement that "all that flows is not water"!

clast size

Debris flows have been known to carry boulder material over 20 km from their source in a single flow (e.g., Sharp and Nobles 1953). Takahashi (1981) reports a famous boulder weighing 3,000 t that was moved several kilometers by a debris flow in Japan. In New Zealand a 37 t boulder was transported 57 km (Macdonald 1972). If channel size and shape remain constant downstream, Oliferov (1970) found that the diameter of the largest transported boulders depended on the average flow depth. Figure 1 shows one unusually large boulder transported by a debris flow in Colorado.

Shear stress equation

Debris flows can be very erosive during passage through steep channels (e.g., Pierson 1980b; Janda, et al. 1981). The total shear stress exerted by a fluid on a stream bed is

$$\tau = \varrho g R S$$

where τ is total shear stress, ϱ is fluid density, g is gravitational acceleration, R is hydraulic radius (approximately average flow depth in wide channels), and S is friction slope (approximated by channel slope for uniform flows).

Scour to bedrk in L. Riv.

During passage of a debris flow, the density can be twice as great as during water floods, and the flow depth during debris surges is greater than that of muddy streamflow between surges. Pierson (1980b) observed debris-flow surges 1 m deep in New Zealand, and streamflow between surges 0.3 m deep. Debris flows can thus exert up to 6 times as much shear stress on channel beds as waterfloods between surges. Pierson (1980b) documented 4 m of channel erosion into sheared bedrock, and 11 m into unconsolidated gravels in less than 24 hours. An adjacent basin with similar characteristics produced no debris flows, only muddy streamflows from the same storm, and downcutting in unconsolidated gravels was less than 1 m. Campbell (1975) also reports extensive erosion by debris flows in the San Gabriel Mountains, California on slopes steeper than about 11°. Other investigators attribute much of the erosion in debris flow channels to scour by more fluid water and mud floods following passage of the debris flow front (e.g., Blackwelder 1928; Temple and Rapp 1972; Morton and Campbell 1974).

Reported volumes of individual debris flow deposits range from less than 0.1 m³ (Hampton, 1972) to over 10⁶ m³ (Sokolovskii, 1968), although smaller and larger flows have probably occurred.

An additional interesting characteristic of debris flows is their ability to travel great distances over low slopes. The mobility of debris flows is highly dependent on

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shear strength, so that no deformation occurs. This is termed a "rigid" plug (Johnson 1970; Hampton 1972). Passage of this rigid plug usually forms a U-shaped channel (Johnson 1970). The U-shape of debris-flow channels is usually modified by subsequent water floods incising V-shaped or rectangular notches into channels.

Unlike turbulent water-flows, the theoretical velocity distribution of debris flows is characterized by this rigid plug. The plug is not sheared; instead it is bounded by zones of laminar flow which move at a uniform velocity (Reiner, 1956; Johnson 1970, 1979) (Fig. 4). Velocity distributions which closely resemble that predicted by theory (rigid plug flows) have been measured or observed in experimental flows (Johnson 1970; Hampton 1972) as well as in natural debris flows (Johnson 1970; Wasson 1978).

Dilatant Model

Based on the experimental results of Bagnold (1954), Takahashi (1978, 1980, 1981) modeled debris flows as dilatant fluids. Using Bagnold's concept of dispersive pressure P (discussed later), the shear stress of debris flows is

$$\tau = P \tan \Theta$$

where τ is shear stress, P is dispersive pressure, and Θ is the dynamic angle of internal friction. In this model, the velocity distribution is given by

$$\frac{V_s - V}{V_s} = \left(1 - \frac{y}{\bar{y}}\right)^{3/2}$$

where V_s and V are velocity at the surface and at height y, and $\bar{y}$ is mean flow depth. Stones moving by inertia on the surface of the flow reault from the increased influence of dispersive pressure P on larger particles, and not necessarily plug flow.

Boulder Transport and Suspension of Solids

process

Many people have reported the transport of unusually large boulders, some several meters in diameter, by debris flows (e.g., Johnson 1970; Fisher 1971; Blackwelder 1928; Wasson 1978). These boulders appear to "float" or weakly tumble along in the debris flow and are later found on fan surfaces or in channels with very shallow slopes, supported and surrounded by finer debris. These isolated, coarse particles tend to sink in the flow due to gravity, but they remain supported in the fine-grained matrix. This leads to the logical question of what keeps the solid load suspended and supported in debris flows and prevents separation. Five mechanisms have been suggested: 1. cohesion; 2. buoyancy; 3. dispersive pressures; 4. turbulence; and 5. structural support. All five mechanisms may be operating to a greater or lesser extent during a debris flow, but only cohesive strength, buoyancy, and structural support can act in static, freshly deposited debris flow sediments.

Cohesion. Cohesion of clay-water slurries has been proposed as a major particle-support mechanism by numerous investigators (e.g., Johnson 1970; Hampton 1975, 1979; Rodine and Johnson 1976). A static clay and water slurry with a density of 1.17 g/cm^3 can indefinitely suspend medium sand, and a slurry with density 1.26 g/cm^3 can support coarse sand (Kuenen 1951). The importance of cohesive strength in supporting solid particles is thus constrained by the amount of clay present in the debris. Many debris flows contain less than 8–10% clay. Slurries with this amount of clay will suspend only sand-sized particles indefinitely, yet boulders over 1 m in diameter have been observed in such flows. Particles coarser than sand must be supported by other forces.

Buoyancy. Buoyancy, in conjunction with cohesive strength, is considered to be another major particle-support mechanism in debris flows (Johnson 1970; Hampton 1975, 1979; Middleton and Hampton 1976). Buoyancy is determined by the difference in density between the submerged solids and the fluid. Rodine and Johnson (1976) and Hampton (1979) believe the buoyant force acting on a boulder in a debris flow is equal to the weight of all the displaced material (solids as well as fluids). In many debris flows, this density difference between displaced material and solid particle can be quite small. For a particle with a density ϱ_s of 2.65 g/cm^3 and a fluid density ϱ_f of 2.0 g/cm^3, the submerged weight is:

$$\frac{\varrho_s - \varrho_f}{\varrho_s} \text{ or } \frac{2.65 - 2.0}{2.65} = 0.25 .$$

This indicates that the submerged weight of a boulder in the debris flow is only about one-quarter its dry weight. If the density of the debris flow is 2.4 g/cm^3, the submerged weight would be only about 10% of its dry weight. Buoyancy could thus support about 75–90% of the particle weight in debris flows.

