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An empirical method for estimating travel times for wet volcanic mass flows

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Abstract Travel times for wet volcanic mass flows (debris avalanches and lahars) can be forecast as a function of distance from source when the approximate flow rate (peak discharge near the source) can be estimated beforehand. The near-source flow rate is primarily a function of initial flow volume, which should be possible to estimate to an order of magnitude on the basis of geologic, geomorphic, and hydrologic factors at a particular volcano. Least-squares best fits to plots of flow-front travel time as a function of distance from source provide predictive second-degree polynomial equations with high coefficients of determination for four broad size classes of flow based on near-source flow rate: extremely large flows ($>1\,000\,000\text{ m}^3/\text{s}$), very large flows ($100\,000\text{--}1\,000\,000\text{ m}^3/\text{s}$), large flows ($100\text{--}100\,000\text{ m}^3/\text{s}$), and moderate flows ($100\text{--}1000\text{ m}^3/\text{s}$). A strong nonlinear correlation that exists between initial total flow volume and flow rate for “instantaneously” generated debris flows can be used to estimate near-source flow rates in advance. Differences in geomorphic controlling factors among different flows in the data sets have relatively little effect on the strong nonlinear correlations between travel time and distance from source. Differences in flow type may be important, especially for extremely large flows, but this could not be evaluated here. At a given distance away from a volcano, travel times can vary by approximately an order of magnitude depending on flow rate. The method can provide emergency-management officials a means for estimating time windows for evacuation of communities located in hazard zones downstream from potentially hazardous volcanoes.

Key words Travel time · Flow velocity · Lahar · Debris avalanche · Volcanic hazards

Introduction

Gravity-driven mass flows of volcanic rock fragments mixed with water (often mixed also with snow and ice fragments) are among the most hazardous processes at stratovolcanoes around the world, occurring at scales ranging from cubic meters to cubic kilometers of debris. Such flows may occur not only prior to or during eruptions, but also during volcanically quiet (and seemingly safe) periods when they can be triggered by heavy rainfall, lake outbreaks, earthquakes, or simply the progressive weakening of rocks in a volcanic edifice by hydrothermal alteration. Commonly, the occurrence of these flows takes people in downstream valleys by surprise, because the flows can occur suddenly without warning, move rapidly, and reach hundreds of kilometers from their sources. Large volcanic mass flows can destroy infrastructure (bridges, roads, railroads, pipelines, and electric transmission lines), temporarily block rivers, and obliterate entire communities located downstream from source volcanoes. Even relatively small flows can have devastating impacts locally. Within the last 50 years, flows spanning the entire size range investigated for this paper have resulted in substantial loss of human life and property in Chile (Best 1992), Colombia (Lowe et al. 1986; Naranjo et al. 1986; Pierson et al. 1990), Costa Rica (Ulate and Corrales 1966; Waldron 1967; Alvarado and Schmincke 1994), Guatemala (Kuenzi et al. 1979; Rose 1987; Flynn et al. 1991), Indonesia (Schmidt 1934; Suryo and Clarke 1985; Hamidi 1989), Japan (Nakamura 1926; Murai 1960; Inokuchi 1985), New Zealand (O'Shea 1954), the Philippines (Pierson et al. 1992, 1996; Major et al. 1996; Umbal and Rodolfo 1996), and the United States (Cummins 1981; Janda et al. 1981; Voight et al. 1983; Fairchild 1985; Gallino and Pierson 1985).

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This paper focuses on gravity-flow processes (sometimes accelerated by volcanic explosions) that involve substantial volumes of poorly sorted rock particles as their primary solid components, hence the term “mass flows.” These flows are considered “wet” because all have some liquid water (inherited from their prefailure location or acquired during flow) in the interstices between particles for the major part of their flow paths, even if they were triggered by truly dry volcanic mass flows such as pyroclastic flows and surges. The mass flows considered here span a wide range of volumes, flow rates (peak discharges), velocities, compositions, bulk rheologies, and flow hydraulics. Rapid wet (but unsaturated) granular flows, usually classified as debris avalanches (Sharpe 1938; Varnes 1958, 1978; Ui 1983; Pierson and Costa 1987), commonly begin as large landslides from the flanks of volcanoes, involve volumes of debris up to several tens of cubic kilometers, and travel at velocities as high as 360 km/h (Siebert 1992). Debris-avalanche flow mechanisms remain controversial, but it is likely that inertial and frictional interactions between clasts largely control motion and are highly rate dependent. There is evidence that debris avalanches stop very abruptly (Siebert 1996), apparently due to “locking up” of the coarse angular debris during deceleration. Thus, debris avalanches, even extremely large ones, typically do not travel more than several tens of kilometers away from their sources, despite their great bulk and initial speed, unless they become water saturated. Water-saturated mass flows, which have the consistency of wet concrete and are termed debris flows (Pierson and Costa 1987), are more mobile because positive pore-fluid pressures ($p > p_{\text{hydrostatic}}$) greatly decrease internal friction within the debris mass (Hampton 1979; Pierson 1981; Johnson 1984; Iverson and Denlinger 1984; Major 1996; Iverson 1997). Debris flows can flow as fast as 150 km/h and are capable of flowing hundreds of kilometers down valleys away from their sources (Janda et al. 1981; Naranjo et al. 1986; Scott 1988; Pierson et al. 1990; Mothes 1992; Scott et al. 1995). Debris flows can initiate from landslides (sometimes transforming directly from debris avalanches) or from sudden releases of water such as lake outbreaks, rapid snowmelt during eruptions, or heavy rainfall. In water-initiation cases, erosion and incorporation of sediment by flowing water on the steep upper slopes of volcanoes typically result in several-fold increases in flow volume. While clay-rich (“cohesive”) debris flows can travel great distances with little or no change in rheology, debris flows containing less than approximately 5% clay-size particles in the matrix (granular or “noncohesive” debris flows) commonly become progressively more dilute as they flow and eventually evolve into flood waves of very muddy water termed hyperconcentrated flows (Beverage and Culbertson 1964; Pierson and Scott 1985; Pierson and Costa 1987; Scott 1988). The term “lahar” has been recently adapted to include both debris flows and hyperconcentrated flows generated on volcanoes (Smith and

Fritz 1989; Smith and Lowe 1991), because a single flow event can involve both rheologic flow types and multiple flow transformations.

