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The dynamics of rapidly emplaced terrestrial lava flows and implications for planetary volcanism

Stephen Baloga
Proxemy Research, Inc., Laytonsville, Maryland
NASA Goddard Space Flight Center, Greenbelt, Maryland

Paul D. Spudis
Lunar and Planetary Institute, Houston, Texas

John E. Guest
University of London Observatory, London, England

Abstract. The Kaupulehu 1800–1801 lava flow of Hualalai volcano and the 1823 Keaiwa flow from the Great Crack of the Kilauea southwest rift zone had certain unusual and possibly unique properties for terrestrial basaltic lava flows. Both flows apparently had very low viscosities, high effusion rates, and uncommonly rapid rates of advance. Ultramafic xenolith nodules in the 1801 flow form stacks of cobbles with lava rinds of only millimeter thicknesses. The velocity of the lava stream in the 1801 flow was extremely high, at least 10 m s^{-1} (more than 40 km h^{-1}). Observations and geological evidence suggest similarly high velocities for the 1823 flow. The unusual eruption conditions that produced these lava flows suggest a floodlike mode of emplacement unlike that of most other present-day flows. Although considerable effort has gone into understanding the viscous fluid dynamics and thermal processes that often occur in basaltic flows, the unusual conditions prevalent for the Kaupulehu and Keaiwa flows necessitate different modeling considerations. We propose an elementary flood model for this type of lava emplacement and show that it produces consistent agreement with the overall dimensions of the flow, channel sizes, and other supporting field evidence. The reconstructed dynamics of these rapidly emplaced terrestrial lava flows provide significant insights about the nature of these eruptions and their analogs in planetary volcanism.

Introduction

The 1800–1801 eruption of the Hualalai volcano and the 1823 eruption of Kilauea volcano provide unique opportunities for investigating the dynamics of rapidly emplaced lava flows. These eruptions featured unusually high rates of discharge combined with relatively low lava viscosities, conditions thought to be present in certain planetary settings. The overall dimensions, planimetric forms, and general morphologic characteristics of these Hawaiian flows seem to suggest physical processes of emplacement that are usually considered commonplace in basaltic volcanism, for example, laminar viscous fluid flow, cooling by radiation, incipient crystallization, and the formation of an insulating surficial crust. However, it has long been known that these flows are relatively unusual among terrestrial basaltic eruptions [e.g., *Stearns*, 1926; *Richter and Murata*, 1961] in featuring such high volumetric rates of effusion and low magma viscosities.

More than 50 years ago, *Nichols* [1939] correctly pointed out that sufficiently slow-moving lava flows should be treated quantitatively by laminar viscous fluid dynamics, rather than a semiempirical formalism for turbulent water flow. Nichols also noted, in effect, the defining condition for a sufficiently slow-moving laminar flow, namely, that the movement of any filament within the flow is locally parallel to all other filaments. With noteworthy insight, *Nichols* [1939, p. 294] proclaimed that inferences about the dynamics of lava flows "can be determined only if it is known whether the flow was turbulent or laminar." A slight overstatement of the case for viscous fluid flow precedes this insightful remark, that "gravity does not succeed in making the liquid of low viscosity flow more rapidly than the more viscous liquid, because the liquid of low viscosity uses up energy in turbulence . . . ." When a free, unconfined upper surface is present, additional volumetric flow rate can always be used to increase the mean flow velocity, although this is not as efficient for turbulent flows. Inferences drawn by Nichols for the Alika flow of the 1919 Mauna Loa eruption and a distal segment of the McCartys flow in New Mexico included a laminar flow regime and flow advance rates of about 2 m/s and 0.2 m/s, respectively.

Numerous works treating lava emplacement with the concept of laminar, viscous fluid flow have appeared since Nichols' publication. With the aid of more detailed observations of basaltic lava flow emplacement, we know today that the concept of laminar flow with all elements moving in parallel is certainly not always valid. In addition, advance rates for many basaltic lava flows have been significantly greater than the range of values inferred by Nichols.