The effect of buoyancy is enhanced by an increase in pore pressure gradient caused by the partial transfer of the weight of solid particles to the pore fluid in the debris flow. The fine-grained matrix prevents rapid dissipation of this pressure, increasing pore pressure and buoyancy in the flow (Hampton 1979). This also greatly reduces shear strength of the debris and increases its mobility.

Coarsen-up

Dispersive Pressure. Bagnold (1954) experimently demonstrated that when a relatively high concentration of poorly sorted grains are sheared by flow, the larger particles tend to drift toward the free surface. This results from lift produced when forces are transmitted between particles in collision or near collision as one is sheared over another. Bagnold (1954) referred to this upward stress as dispersive pressure, and formulated the equation:

$$P = 0.042\,\lambda\,D^2\left(\frac{dv}{dy}\right)^2 \cos\Theta$$

where P is dispersive pressure, λ is linear grain concentration, dv/dy is velocity gradient, Θ is dynamic angle of internal friction, and D is particle diameter.

The dispersive pressure on a given particle increases as the square of the diameter. Since dispersive forces act more strongly on the largest particles, forcing them away from zones of maximum shearing near the channel bed, the coarsest particles should migrate to the front and top of debris flows. This is commonly the case observed in nature (Fig. 1).

Turbulence. As fine-grained material is added to water, fall velocity of particles decreases (Graf 1971). Five percent silt by weight in flowing water dampens eddy currents, decreasing turbulence (Lane 1940; Vanoni and Nomicos 1960). Turbulence is the variation in direction and magnitude of velocity vectors with time. It is generally acknowledged to be an important component of sediment entrainment and transport in water (Vanoni 1975). But the efficacy of turbulence in debris flows is questionable because of the high viscosity and cohesion, as well as the laminar appearance of most debris flows (Johnson 1970; Hampton 1972).

The preservation of intact brittle shale fragments and fractured boulders, blocks of unconsolidated colluvium, and chunks of soil source materials (Johnson 1970; Lawson 1982; Janda et al. 1981) is strong evidence for laminar flow, or at least greatly surpressed turbulence, in some debris flows. Lawson (1982) observed a much more active role of turbulence in sediment transport in debris flows as water content increased. In New Zealand, high velocity debris flows were quite turbulent (Pierson, 1980a). Enos (1977) describes some sedimentologic evidence for laminar flow from debris flow deposits, including clast fabric, preservation of delicate clasts, projection of large boulders from the top of deposits, and the absence of flutes in sediments associated with debris flows.

Structural support. In fresh, static debris-flow deposits which support large particles, dispersive pressures and turbulence cannot be acting as particle-support mechanisms. Pierson (1981) argues that for typical densities of boulders and matrix materials in debris flows, the submerged weight of large particles is only about 1/4 of their dry weight. Neither does the small amount of clay (11%) in the matrix material of the New Zealand deposits he studied provide enough cohesion to support the submerged weight of the large particles. Pierson (1981) reports that large clasts "floating" on fresh debris-flow deposits could be pushed into the deposit, and when the force ceased, would remain at the attained level. Apparently some support mechanism other than buoyancy and cohesive strength must exist in fresh, static debris-flow deposits.

In static debris-flow deposits, grain-to-grain contacts, or structural support are provided by a framework of particles in contact with the bed and with each other. Structural support comes into play at sediment volume concentrations of 35 to 58% (Pierson 1981), and supports about 1/3 of the weight of coarse particles. This structural support, in addition to cohesive strength and buoyancy, is necessary to keep large boulders suspended in fresh debris-flow deposits. In dynamic or rapidly moving debris flows, structural support will not operate, and dispersive pressure and/or turbulence will occur in addition to buoyancy and cohesive strength. As the physical properties of debris flows vary, so will the relative importance of the various particle-support mechanisms.

Deposition of Debris Flows

Although exceptionally large, fluid debris flows can flow for many kilometers beyond their source areas, viscous flows tend to stop upon reaching a relatively low gradient or in areas of decreased confinement such as on alluvial fans at the mouth of small basins and canyons. These flows can spread out, thin, and stop in place when internal shear stress is exceeded by the shear strength of the flow.

Alluvial fans are found in valleys or in foothills of mountains in all latitudes, irrespective of climate (Rachocki 1981). It was not until after the publication of Blackwelder's (1928) paper on mudflows in semiarid mountains that people began to appreciate the importance of debris flows in building alluvial fans (Beaty 1963, 1974). Earlier workers had believed water-flooding was the only depositional mechanism on alluvial fans (e.g., Gilbert 1882; Trowbridge 1911; Lawson 1915). Probably the importance ~~of debris flows in~~ alluvial fan formation ~~was~~ unrecognized for so long due to the long recurrence intervals between debris flows (Table 3) and the ex~~tensive reworking of deposits by more frequent water flows.~~

NOT VOLCANIC

The recurrence intervals of debris flows are not controlled solely by rainfall frequency. A small rainstorm may produce a debris flow in a basin at one time, but a more rare (larger) storm another time may produce ~~only flash flooding.~~ Sediment availability must ~~be considered (Lumb 1975).~~ The rate of poorly sorted colluvium formation on steep side slopes in source areas, which is controlled by rates of weathering, is also an important factor ~~controlling the frequency of debris flows.~~

Generally, more than one type of deposition occurs on most alluvial fans (e.g., Bull 1964; Ryder 1971a, 1971b). However, some fans consist almost entirely of debris-flow deposits (e.g., Pierson 1980a; Wasson 1978; Mills 1982; Williams and Guy 1973). Other fans consist almost exclusively of water-laid sediments, where conditions for debris-flow formation are surpressed. For example, small basins underlain by limestone or quartzite produce very little fine material, and debris flow formation is thus inhibited (Hooke 1967). These types of fans seem to be infrequent compared to debris-flow or compound fans, however. Blissenbach (1954) reports a decrease in the amount of mudflow deposits in fans in Arizona as mean annual precipitation increases from 279 mm (20–40% mudflow deposits) to 483 mm (5–10% mudflow deposits). McPherson and Hirst (1972) also report that water-laid deposits predominate over debris-flow deposits in cold, temperate climates. This is at variance with the conclusions of Winder (1965) who thought mudflows were the dominant sediment source on Canadian alpine alluvial fans.