The primary objective of this study was to develop a reliable method for predicting travel times of wet volcanic mass flows that would be useful to (a) emergency-management officials facing decisions about how and when to notify or evacuate communities downstream from a volcano (see Fig. 1) after the onset of such a flow, and (b) land-use planners making decisions about the location and density of business, residential, and recreational development near volcanoes. Such a method has not been previously available. Various computer models using both theoretical and hydrologic (water flood routing) approaches have been developed or adapted in recent years for generalized grain flows or debris flows (Takahashi 1978, 1980; Savage 1984, 1989, 1993; Savage and Hutter 1989, 1991; Mizuyama and Yawata 1987; Chen 1988a, b; Laenen and Hansen 1988;

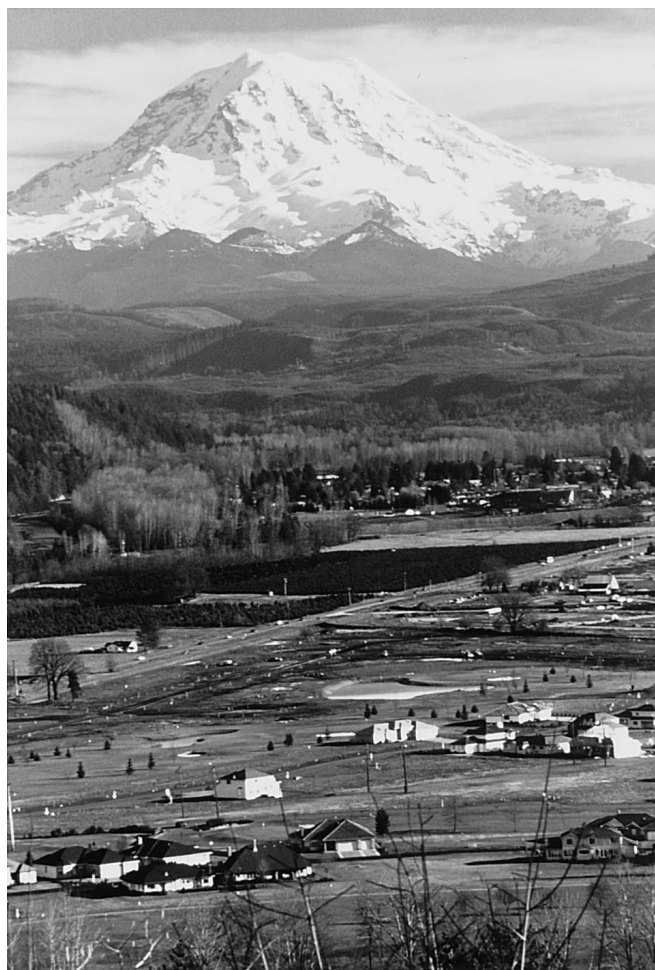


Fig. 1 Town of Orting, Washington, situated in a valley inundated approximately 500 years ago by an extremely large wet volcanic mass flow (the “electron mudflow”; see Scott et al. 1995) that traveled 50 km from Mount Rainier (background) to reach this site. Town leaders are currently developing a warning system and emergency plans to deal with the possibility of another such flow. (Photo courtesy of Dave Wieprecht)

Vignaux and Weir 1990; Macedonio and Pareschi 1992; Caruso and Pareschi 1993; O'Brien et al. 1993; Costa 1997; Iverson 1997), but these models have not yet been adequately developed, scaled, parameterized, calibrated, or tested for reliable prediction of travel time.

Sufficient historical data exist to permit an empirical approach to predicting flow travel times. This paper uses those data to provide a method for estimating time of travel of wet volcanic mass flows. The time available for evacuation from the area or local escape to high ground within flow hazard zones varies from minutes to hours depending primarily on: (a) distance of the site from the volcano; (b) size of the flow; and (c) whether timely warning of an approaching flow is received by the population at risk. If evacuation and crisis-response plans at potentially hazardous volcanoes are to be effective, knowledge of probable flow travel times is critical.

Study methods and results

Single wet mass flows of all sizes, which usually originate from point sources, show a gradual slowing down with distance traveled away from their sources on or near volcanoes (Fig. 2). This occurs because mass flows spread out (flow depths decrease) as channels widen and flatten with increasing distance away from a volcano. Uniform flow equations, such as the Manning equation, which have been shown to provide reasonable approximations for open-channel debris flow as well as for water flow (Costa 1997), predict that velocity should decrease as depth or slope decrease (Chow 1959). This behavior is in contrast to rainfall-generated flood flows (not originating from a point source), which generally increase enough in depth as tributaries add more water to more than compensate for the decrease in slope and thus cause velocities to increase slightly as the floods progress downstream.

The main approach used in this study was to: (a) obtain reliable empirical data on time of travel for wet mass flows of different flow rates and types from numerous volcanoes (some data on nonvolcanic flows were used where data on volcanic flows were not available); (b) group these data according to flow rate, a factor related to flow velocity and thus to its inverse, travel time ($Q = vA = vdw$, where Q is flow rate, A is cross-sectional area of the flow, v is velocity, d is channel depth, and w is channel width); and (c) correlate by groups *distance from source* (independent variable) against *travel time* (dependent variable), in order to obtain regression equations that could be used to estimate flow travel times at other volcanoes. Because flow rates typically decrease with distance traveled (Pierson 1995), *near-source* flow rates were used in the regressions. Near-source flow rate is defined as peak discharge measured or estimated as close as possible to the flow source, typically between 0 and 15 km from the volcano summit, which is the zone where peak discharges are characteristically greatest (Pierson 1995).