The classical view of turbulence in fluid flow features statistically correlated, time-dependent fluctuations in the components of flow velocity. We still cannot make the assertion of turbulence in lava flows in the classical sense. However, pure

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laminar flow is clearly somewhat of a special case due to mixing, disruption of streamlines, irregular flowbed topography, breaks in slope, and other factors [see *Crisp and Baloga*, 1994]. It is unfortunate that the term "disrupted flow" [*Daugherty and Franzini*, 1965] has never been widely accepted in the scientific literature for naturally occurring conditions between pure laminar flow and classical turbulence.

Field evidence and several critical observations indicate that the emplacement of the 1800–1801 Hualalai and 1823 Kilauea flows resembled a rapid flooding by lava, rather than a relatively slow-moving, laminar advance of a cooling and crystallizing viscous fluid. In this work we resurrect an approach that has resided in desuetude for more than half a century and propose a floodlike model for low-viscosity, high effusion rate flows. We show that a reconstruction of the flow dynamics provides quantitative and qualitative results that are consistent with several types of field evidence. The dynamics of such flows are scaled for gravity differences to illustrate the influence of a planetary setting on these unique terrestrial analogs.

Geological Observations of the 1801 and 1823 Lava Flows

Reconstructing the dynamics of flow emplacement requires several types of geological observations to ascertain the important physical processes that governed flow behavior. These observations include the fluid behavior of the molten lava at various stations along the path of the flow, field evidence on flow behavior and fluidity, and dimensional and morphologic considerations. Details of the known and inferred properties of the flows considered here are summarized below.

The 1801 Hualalai Lava Flow

The geology of this flow, also known as the Kaupulehu flow, is described in detail by *Richter and Murata* [1961], *McGetchin and Eichelberger* [1975], *Jackson et al.* [1981], and *Guest et al.* [1995]. Briefly, it is an alkalic basaltic aa flow, having abundant channels, erupted from a fissure zone at about 1700 m elevation along the northwest rift of Hualalai volcano (Figure 1). The flow trends down slope to the north for a distance of about 15 km to the sea and, from there, about another 5–6 km into the ocean. The remaining deposit of lava is relatively thin at the margins of the flow with thicknesses of 2–3 m near its source vent and thickening only slightly at the margins to about 5–8 m near the ocean.

The 1801 flow is remarkable for its beds of xenolith nodules [*Richter and Murata*, 1961]. The xenoliths appear to have been emplaced as a cobble lag deposit and, in some places, as an overbank levee deposit [*Guest et al.*, 1995]. Rinds of lava on the nodules (Figure 2) indicate that the effective viscosity of the host lava during nodule emplacement was extremely low, of the order of 10^2–10^3 P (10^1–10^2 Pa s) [*McGetchin et al.*, 1976]. Drainage is similar at all locations along the path of the flow where nodule beds are exposed. There is no evidence from the rinds for a significant systematic change in viscosity, or other rheologic parameters, with distance from the vent.

The flow traversed the 15 km from the vent to the sea extremely rapidly. The timescale for the development of a significant crust is typically tens of minutes to a few hours [*Crisp and Baloga*, 1990]. The lack of any significant crust on the flow suggests an emplacement time of that order. This inference is also supported by the documented time of approximately 1 hour quoted by *Ellis* [1842]. *Guest et al.* [1995] independently estimate a flow velocity near the vent of at least 10 m s^{-1}, based upon the uphill flow of lava around a preexisting cone.

Lava channels near the beds of xenolith nodules can be quite large, up to 80 m across and 18 m deep (Figure 3). The drainage of the nodule beds both toward the upper reach of the flow and below the highway, the presence of lava overspills, and the lack of a dramatic longitudinal increase in flow thickness, all provide evidence for a very low viscosity host fluid.