A possible decrease in the importance of debris flows in the formation of alluvial fans as mean annual precipitation increases is not supported by the work of Mills (1982) in the southern Blue Ridge Mountains. Fans in this region are almost exclusively formed by debris flows. Debris-flow deposits are commonly reworked, sorted, and stratified by the watery tails of debris flows or by subsequent water flows when the water flows are competent to winnow deposits (e.g., Selby 1974; Johnson and Rahn 1970). This effect would be enhanced in ~~more humid climates because of increased~~ streamflows. Thus ~~debris~~ flows need not be subordinate to water-laid sediments in humid environments; they just have more opportunity to be re~~worked, destroying the stratigraphic evidence for the original genesis of the de~~posits. The reason Blue Ridge debris flows are preserved and little affected by subsequent streamflows is that they are so coarse, normal stream-flows are incapable of reworking the deposits. Fan sediments in Arizona and Canada mentioned above are much finer than sediments in Blue Ridge fans.

The proportion of debris-flow deposits and water-laid deposits can vary vertically and downslope from the head of the fan, as well as throughout the history of

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Table 3. Estimated recurrence intervals of debris flows

Location	Estimated recurrence intervals (years)	Basis of Estimate	Reference
Montgomery Creek, White Mts., California	300 – 500	Geomorphology	Beaty 1974
Mayflower Gulch, Tenmile Range, Colorado	150 – 400	Lichenometry	Curry 1966
Pfeiffer-Redwood Creek, Santa Lucia, California	140	Dendrochronology, stratigraphy, radiocarbon	Jackson, 1977
Alesätno, Nissunvagge Rivers and Tribs., Northern Sweden	50 – 400	Historical data, lichenometry, ppt. records, geomorphology	Rapp and Nyberg 1981
Takahara River, Japan	300	Historical data, radiocarbon	Iso, et al. 1980
Volcanic Mountains, Japan	0.2– 0.4	Historical data	Okuda 1978
Andöya Island, Norway	50 – 60	Historical data	Rapp and Strömquist 1976
Torneträsk-Narvik Area, Lappland	8	Ppt. records	Rapp and Strömquist 1976
Steel Creek, Yukon	(decades)	Vegetation, historical data	Broscoe and Thomson 1969
Mt. St. Helens, Washington	500– 3,000	Stratigraphy, radiocarbon	Crandell and Mullineaux 1978
Mt. Shasta, California	600– 5,000 ("large" debris flow) 10– 25 ("small" debris flow)	Stratigraphy, radiocarbon	Miller 1980
Davis Creek, Virginia	3,000– 6,000	Stratigraphy, radiocarbon	Kochel et al 1982
Blanco Mt., White Mts., California	300	Dendrochronology	La Marche 1968
Mt Baker, Washington	150	Stratigraphy, radiocarbon	Hyde and Crandell 1978
Nisqually River, Mt. Rainier, Washington	800	Stratigraphy, radiocarbon	Crandell 1971
White River, Mt. Rainier, Washington	600	Stratigraphy, radiocarbon	Crandell 1971
Reshūn Stream, Pakistan	> 30 ("large" debris flow) < 10 ("small" debris flow)	Historical data	Wasson, 1978
Portland and Cascade Creeks, Ouray, Colorado	10	Historical data	Simons Li and Associates 1982
Santa Monica Mts., California	75– 150	Ppt. records	Campbell 1975
Cambria County, Pennsylvania	5,000–10,000	Ppt. records	Pomeroy 1980
Rocky Mts., British Columbia	15– 25	Dendrochronology	Gardner 1982
Fall Creek Tributary, New York	10– 70	Ppt. records	Renwick 1977

the fan. Hooke (1967) reports that much of the deposition near fan heads originates from debris flows overtopping channel banks. Deep entrenchment of fan-head streams prevents sheet flow deposits from forming here. Thus water-laid deposits would be expected to be more common in the middle and lower parts of fans. This may partly explain the decrease in maximum and mean size of particles down the surface of alluvial fans reported by many workers (e.g., Bull 1977).

The exact mechanism by which debris flows stop flowing is uncertain. Lateral spreading may permit thickness of flows to decrease below that needed to flow. Some debris flows spread out onto fan surface as sheets, closely paralleling older fan topography. The aerial extent of debris-flow deposits is controlled by volume and strength of the flows, and by the slope of the fan surface. Debris flows also stop flowing when escaping pore fluids (water, clay, and fine silt) cause an increase in internal friction. The rate of pore-fluid escape is a function of the sorting of the deposit.

Takahashi (1981) experimentally determined that the stable slope angle for debris-flow deposition is a function of grain concentration by volume in the static debris deposits, density of fluid and solids, particle size, depth of flow, and angle of internal friction. Debris flows cease flowing in drainage channels when internal friction increases and volume, thickness, strength, and channel slope decrease, causing deposition. Debris flows that stop flowing in channels can form temporary debris dams that can be remobilized by another surge. Alternatively, such deposits can remain in the channel for some time before being eroded by either another debris flow or by normal stream flows.

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At the distal and marginal edges of flows, lobes with steep fronts and concentrations of boulders frequently occur (Fig. 1). Some debris flow sheet deposits can be 0.3 to 1–2 m thick (Pierson 1980a). Thickness of deposits decreases downfan. Bull (1964) describes a mudflow at the mouth of Arroyo Cierro, California where thickness decreases from 48 cm at the fan apex to 9 cm at a distance 1.1 km downslope.

Other debris flows remain in discrete channels when they flow onto fans, shifting course frequently as channels become clogged with sediment (Beaty 1963; Hooke 1967). Freshly-deposited flows can be remobilized by subsequent debris-flow surges if not completely dry (Sharp and Nobles 1953), or reworked by subsequent water floods (e.g., Johnson and Rahn 1970).

Just prior to deposition, debris flows must be moving quite slowly since small vegetation on fans and in channels is capable of diverting debris flows transporting very coarse boulders, without being knocked over or scarred. Figure 3 shows an unscarred base of a willow with a diameter of 30 mm that was engulfed by the lobe of a debris flow in 1977 (Costa and Jarrett 1981).