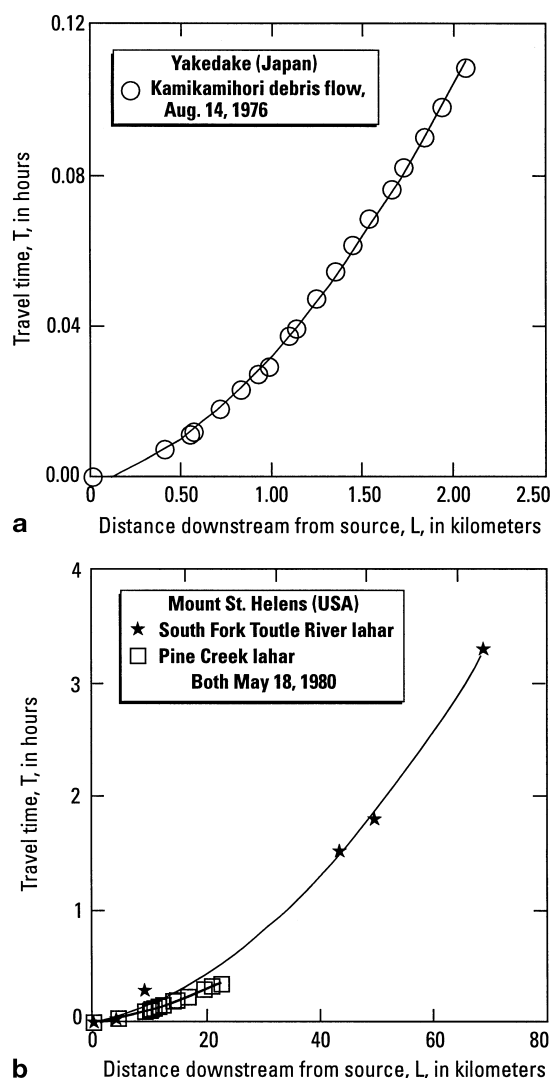


Fig. 2a,b Plots of distance from source vs flow-front travel time for single wet mass flows of different sizes. **a** Debris flow having a near-source flow rate of 100–200 m³/s from Yakedake Volcano, Japan (see Table 4); **b** Debris flows having near-source flow rates of between 50000 and 100000 m³/s from Mount St. Helens, USA (see Table 2). Lines through data points are second-degree polynomial best fits

Empirical data on the arrival times of flow fronts of 40 historical wet mass flows were collected for this analysis; 33 of the flows were lahars (15 having debris-flow rheology for the entire flow path and 18 transforming to hyperconcentrated flow for part of the flow path where travel times were recorded) from nine volcanoes, and seven of the flows were debris avalanches (volcanic and nonvolcanic) from six locations (Tables 1–4). All of the debris avalanches except one were much larger than the lahars. Each data point used in the analysis represents a different mass flow.

The travel-time data were obtained from: (a) direct instrumental measurements of times of flow arrival (video, seismograph, gauging station, and acoustic flow-monitor records); (b) indirectly calculated velocities us-

Table 1 Travel time/distance data for extremely large mass flows; these examples moved predominantly as debris avalanches, although some transformed to debris flow at their distal ends, apparently without first stopping. Peak discharges are assumed to be $>10^6$ m³/s. Only the Mount St. Helens avalanche was a vol-

canic debris avalanche; the others were nonvolcanic and involved a variety of rock types or rock and ice. In cases where more than one travel time was recorded for a given flow, data pairs marked by asterisks (*) were used for the regressions in Figure 3

Location and date of flow; reference	Elapsed time from flow onset to arrival at timing point <i>p</i> (h)	Distance downstream from source to timing point <i>p</i> (km)	Average front velocity from source or timing point <i>p</i> – 1 to timing point <i>p</i> (m/s)	Estimated volume (m ³); origin; type(s) of flow
Mount St. Helens: North Fork Toutle River (1980); Voight et al. (1983)	0.17	23	38	2.8×10^9 ; slope failure; debris avalanche (volcanic)
Mount Cook, New Zealand (1991); Chinn et al. (1992)	0.025	7.3	81	5.5×10^7 ; slope failure; mixed (rock, ice, snow) avalanche (nonvolcanic)
Elm, Switzerland (1881); Hsü (1975; 1978)	0.013	2.2	50	10^7 ; slope failure; debris avalanche (nonvolcanic)
Mayunmarca, Peru (1974); Kojan and Hutchinson (1978)	0.050	8.0	44	10^9 ; slope failure; debris avalanche, debris flow (nonvolcanic)
Huascaran, Peru (1962); Plafker and Ericksen (1978)	0.083	16	53	1.3×10^7 ; slope failure; debris avalanche, debris flow (nonvolcanic)
Huascaran, Peru (1970); Plafker and Ericksen (1978)	0.033 0.058*	9 16*	75 76	$5\text{--}10 \times 10^7$; slope failure; debris avalanche, debris flow (nonvolcanic)

ing superelevation and flow runup equations (Chow 1959; Apmann 1973) that were calibrated by direct measurements; or (c) eyewitness accounts of arrival times that were corroborated by reenactments of survivor actions or by concordance of reports by multiple witnesses. Unverified, indirectly computed velocities were not used.

Data from five nonvolcanic debris avalanches were added to those from the 1980 volcanic debris avalanche at Mount St. Helens and the 1984 debris avalanche from Mt. Ontake, Japan. Although flow mechanics of volcanic and nonvolcanic debris avalanches are assumed to be fundamentally the same, some researchers (Ui 1983; Voight et al. 1985; Hayashi and Self 1992) have documented a statistically significant difference between the mobilities (runout distances) of volcanic and nonvolcanic debris avalanches, possibly due to the presence of hydrothermal alteration products in the volcanic avalanches. This difference may also affect travel time/distance relationships, but given the scarcity of velocity data for volcanic debris avalanches, the nonvolcanic flows were included as the best available approximations.

Mass flows in this study are divided into four broad classes on the basis of near-source flow rate: extremely large (>1000000 m³/s), very large ($10000\text{--}1000000$ m³/s), large ($1000\text{--}10000$ m³/s), and moderate ($100\text{--}1000$ m³/s), and regressions for each size class were run. In addition to the measured distance–time data pairs, hypothetical initial-condition data pairs (0 time for 0 distance traveled) were added to each of the four regressions in order to keep the curves physically meaningful. The largest size class includes data only

from mass flows that were debris avalanches (one volcanic and five nonvolcanic) in the measured reaches, although some transformed to debris flows farther downstream. The last three categories include data from lahars (both pure debris flows and some transforming to hyperconcentrated flows) and one debris avalanche. Flow sources for the lahars and the two volcanic debris avalanches are located at the summits of volcanic cones or on their steep flanks, with the exception of the 1980 North Fork Toutle River lahar, which was generated from liquefied debris-avalanche deposits in a valley 15 km downstream from Mount St. Helens. Flow sources for the nonvolcanic debris avalanches are steep mountain flanks or summits. No flow-rate data were available for the extremely large mass flows used in this study, but given the large volumes of material involved (Table 1) and known valley dimensions, simple calculations indicate that initial discharges would be in excess of 1000000 m³/s. A prehistoric flow of comparable size, the 3.8-km^3 Osceola Mudflow from Mount Rainier (debris avalanche that transformed to debris flow), was estimated by Vallance and Scott (1997) to have had a flow rate of 12.4 million cubic meters per second 10 km from the summit source area.

Correlation of the data shows that flow travel time is positively and nonlinearly dependent on distance from source for the combined flow data in each of the four flow-rate classes (Fig. 3), as it is for individual flows (Fig. 2). Least-squares best-fit regressions of the data show that second-degree polynomial equations provide the best fits to the data. Second-degree coefficients of determination (R^2) are quite high, varying from 0.917 to 0.995. Travel times for the extremely large flows are

Table 2 Travel time/distance data for very large mass flows; these examples moved predominantly as debris flows but commonly with extended runouts of hyperconcentrated flow. Initial peak discharges were in the range of 10⁴ to 10⁶ m³/s. In cases where more than one travel time was recorded for a given flow, data pairs marked by asterisks (*) were used for the regressions in Figure 3

Location and date of flow; reference	Elapsed time from flow onset to arrival at timing point <i>p</i> (h)	Distance downstream from source to timing point <i>p</i> (km)	Average front velocity from source or timing point <i>p</i> – 1 to timing point <i>p</i> (m/s)	Peak discharge nearest source (m ³ /s); origin; type(s) of flow
Cotopaxi, Ecuador: Rio Esmeraldas (1877); Wolf (1878); Mothes (1992)	18	270	4.2	~ 200 000; pyroclastic flow?; debris flow, hyperconcentrated flow
Cotopaxi, Ecuador: Rio Cutuchi (1877); Wolf (1878); Mothes (1992)	0.83	43	14.4	45 000–70 000 at 43 km; pyroclastic flow; debris flow
Mount St. Helens: North Fork Toutle River (1982); Pierson and Scott (1985)	2.20	48.5	6.2	14 000 at 7 km; water flood; debris flow, hyperconcentrated flow
	4.13	73.1	3.5	
	4.78*	81.2*	3.5	
	5.53	86.6	2.0	
Mount St. Helens: Pine Creek (1980); Pierson (1985)	0.034	4.4	35.7	~ 50 000–100 000 at 4 km; 28 600 at 9.9 km; pyroclastic surge/flow; debris flow
	0.094	9.1	27.0	
	0.10	9.9	20.4	
	0.11	10.1	23.9	
	0.12	10.8	15.1	
	0.13	11.5	14.3	
	0.15	12.2	12.5	
	0.18	14.1	16.3	
	0.19	14.8	24.3	
	0.22	16.8	17.6	
	0.29	19.5	10.7	
	0.29	19.6	12.7	
Mount St. Helens: South Fork Toutle River (1980); Cummans (1981); Fairchild (1985); unpublished data	0.32	20.9	13.8	68 000 at 4 km; pyroclastic surge/flow; debris flow, hyperconcentrated flow
	0.33*	22.5*	–	
	0.029	4.0	38.0	
	0.28	9.0	8.9	
	1.50	43.5	7.9	
Nevada del Ruiz, Colombia: Rio Azufrado (1985)/Pierson et al. 1990; unpub. data	1.78*	49.8*	6.3	48 000 at 9.6 km; pyroclastic flow; debris flow
	3.28	69.1	3.6	
	1.37	49.5	10.0	
Nevada del Ruiz, Colombia: Rio Guali (1985)/Pierson et al. 1990; unpub. data	2.28*	72.0*	6.9	20 500 at 17.9 km; pyroclastic flow; debris flow
	2.37	74.0	8.7	
Nevado del Ruiz, Colombia: Rio Chinchina (1985); Pierson et al. (1990); unpublished data	4.37*	102.6*	4.0	19 900 at 11.2 km; pyroclastic flow; debris flow
	0.33	11.3	9.5	
	0.50	16.4	8.3	
	1.17*	33.0*	6.9	
	1.37	37.5	6.3	
	2.40	58.9	5.8	
Mount Ontake, Japan: Denjogawa and Ohtakigawa (1984); Inokuchi (1985)	2.87	68.6	5.7	~ 150 000 at 7–9 km; slope failure; debris avalanche, debris flow
	0.17	11.4	18.6	
	0.13	7.6	16.2	
Redoubt Volcano: Drift River (16 Jan 1990); Dorava and Meyer (1994), unpublished data	0.45*	18.5*	9.5	~ 12 000 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	2.53	44	4.8	
Redoubt Volcano: Drift River (28 Feb 1990); Dorava and Meyer (1994), unpublished data	2.53	44	4.8	~ 10 000 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	0.016	2.4	42.1	
	0.059	6.4	26.0	
	0.29	20.0	16.0	
Tokachidake, Japan: Huranogawa (1926); Nakamura (1926); Murai (1960)	0.48*	23.2*	5.0	14 800 at 8 km; pyroclastic surge/flow; debris flow