Long-term average rates of accretion of alluvial fans vary widely. In the White Mountains of California, Beaty (1970) esimates an average accretion rate of 1.5 cm/100 years. Ryder (1971a) estimates a rate of 5.1 cm/100 years for fans in British Columbia, Canada, and Bull (1964) estimates an accretion rate of 34 cm/100 years for Arroyo Ciervo, California. At Mt. Thomas, New Zealand, aggradation occurred at a rate up to 2727 cm/100 years (Pierson 1980a).

United States indicate that the transition from water floods to debris flows, as expressed by deposits and landforms left by the flows, is abrupt and easy to recognize. Geomorphic and sedimentologic criteria, as discussed in the following pages, are thus a more practical and meaningful way to ascertain process than a sediment concentration boundary or threshold that has not yet been adequately defined for the range of materials encountered in natural flows.

Several kinds of geomorphic and sedimentologic evidence remain in small mountain basins following water and mud floods, and following debris flows that can be used to differentiate the two processes. This evidence includes: 1. the presence or absence of coarse, poorly sorted levees and terminal lobes on fans and bordering channels; 2. sedimentology of deposits; 3. the extent of damage to vegetation on fans at the mouths of basins and in stream valleys; 4. the extent of ground-litter disruption below high-water marks; and 5. analysis of records from gaging stations downstream from a basin.

A problem of interpretation exists however, when a debris flow is followed by a more fluid mud or water flood, or reworks old debris-flow deposits. The reworking and sorting of debris-flow deposits by subsequent water flows is apparently common (Hooke 1967; Broscoe and Thomson 1969; Bluck 1964; Sharp 1942; Beaty 1963; Vinogradov 1969; Wasson 1978; Blackwelder 1928; Johnson and Rahn 1970; Temple and Rapp 1972).

Levees and terminal lobes. As debris flows progress downslope, dispersive forces cause migration of large particles to the margins of the flow. Lateral areas of the flow mass are pushed to the sides and sheared from it as the rigid plug passes through the middle of the flow, leaving distinctive levees which are often studded with large boulders (Fig. 1) (Sharp 1942). Debris flows will commonly continue downslope until they deposit most of their mass as levees.

When debris flows stop, the strength of the material or concentrations of coarse clasts at the margins of the flow allows the formation of steep fronts and sides, creating terminal lobes of finite thickness on sloping ground.

Boulder berms. Boulder berms are open-framework coarse gravels and boulders deposited across and adjacent to stream channels in valleys and on fans. They have been observed following large water and mud floods along steep mountain channels in California (Stewart and LaMarche 1967; Scott and Gravlee 1968), and by me after large rainfall and dam-break floods in the Southern Rocky Moutains in Colorado (Fig. 7). Krumbein (1942) refers to similar features in the Arroyo Seco, California as "boulder jams", but they in fact may be debris flow ~~levees and lobes.~~ Boulder berms are created by water and mud ~~floods, whereas~~ levees and lobes are the result of debris ~~flows.~~

Boulder berms, while similar in some ways to debris-flow levees and lobes, are distinctly different. The largest boulders are found at or near the surface of berms, but they have no fine-grained matrix. Some debris flow deposits may have the upper portions of fine-grained matrix washed away by rainfall or stream flows, but boulder berms are grain-supported from the top of the deposit to the bottom. The tops of some of the largest boulders may protrude above the high-water marks on valley sides. Boulder berms are highly localized along stream channels, unlike de-

Fig. 7. Photograph of boulder berm formed following dam-break flood along Fall River. Rocky Mountain Park, Colorado. Backpack sitting on boulder in left-center of photograph for scale

bris-flow levees which can be continuous for long reaches. Berms tend to form below areas of extensive erosion and in expanding valley reaches where more coarse material is supplied to the channel by landslides and channel erosion than can be transported by the available water.

The exact origin of boulder berms remains a mystery. They may form as slip faces and sides of large dune or delta bedforms. or they may represent the front of subaqueous viscous flows in which the bedload moves as a churning mass. They may originate from macroturbulence effects. Matthes (1947) describes a number of forms of macroturbulence that occur in swift and deep streams. Vortex action. called kolks. similar to tornadoes in air. cause upward suction and lift of coarse bed materials. They may be the margins of debris-torrent deposits (Miles and Kellerhals 1981). Boulder berms remain an important but poorly understood bedform in gravel rivers.

Sedimentologic Evidence

really?

When debris flows stop. the resulting deposits consist of a uniform distribution of sizes up through boulders in a matrix of fine-grained debris, forming a pebbly-mudstone deposit (diamicton). Boulders are supported in a matrix containing substantial amounts of fine-grained sediment. Some debris-flow deposits can be clast-supported if the matrix drains or is washed away. However some fine matrix material may occasionally be found beneath the washed surface boulders (Costa and Jarrett 1981).

Despite this complication, the distinguishing feature of undisturbed debris flows is a mud matrix surrounding larger particles (Blackwelder 1928; Crandell 1971). Debris-flow matrix may also contain light-weight materials such as wood and bark fragments, pine needles and cone chips, and animal droppings which should have floated away if water and mud floods were reponsible for the deposits (Sharp and Nobles, 1953).

Abundant bubble holes (vesicles) are also more common in the fine matrix material of debris flows than in water-deposited fine sediments (Sharp and Nobles 1953; Bull 1964; Crandell 1971). Bubble holes form when air is incorporated into debris flows as they move down channels, and from soil air moving upward from ground freshly covered by debris-flow sediments.

Water-laid sediments may be sorted, cross-bedded, stratified, or massive, with gradational boundaries. Debris-flow deposits are much more poorly sorted than water-laid deposits, and bedding is virtually non-existent. Contacts tend to be sharp. On alluvial fans, water-laid sediments consist of (a) sheets of gravel, sand, and silt deposited by braided distributary channels; (b) fill of entrenched channels with coarse, poorly-sorted sediments; and (c) lobes of coarse material which form where fan surfaces are so porous that water rapidly infiltrates into the ground. These lobes of coarse, open-framework sediments are rare, but distinctive, and have been named "sieve deposits" (Hooke 1967). They apparently do not form if too much fine material is present to plug underlying fan materials.