Table 3 Travel time/distance data for large mass flows having initial peak discharges of 10^3 to 10^4 m³/s. These examples moved initially as debris flows but most of the Drift River flows apparently had extended runouts of hyperconcentrated flow. In cases

where more than one travel time was recorded for a given flow, data pairs marked by asterisks (*) were used for the regressions in Figure 3

Location and date of flow; reference	Elapsed time from flow onset to arrival at timing point <i>p</i> (h)	Distance downstream from source to timing point <i>p</i> (km)	Average front velocity from source or timing point <i>p</i> – 1 to timing point <i>p</i> (m/s)	Peak discharge nearest source (m ³ /s); origin; type(s) of flow
Mount St. Helens: North Fork Toutle River (1980); Cummins (1981); Fairchild (1985)	0.25	6.0	6.7	7200 at 10.5 km; liquefaction of debris-avalanche deposit; debris flow
	0.55	18.9	6.5	
	1.83	34.0	4.1	
	4.67	48.8	1.5	
	7.33*	71.8*	2.4	
Redoubt Volcano: Drift River (9 March 1990); Dorava and Meyer (1994), unpublished data	0.87	18.5	5.9	~ 2500 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	2.90*	44.0*	3.5	
Redoubt Volcano: Drift River (14 March 1990); Dorava and Meyer (1994), unpublished data	0.22	8.5	10.7	~ 2500 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	0.35	11.8	6.6	
	0.48	18.5	14.3	
	2.35*	44.0*	3.8	
Redoubt Volcano: Drift River (23 March 1990); Dorava and Meyer (1994), unpublished data	0.13	8.5	18.2	~ 2500 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	0.22*	11.8*	10.3	
	1.02	18.5	1.8	
Redoubt Volcano: Drift River (29 March 1990); Dorava and Meyer (1994), unpublished data	0.10	8.5	23.6	~ 2500 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	0.17	11.8	13.2	
	0.27*	18.5*	18.6	
	2.55	44.0	3.1	
Redoubt Volcano: Drift River (6 April 1990); Dorava and Meyer (1994), unpublished data	0.12	8.5	19.7	~ 1000 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	0.20	11.8	11.5	
	0.60*	18.5*	4.6	
Redoubt Volcano: Drift River (15 April 1990); Dorava and Meyer (1994), unpublished data	0.10	8.5	23.6	~ 1000 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow
	0.20*	11.8*	9.2	
	2.95	44.0	3.3	
Redoubt Volcano: Drift River (21 April 1990); Dorava and Meyer (1994), unpublished data	0.23	8.5	10.3	~ 1000 at 19 km; pyroclastic flow; debris flow, hyperconcentrated flow

approximately an order of magnitude less than for the moderate flows.

Interpretation and application

The goodness of fit of the polynomial model to the data indicates that a very large proportion of the total variation in travel time for flows of different size classes can be accounted for simply by distance from source. Flow rate also appears to play an important role in determining travel time (Fig. 4); basically, the larger the flow class, the faster flows travel. This is because two of the three variables that control velocity in open-channel flow (depth and channel slope) are imbedded in flow rate, and channel slope is a direct function of distance from source. However, differences between some classes may not be statistically significant; this was not tested.

Variation in travel time not explained by flow rate and distance from source probably originates from: (a)

physical mechanisms within the flows not related to flow depth and channel slope, such as velocity-dependent shifts in partitioning of energy dissipation between inertial, frictional, and viscous forces (see Iverson and Denlinger 1987; Iverson 1997) or flow transformations; (b) size differences between flows within the same class; (c) conditions and physical mechanisms acting at flow boundaries, such as channel constrictions, changes in channel sinuosity, and bed roughness; (d) error in measurement or reporting of arrival times of flow fronts; and (e) random error arising from how representative this sample is of the population of all such flows.

The difference in relative travel times between debris avalanches and lahars cannot be satisfactorily evaluated with the data presented here. Although one very large debris avalanche (Mt. Ontake, 1984) plots very similarly to lahars of similar magnitude, only debris avalanches make up the “extremely large” category and are much larger than the lahars in this data set. It is likely that transformation to debris flow would slow a

Table 4 Travel time/distance data for moderate mass flows (initial peak discharge 10^2 to 10^3 m³/s). These flows moved predominantly as debris flows but commonly had extended runouts of hyperconcentrated flow. In cases where more than one travel time was recorded for a given flow, data pairs marked asterisks (*) were used for the regressions in Figure 3

Location and date of flow; reference	Elapsed time from flow onset to arrival at timing point <i>p</i> (h)	Distance downstream from source to timing point <i>p</i> (km)	Average front velocity from source or timing point <i>p</i> – 1 to timing point <i>p</i> (m/s)	Peak discharge nearest source (m ³ /s); origin; type(s) of flow
Mt. Pinatubo, Philippines: Sacobia River (25 Aug 1991), unpublished data	0.2	8.7	12.1	400–800 at 6 km; heavy rainfall; debris flow
Mt. Pinabuto, Philippines: Sacobia River (20 Aug 1991), unpublished data	0.26	8.7	9.3	400–800 at 6 km; heavy rainfall; debris flow
Mt. Pinatubo, Philippines: Sacobia River (21 Aug 1991), unpublished data	0.67	8.7	3.6	400–800 at 6 km; heavy rainfall; debris flow
Mt. Pinatubo, Philippines: Sacobia River (18 Aug 1991), unpublished data	0.49	8.7	4.9	400–800 at 6 km; heavy rainfall; debris flow
Redoubt Volcano: Drift River 26 May 1990, Dorava and Meyer (1994); unpublished data	0.53	8.5	4.5	<1000 at 19 km; pyroclastic flow; hyperconcentrated flow?
	0.68*	11.8*	6.2	
	2.53	18.5	1.0	
Mt. Ruapehu, New Zealand: Whangaehu River (1953); O’Shea 1954	2.28	40	4.9	850 at source; rapid lake drainage; debris flow
Mt. Ruapehu, New Zealand: (LH1) Whangaehu River, (18 Sept 1995); S. J. Cronin, written commun., 1997	7.13	56	2.2	220 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Ruapehu, New Zealand: (LH2) Whangaehu River, (20 Sept 1995); S. J. Cronin, written commun., 1997	7.38	56	2.1	>100 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Ruapehu, New Zealand: (LH3) Whangaehu River, (23 Sept 1995); S. J. Cronin, written commun., 1997	4.57	56	3.4	650–1100 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Ruapehu, New Zealand: (LH3d) Whangaehu River, (24 Sept 1995); S. J. Cronin, written commun., 1997	0.22	9	11.4	220–650 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Ruapehu, New Zealand: (LH5f) Whangaehu River, (27 Sept 1995); Cronin et al. 1997; S. J. Cronin, written commun., 1997	1.33	23.5	4.9	600–975 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Ruapehu, New Zealand: (LH6c) Whangaehu River, (29 Sept 1995); S. J. Cronin, written commun., 1997	3.13	42	3.7	400–650 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Ruapehu, New Zealand: (LH10a) Whangaehu River, (7 Oct 1995); S. J. Cronin, written commun., 1997	4.75	56	3.3	235 at 9.5 km; eruptive expulsion of lake water; debris flow, hyperconcentrated flow
Mt. Yakedake, Japan: Kamikamiho-ri (1976); Okuda et al. (1980)	0.0070	0.41	16.5	~100–200 at ~1.5 km; heavy rainfall; debris flow
	0.011	0.55	9.9	
	0.012	0.57	7.0	
	0.018	0.72	6.1	
	0.023	0.83	6.3	
	0.027	0.93	7.7	
	0.029	0.99	6.8	
	0.037	1.10	4.1	
	0.039	1.14	4.0	
	0.047	1.25	3.9	
	0.054	1.35	4.0	
	0.061	1.45	4.0	
	0.068	1.54	3.8	
	0.076	1.66	4.0	
	0.082	1.73	3.1	
	0.090	1.84	4.2	
	0.098	1.94	3.4	
	0.108*	2.07*	3.5	