Sieve deposits can be differentiated from debris-flow sediments by various criteria (Hooke 1967). First, recent sieve deposits have an open framework of coarse materials with no fine matrix. Some fine matrix can form over time by post-depositional weathering. Recent debris-flow deposits have a fine-grained matrix if it has not drained or been washed away. Second, debris flows may contain unusually large boulders (greater than one meter diameter). Third, debris-flow deposits are thinner and wider spread. Sieve deposits, on the other hand, tend to be narrow and taller. Fourth, contacts between debris-flow sediments and underlying materials tend to be sharp and well defined, whereas sieve deposit contacts are gradational. Fifth, sieve deposits tend to be relatively short and deposited with slopes less than the fan slope; debris-flow deposits are relatively long and are deposited at slopes approximately equal to the fan slope. Sixth, debris flows commonly form lateral levees, while sieve deposits do not. Seventh, fresh sieve deposits are associated with stream channels, while on fan surfaces debris flows deposit sediment on interfluves as well as in channels.

Because of the small difference in density between boulders and fluid material in debris flows, buoyant forces and dispersive pressures may concentrate boulders at the top of the deposit, forming reverse grading (Fisher 1971). However some debris-flow deposits are normally graded. Clast fabric can also be used to identify debris-flow deposits. In thick, viscous flows with a relatively small water content, the larger clasts have a random orientation and distribution throughout the deposit. In more fluid flows with lower viscosities, particles may show a poorly preferred orientation parallel or perpendicular to the flow direction. Water-deposited sediments can exhibit graded bedding and horizontal imbrication of gravel clasts (Bull 1977; Lawson 1982).

It has been suggested that some measure of sorting might be a valuable clue to process in mountain channels (Costa and Jarrett 1981). The Trask sorting coefficient hydrology methods must be increased by a factor based on measured sediment accumulations in reservoirs or debris basins, or by assuming that the debris flows contain some average amount of solids by weight. This factor used to increase clear-water discharge estimates is called a "bulking factor." Bulking factors for clear-water peak discharge estimates applied to debris flows range from 1.38 for flows with 50% solids by weight, to 4.40 for flows with 90% solids by weight (Table 8). Bulking factors for "average" debris flows will probably vary between 1.5–2.0.

The Los Angeles County Flood Control District has developed a series of curves for estimating peak discharge bulking factors as a function of drainage area for different regions in the Los Angeles and Santa Clara River Basins (Hydraulics Division 1971). These curves are based on extensive data from numerous debris basins. Bulking factors decrease gradually as drainage area increases (Fig. 13). A maximum bulking factor of 2.0 was determined as applicable to peak flows within maximum debris potential areas. For a drainage area of 1.6 km² (1 mi²), bulking factors in the Los Angeles Basin range from 1.53 to 2.0 based on differences in topography, geology, and rainfall. This is equivalent to a range of 58–73% solids by weight in debris flows. The Los Angeles County data may not apply to other geographic areas.

Superelevation of Flows Around Channel Bends and Runup

The tendency for fluid flows to reach higher elevations on the outside of channel bends than on the inside has been observed by several investigators, and can be used to estimate mean velocity through the bend (Chow 1959; Guy 1971; and Apmann 1973). The principle is based on the radial acceleration of flow around a bend, and is independent of fluid density. Applying Newton's second law of motion to centrifugal action in the curve and assuming that all stream lines have equal velocity and equal radius of curvature (r_c), the flow surface can be approximated as a straight line (Fig. 14). Johnson (1979) derives the relationship:

$$a_r = \frac{\bar{v}^2}{r_c}$$

where a_r is radial acceleration, $\bar{v}$ is mean velocity, and r_c is radius of curvature. For Fig. 14:

$$\tan\Phi = \frac{a_r}{g \cos S} = \frac{\bar{v}^2}{r_c\, g \cos S}$$

where S is channel slope, $\tan \Phi$ is $\Delta h/w$, w is width, and Δh is elevation difference between the flow surface on the inside and outside of the bend. Solving for mean velocity, $\bar{v} = (r_c\, g \cos S \tan \Phi)^{0.5}$. If channel slope is less than about 15 degrees, $\bar{v} = (r_c\, g \tan \Phi)^{0.5}$.

The mean velocity of a debris flow can thus be estimated if the tilt of the flow surface and radius of curvature of the channel bend can be measured. This method assumes the flow is a perfect fluid. For flows with low strength, this probably does not introduce any greater errors than other indirect methods of calculating velocity for unsteady flows in steep channels. When this technique is applied to mudflow channels draining from Mount St. Helens, Washington, average velocities as great as

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30 m/s are computed (Wigmosta et al. 1981; Janda et al. 1981). This unbelievable average velocity is probably a result of (a) super-acceleration from the lateral blast of the volcanic eruption, or (b) result of poor site selection, or (c) wave splatter setting unrealistically high mud marks. The fundamental validity of velocity values calculated from superelevation, and assumptions therein, remain unverified for use with mud and debris flows (Ikeya and Uehara 1982). Since this method is promising because of its wide applicability, more research is needed on the influences of fluid strength on ~~superelevation.~~

Earlier it was noted that mudlines on surviving trees in the path of debris and mud flows show runup of the fluid on upstream sides. This runup reflects the point surface velocity of the moving fluid, and for debris flows can probably ~~be assumed~~ nearly equal to the mean velocity at that point. ~~Runup from mud~~lines on trees, hills, or canyon ~~walls might~~ be ~~used to obtain any~~ number of point mean velocities of flows by substituting the amount of runup into the velocity head equation ($\Delta h = \alpha v^2/2g$). Strictly speaking, hydrostatic pressure relationships are not valid for flows with strength. However, no data exist to my knowledge to verify just how well or how poorly runup on trees or other obstructions compares with measured velocities for mud and debris flows. Studies have been made on clear-water streamflow (Wilm and Storey 1944). This method needs further investigation.

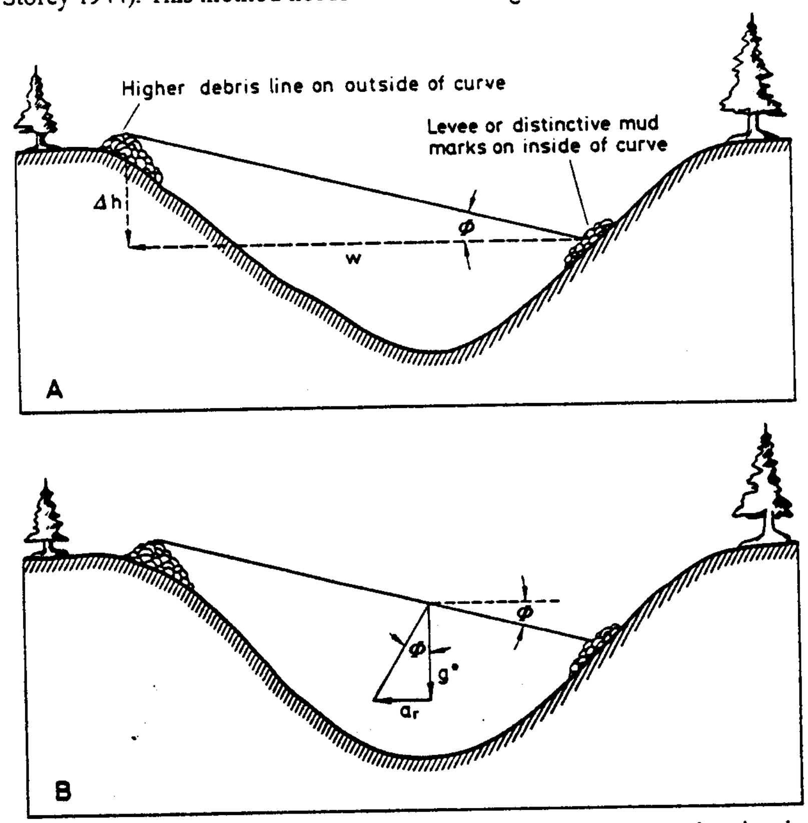

Fig. 14. Technique for estimating average velocity of debris flows from superelevation in curves. (From Johnson 1979)

Photographic Techniques

Mean velocity of debris flows can also be measured using a number of different photographic techniques. Johnson (1970) timed debris flows in Wrightwood, California by taking time-exposure photographs from an overhang with the axis of the camera normal to the surface of flow. The reflective sunlight from mud-coated clasts produce light streaks on the photographs, and by measuring the lengths of the streaks and knowing the exposure time, velocities at many places on the surface of the flow can be calculated. Alternatively, the velocity of debris flows can be calculated from motion picture film if reference points or objects are available, and the number of frames per second is known (Curry 1966; Watanabe and Ikeya 1981). Radar guns have been used to measure velocity of debris flows in Japan (Okuda et al. 1980). Photographic techniques have limited applicability because the investigator must know in advance of the event's pending occurrence, unless remote systems are used.

Mitigation of Debris-Flow Hazards

As with most natural hazards, it may be impossible to provide complete protection from all kinds of mass movements, including debris flows. Techniques for mitigation of debris flow losses may be grouped into four categories: 1. avoidance; 2. control of grading, clearing, and drainage; 3. protective structures; and 4. warning and evacuation.

Avoidance of Hazardous Areas

Because of their elevation above floodplains, debris fans have long been favored sites for development. Some floodplains or other valley-bottom locations are also hazardous. Unfortunately, compared to water floods, mitigating procedures and identification of risk areas for mud and debris flows are poorly developed.

<table>
<tr><th>Class of Phenomenon</th><th>Event</th><th colspan="2">Location</th><th>Covered by NFIP</th><th>Status of Hazard Mapping</th></tr>
<tr><td rowspan="2">FLOODS</td><td>Clear Water Floods</td><td rowspan="2">Floodplains</td><td rowspan="3">Drainage Channels and Alluvial Fans</td><td rowspan="3">Yes</td><td>Now Mapped</td></tr>
<tr><td>Hypercon-centrated Flows</td><td rowspan="3">Not Now Mapped</td></tr>
<tr><td rowspan="2">LANDSLIDES</td><td>Debris Flows</td><td rowspan="2">Hillslopes</td></tr>
<tr><td>Other Landslides</td><td>Hillslopes</td><td>No</td></tr>
</table>

Fig. 15. Classification of floods and landslides by location, and their status under the National Flood Insurance Program. (From Committee for Methodologies for Predicting Mudflow Areas 1982)

Losses from mud and debris flows for insured structures are covered in the United States by the National Flood Insurance Program, however the frequency of such events at a given site, and the inundation areas used to set insurance rates are unknown. It is difficult to regulate and avoid debris flow hazard areas when the dangerous areas and the frequency of inundation cannot be determined systematically and consistently. This is dramatically documented in a recent National Research Council Report on mud and debris flows for the Federal Emergency Management Agency (Committee on Methodologies for Predicting Mudflow areas, 1982) (Fig. 15). The application of conventional engineering hydrology to debris flows leads to greatly erroneous estimates of discharge (Scott 1971; Costa and Jarrett 1981). One flood insurance study for two small tributaries to the Uncompahgre River in southwestern Colorado demonstrates well the frustrations of trying to apply conventional engineering hydrology to debris flows:

"As the nature of flooding of Portland and Cascade Creeks was studied, it became increasingly apparent that the flooding did not follow patterns which could be evaluated by normal hydraulic methods. After evaluating other techniques which might be applied to rivers carrying high loads of silt or debris, a basic conclusion was reached – the floodplains of Portland and Cascade Creeks were variable, unpredictable, and could not be defined."

These two basins appear to be typical of many other small basins in upland areas subject to debris flows.

The accurate identification of a "floodplain" across a debris fan, using conventional hydraulic and hydrologic procedures is not possible (Magura and Wood 1980). Channel blockage and debris-flow deposition result in continually changing channel patterns and locations of deposition (Gundlach 1977–1978). The constantly shifting path of a debris flow across a debris fan makes identification of a single floodplain impossible. In 1914 a debris flow along Cornet Creek caused severe damage to the eastern parts of Telluride, Colorado, built on a debris fan. In 1969 another debris flow down the same stream damaged the western part of the town. Consequently, entire debris fans of small upland streams could be classified as hazard areas. Standard procedures for dealing with inundation hazards on debris fans and alluvial fans have been slow in developing (Magura and Wood 1980; Gundlach 1977–78; Dawdy 1979).