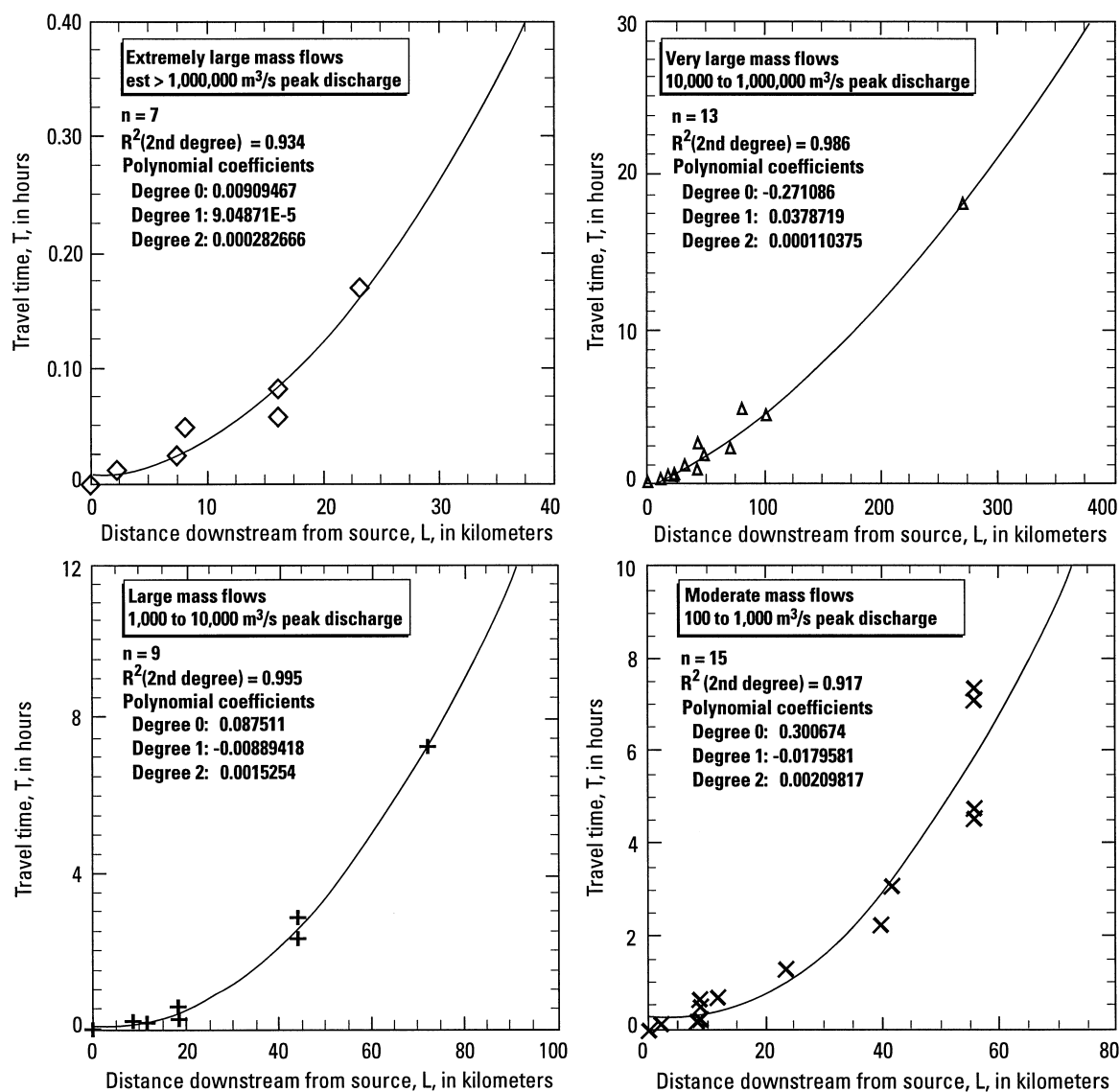


Fig. 3 Best-fit regression curves of flow-front travel time plotted as a function of distance downstream from source for four size classes of wet volcanic mass flows. Number of points in the sample (n), coefficients of determination (R^2), and coefficients of best-fit polynomial equations are shown next to each plot. See Tables 1–4 for data used in regressions (hypothetical initial-condition data pairs (0 travel time at 0 distance from source) have been added to each plot

mass flow that begins as a debris avalanche because of changes in the dominant form of energy dissipation (Iverson 1997). Therefore, the regression curve presented here for extremely large flows should not be used for lahars or for debris avalanches past the point where flow transformations may occur.

One comparison from the sample data demonstrates the effect of depth and slope. The 1980 lahars in the Pine Creek (PC) and South Fork Toutle River (SFT) valleys were triggered at the same time by the same process during the first minutes of the 18 May 1980 eruption of Mount St. Helens (Janda et al. 1981; Fairchild 1985; Pierson 1985). Both lahars were composed

of cohesionless coarse sandy gravels and both began with approximately similar velocities at the 4-km mark (SFT = 33 m/s; PC = 31 m/s). Both had approximately similar peak discharges at 4 km (SFT = 68 000 m³/s; PC = 50 000–100 000 m³/s) that attenuated rapidly downstream (SFT = 18 000 m³/s at 25 km; PC = 9 000 m³/s at 20 km). Both channels are relatively straight. The major difference was that the upper SFT channel averaged 450 m wide and flow depths averaged approximately 5 m (Fairchild 1985), whereas the PC channel averaged only 105 m wide and flow depths averaged approximately 12 m (Pierson 1985). The PC channel was also slightly steeper on average (0.05 m/m) than the equivalent reach of the SFT channel (0.04 m/m). The effect of these differences can be seen in Fig. 2, where PC travel times are consistently less (velocities faster) than SFT travel times.

The regressions in Fig. 3 can be used to forecast flow travel times at specific distances away from a potential flow source on a volcano *if* the flow-rate class of a future flow can be determined. Two approaches can be

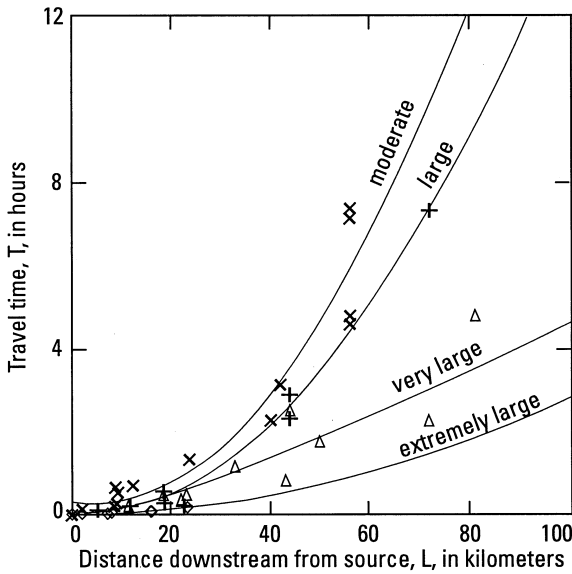


Fig. 4 Data points and regression curves for the four size classes of mass flow plotted and extrapolated for comparison at the same scale

used to make order-of-magnitude estimates of flow rate beforehand: (a) calculation of peak discharge near the volcano from measured depths and widths and backcalculated velocities of previous flows; and (b) prediction of peak discharge on the basis of expected flow volume. The first approach can be applied when the height of previous flows in channels or on valley sides is known from historical records or geologic evidence. Because stratovolcanoes commonly follow similar patterns of eruptive behavior over long periods of time, there should be a fairly high probability that future flow depths would be similar to past depths. Simple surveying of channel width, depth, and slope in a relatively straight, uniform channel reach and the application of basic hydraulic equations permit computation of approximate peak discharge (see Gallino and Pierson 1985). The second approach would be applicable when one knows or can estimate (a) the volume of rock weakened by hydrothermal alteration on a volcanic cone, (b) the volume of rock situated in an oversteepened part of the volcano, (c) the volumes of ice and snow situated in the upper parts of drainages that could be rapidly melted during an eruption, or (d) the volume of water in a lake at risk of ejection during an eruption. For the case of a rock-mass failure, the in situ rock mass should be multiplied by 1.2 to estimate the bulk volume of dilated loose debris (Voight et al. 1983). For the case of water release, the volume of water should be multiplied by 3 to obtain the approximate volume of a fully bulked debris flow (Pierson and Costa 1987). Flow peak discharge then can be predicted from flow volume by the regression equations determined by Mizuyama et al. (1992) for clay-rich and granular debris flows (Fig. 5):