No standard procedures exist for identifying debris flow prone areas, or for calculating the degree of debris flow risk. Consequently, although debris flow ("mudslide," i.e., "mudflow") coverage is included in the standard flood insurance policy, no debris flow maps have been published under the flood insurance program, and no formal system of debris flow management and mitigation has been established.

As a general rule, the bottoms and mouths of small, steep ravines that originate in steep, hilly or mountainous terrain (especially volcanic areas), or in areas of historic and prehistoric debris flows, should be considered potential debris flow areas and avoided. In general, the steeper the slope, the greater the risk.

Control of Grading, Clearing, and Drainage

It is generally believed that erosion by debris flows can be reduced by strict controls of land use, grading, and drainage (Pierson 1980b; Campbell 1975; Hollingsworth and Kovacs 1981). On artificial slopes, this could include limiting the heights of slopes, properly compacting fills, and adequate drainage provisions. Many culverts in debris flow hazard areas are inadequate because they were designed under the assumption that only water flows occur (Campbell 1975).

In the Los Angeles Basin and in mountainous regions of the Soviet Union, China, and New Zealand, devegetation by fire or overgrazing in source areas greatly increases the chance of debris flows (Campbell 1975; Gagoshidze 1969; Li and Luo 1981; Pierson 1980b). Table 9 summarizes available techniques for control of debris-flow gully and slope erosion. Methods are arranged roughly in order of increasing effectiveness and cost. Most of the methods have been widely used with success in Europe, Japan, and Indonesia.

Table 9. Methods for controlling debris-flow gully erosion (Pierson 1980b)

Stabilizing the gully floor	Stabilizing the gully sides	Preventing upslope water and debris from reaching gully
Channel linings (small gullies only) – Hydraulically rough linings: Rubber tires, coarse rock, etc. secured by wire ties or netting	Revegetation (only in relatively small gullies) – Oversowing and top-dressing – Tree planting	Oversowing and top-dressing Tree planting Mulching
– Hydraulically smooth linings: Fiberglass matting, asphalt, concrete	Regrading steep gully walls to flatter slope angles (followed by revegetation)	Contour trenches and barriers Diversion dikes and ditches
Weirs	Internal drainage of rock and soil masses in gully walls with horizontal or vertical drains	Filter berms
Spillways	Reinforcement of sites with terraces, concrete facing, or retaining walls	Gully bypass chutes and flumes
Chutes Sills Openwork sediment trapping dams (interlocking concrete cribbing) Cable net dams Debris dams (check dams) – earth, rock, gabion, concrete cribbing filled with rock Sabo dams		

Protective Structures

Construction of protective barriers and other structures may be necessary if avoidance of debris flow hazard areas is not possible (e.g., due to preexisting development). The purpose of these structures is to stop, slow, or divert debris flows.

Protective measures to reduce property damages from debris flows are different from mitigating measures for water floods. For example, channelization for debris flows is usually ineffective because channels can quickly become blocked, causing subsequent surges to flow in new directions. Channel improvements during the 1964 dry season in the Rio Reventado channel in Costa Rica proved unsuccessful. The first storm of the rainy season promptly filled the enlarged channel with mud and rock debris (Waldron 1967). Reservoirs can become filled quickly and require extensive dredging to maintain design capacity.

Closely-spaced trees have been observed to be quite effective in stopping boulders and other large debris, and allowing the finer fraction of the flow to pass (Fig. 16). In addition to planting closely spaced trees on upslope sides of structures in debris-flow hazard areas, artificial arresting and separating structures such as open-work dams and structural fences of steel and reinforced concrete, steel cable nets, debris fences, and sediment barriers can be effective in stopping and separating large boulders from debris flows (Gagoshidze 1969; Hollingsworth and Kovacs 1981). These structures can be constructed along with debris-storage basins to trap

Fig. 16. Closely spaced trees that were effective in trapping large boulders from a debris flow, Glenwood Springs, Colorado. (From Mears 1977)

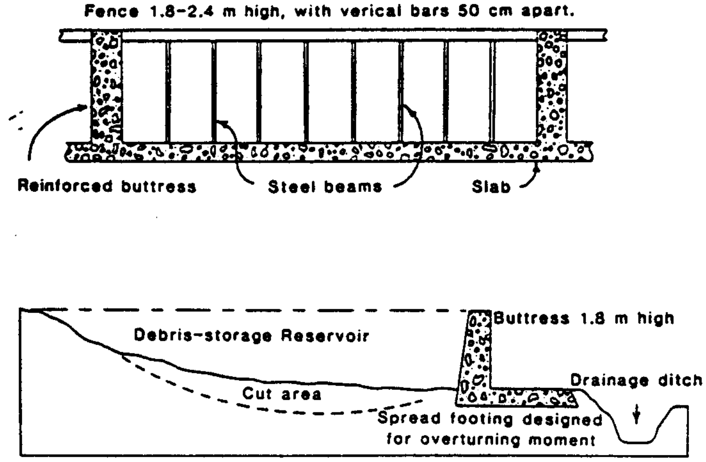

Fig. 17. Structural defense for arresting and separating debris flows. Large boulders and debris are stopped, smaller material and mud are washed through and over the structure and continue as a water flood in the drainage ditch. (From Mears 1977)

sediment, and with drainage channels to contain the separated fluid component of the debris flow (Fig. 17).

Retaining walls are the most commonly recommended structural protection against debris flows. Poured concrete walls have greater impact resistance than block walls (Hollingsworth and Kovacs 1981). Uphill walls, windows, and doors can be reinforced to withstand the impact forces of debris flows. Deflecting or encircling walls are built at an angle to the axis of the moving debris flow to direct it along the wall and away from the protected structure. Dikes to contain and deflect debris flows at Port Alice, British Columbia, Canada, were successfully constructed (Nasmith and Mercer 1979).

Structural works to reduce debris flow hazards have also been attempted in debris source areas. In 1919, lahars from the expulsion of the crater lake on Kelut Volcano, Java, claimed 5,100 lives, destroyed or damaged 100 villages, and destroyed 200 km^2 of farmland (Kemmerling, 1921). Dutch engineers subsequently dug a series of drainage tunnels into the volcano to reduce the volume of lake water from 65×10^6 to 3×10^6 m^3. An eruption in 1951 was much less damaging, but the tunnels were destroyed and the crater lake was deepened by 10 m. The lowermost tunnel was repaired, but the lake volume below the tunnel level had increased to 40×10^6 m^3. A new lower tunnel was driven by the Indonesian Government but stopped short of the lake. It was thought that natural seepage into the tunnel would lower the lake level (Zen and Hadikusumo 1965). This did not succeed because of the low permeability of the volcanic material, and another eruption in 1966 generated lahars that killed hundreds of people and destroyed a large amount of farmland. A new tunnel was constructed in 1967 and the lake levels have been greatly reduced (Bolt et al. 1975).

Warning and Evacuation

Since debris flows frequently result from sudden ground failures and travel at high velocities, it is difficult to provide a direct warning of a specific flow. Ground shaking or the loud sound of approaching debris flows may provide a short warning (Okuda et al. 1980). Sensors or tripwires have been placed along debris-flow channels in Japan to detect the passage of debris flows (Okuda et al. 1980). Following the 1953 Mt. Ruapehu disaster in New Zealand where a railroad bridge was washed out and 151 people killed, a lahar detection system of tripwires was installed (Neall 1976). Detailed instrumentation of hillslopes in debris-flow source areas has been suggested by Campbell (1975), but this is expensive and results are uncertain.

The most promising warning system may be the successful identification of minimum precipitation threshold conditions for slope failures in a particular geographic region. Residents could then be warned when debris flows would be likely if high intensity rain continues (Campbell 1975; Nilsen et al. 1976; Okuda et al. 1980).

Despite expenditures of large sums of money on protective and warning devices, debris flows will probably continue to reap a large toll in property and lives throughout the world. The key to reducing losses probably will come when debris-flow hazard areas can be accurately identified and the risk of occurrence quantified. In the United States, this task has not even been accomplished for relatively well understood and well studied riverine water floods, despite a concentrated effort extending over nearly two decades (Costa 1978). A great deal of work remains to be done on the mechanics, behavior, characteristics, and frequency of debris flows. It is an international problem that may require an international effort to solve, and geomorphology will be a focal point of the resolution.

Further Research

The information in this summary report identifies six general areas of ignorance about mud and debris flows. The following questions identify potentially valuable research areas:

1. Cause. Small rainstorms can trigger debris flows, while a larger rainstorm in the same region may only cause flash flooding. Why? What are the source area characteristics of debris flow prone regions? What are the threshold soil moisture and precipitation conditions for failure? What are the failure mechanisms of debris flows on hillslopes?
2. Identification. Since it is unlikely that trained professionals will observe debris flows in the field, what kind of evidence remains to identify when and where sediment slumps and slides change to form debris flows? At the other end, when do debris flows become mud floods or water flows?
3. Mechanisms. What is the sediment transport mechanism(s) in debris flows, and how does it (do they) vary from area to area, and based on what conditions? What are the mechanisms of debris flow disposition?
4. Characteristics. How can water-laid and debris flow deposits from modern and ancient sediments be recognized? What are the different sedimentological, stratigraphic, and textural differences?

5. Flow parameters. How can indirect velocity and discharge estimates be applied to non-Newtonian debris flows? The validity of velocity estimates from superelevation during flows with high viscosity and strength is unverified. What are roughness coefficients for debris flows?
6. Occurrence. How can areas inundated by debris flows be systematically and consistently identified. Is it possible to make recurrence frequency estimates? Are the characteristics of each debris flow unique and site specific, or can generalizations be made? What aspects of flow behavior must be better known for adequate design of protective structures such as retaining walls and debris fences.

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References

Apmann RP (1973) Estimating discharge from superelevation in bends: Hydraul Div Am Soc Civ Eng 99: HY1, 65–79

Aramaki S (1956) The 1783 activity of Asama Volcano, part 1. Jpn J Geol Geogr 27 (2–4): 189–229

Bagnold RA (1954) Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear. Proc R Soc London Ser A 225: 49–63

Bagnold RA (1956) Flow of cohesionless grains in fluids. Philos Trans R Soc London Ser A 249: 234–297

Balteanu D (1976) Two case studies of mudflows in the Buzau Subcarpathians. Geogr Ann 58 A: 165–171

Beaty CB (1963) Origins of alluvial fans, White Mountains, California and Nevada. Ann Assoc Am Geogr 53: 516–535

Beaty CB (1970) Age and estimated rate of accumulation of an alluvial fan, White Mountains, California: USA. Am J Sci 268: 50–77

Beaty CB (1974) Debris flows, alluvial fans and a revitalized catastrophism Z. Geomorphol 21: 39–51

Benson MA, Dalrymple T (1967) General field and office procedures for indirect discharge measurements. US Geol Surv Tech Water Resour-Invest B 3: Chap A-1, 30

Beverage JP, Culbertson JK (1964) Hyperconcentrations of suspended sediment. Hydraul Div Am Soc Civ Eng HY6: 117–126

Bingham EC, Green H (1919) Paint, a plastic material and not a viscous liquid: the measurement of its mobility and yield value. Proc Am Soc Test Mater 19: part II, 640–664

Blackwelder E (1928) Mudflow as a geologic agent in semi-arid mountains. Geol Soc Am Bull 39: 465–484

Blissenbach E (1954) Geology of alluvial fans in semi-arid regions. Geol Soc Am Bull 65: 175–190

Bluck BJ (1964) Sedimentation of an alluvial fan in southern Nevada. J Sediment Petrol 34: 395–400

Bolt BA (1978) Earthquakes – a primer. Freeman, San Francisco, 241 p

Bolt BA, Horn WL, Macdonald GA, Scott RF (1975) Geological Hazards. Springer, Berlin Heidelberg New York, 328 p

Boulton GS (1968) Flow tills and related deposits on some Vestspitsbergen glaciers. J. Glaciol 7: 391–412

Broscoe AJ, Thomson S (1969) Observations on an alpine mudflow, Steel Creek, Yukon. Can J Earth Sci 6: 219–229

Bull WB (1962) Relation of textural (CM) patterns to depositional environment of alluvial-fan deposits. J Sediment Petrol 32: 211–216