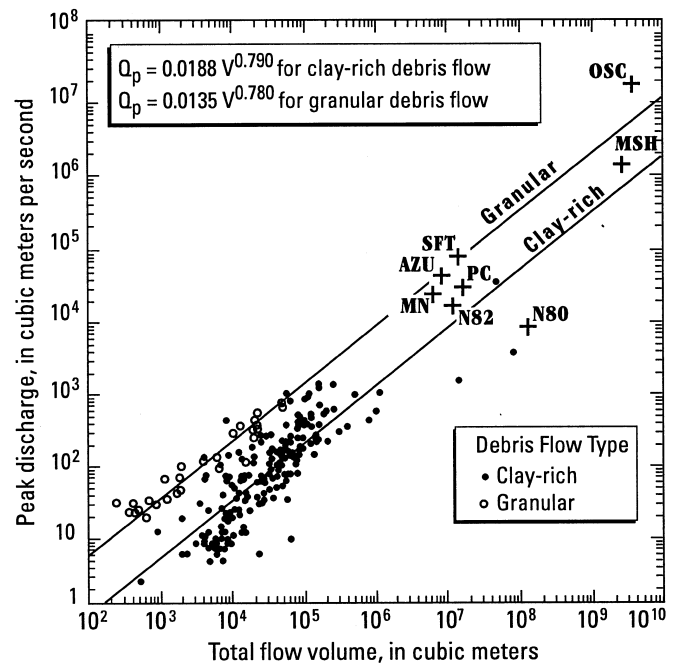


Fig. 5 Peak discharge of debris flows plotted as a function of fully bulked mass-flow volume at source. Unlabeled data points and best-fit regression lines from Mizuyama et al. (1992). Labeled data points are source volumes and near-source peak discharges for eight large to extremely large flows from this study (not used in formulating these regressions; see Table 5 for explanation)

$$Q_p = 0.0188 V^{0.790} \text{ (for clay-rich flows)}$$

and

$$Q_p = 0.0135 V^{0.780} \text{ (for granular flows),}$$

where Q_p is peak discharge and V is flow volume. Mizuyama et al. (1992) did not state whether they used near-source peak discharges or attenuated peaks from farther downstream in forming these correlations, but their data are predominantly from small to moderate mass flows where downstream attenuation of peak discharge is not great. Their data probably include a combination of near-source and downstream values. Although not used to modify the regression equations, data from eight other large to extremely large mass flows (Table 5) were also plotted in Fig. 5. Even though the curves were developed only for debris flows, a debris avalanche (MSH), a debris avalanche that transformed to a clay-rich debris flow (OSC), and a lahar that transformed from debris flow to hyperconcentrated flow (N82) generally follow the same trend. This suggests that the regression equations are adequate to predict order-of-magnitude peak discharge for all mass flows.

However, the flow-rate/volume correlation approach shown in Fig. 5 assumes a relatively instantaneous release of the debris or water mass. Protracted releases by an ongoing process will result in lower peak discharges for a given volume than would result from a relatively instantaneous release. For example, the

Table 5 Correlation of fully bulked flow volume (at source) with peak discharge (near source) for large to extremely large wet mass flows. These data were not used by Mizuyama et al. (1992) to determine regression equations, but they have the same trend (see Fig. 5)

Flow	Peak discharge (m ³ /s)	Distance from source of meas- urement point (km)	Source volume, fully bulked (10 ⁶ m ³)	Reference
AZU: Azufrado debris flow, 1985 (Nevado del Ruiz, Columbia)	48000	10	9	Pierson et al. (1990)
MN: Molinos/Nereidas debris flow, 1985 (Nevado del Ruiz, Colombia)	19000	11	6.5	Pierson et al. (1990)
PC: Pine Creek debris flow, 1980 (Mount St. Helens, Washington)	24300 (average value)	~ 10	15	Pierson (1985)
SFT: South Fork Toutle River debris flow, 1980 (Mount St. Helens, Washington)	68000	4	13	Fairchild (1985)
N80: North Fork Toutle River debris flow, 1980 (Mount St. Helens, Washington)	7200	4.5	140	Fairchild (1985)
N82: North Fork Toutle River debris flow, 1982 (Mount St. Helens, Washington)	14000	6	12	Pierson and Scott (1985)
MSH: Mount St. Helens debris aval- anche, 1980 (Washington)	~1000000	18	2800	Voight et al. (1983)
OSC: Osceola mudflow (debris flow), 5600 years BP (Mount Rainier, Washington)	12400000	10	3800	Vallance and Scott (1997)

North Fork Toutle River lahar (Tables 3, 5) had a volume that was an order of magnitude larger than that of the South Fork Toutle River lahar (Tables 2, 5), yet its peak discharge was lower and it does not plot on the same trend in Fig. 5. The North Fork lahar was generated over several hours by gradual liquefaction of parts of the just-deposited 1980 debris-avalanche deposit, whereas the South Fork lahar was triggered within minutes by rapid snowmelt and mixing with volcanoclastic debris at the beginning of the 18 May 1980 eruption.

Conclusion

Strong correlations were found between travel time and distance from source for wet volcanic mass flows of different flow-rate classes. Data from nonvolcanic flows were included for the largest class (>1000000 m³/s) due to lack of data for volcanic examples. Each class of mass flow displays a nonlinear relationship between travel time and distance from source, for which second-degree polynomial equations provide the best fits to the data. Strong correlations were obtained for sample data with coefficients of determination ranging from R^2 (second degree)=0.917–0.995, despite variability in geomorphic and material-property controls on flow velocities. The regression curves provide an empirical basis for estimating travel times of potentially catastrophic mass flows from restless volcanoes whenever order-of-magnitude estimates of flow rate (near-source peak

discharge) can be made beforehand. These relationships also provide an empirical basis for testing or calibrating predictive computer models.

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