The 1823 Kilauea Lava Flow

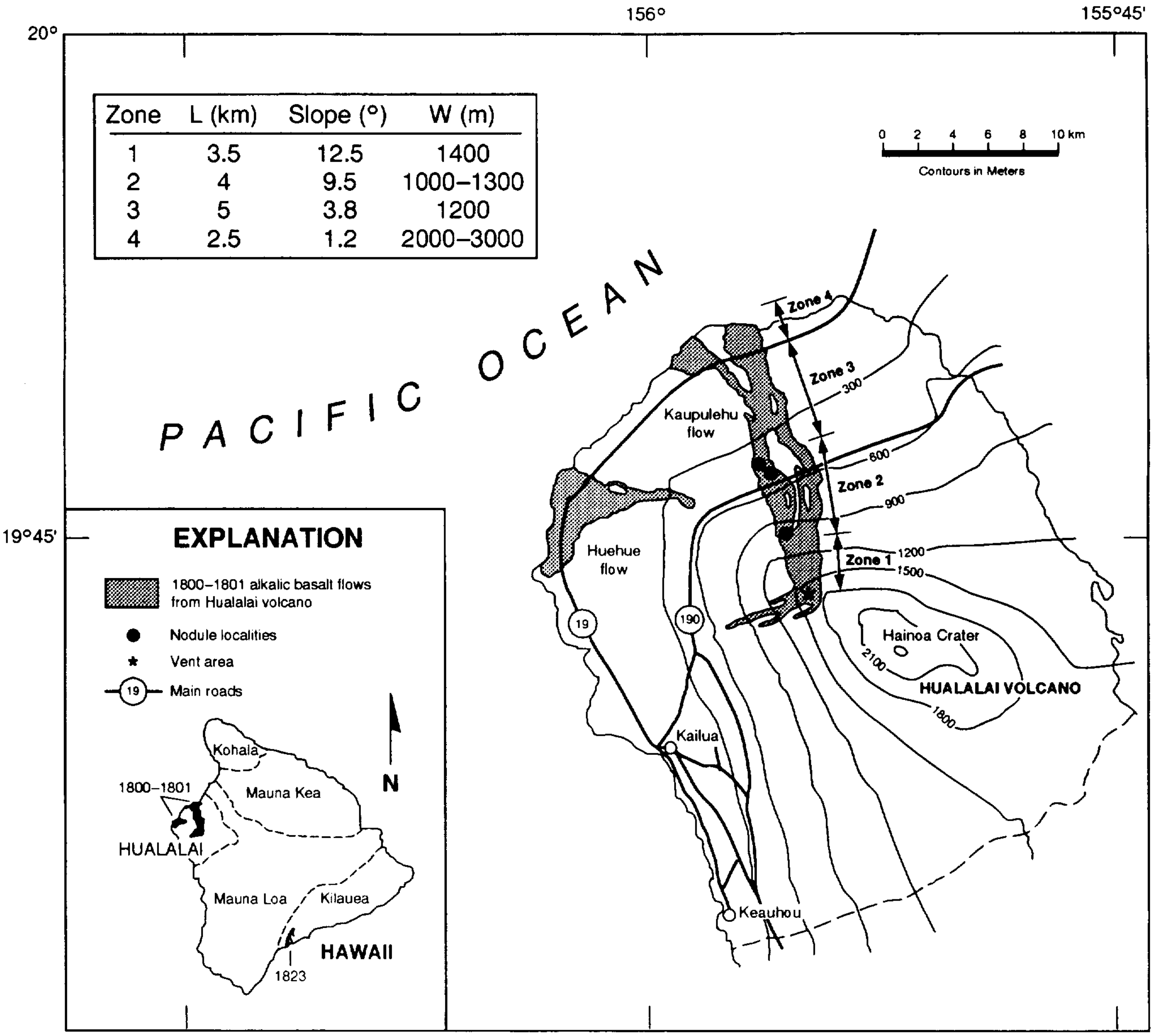
The 1823 Keaiwa lava flow from the southwest rift zone of Kilauea volcano also possesses several remarkable features that suggest unusual eruption conditions [*Stearns*, 1926]. The flow is a basaltic tholeiitic pahoehoe lava flow that changes to a slabby aa surface at its distal margins. The flow vent is a linear feature called the Great Crack and is between 1 and 10 m wide (in most places averaging about 3 m) and about 10 km in length (Figure 4). The flow near its source vent is thin, ranging from 2 to 30 cm. Nowhere does the flow thickness exceed about 2 m [*Stearns*, 1926] and 1.5 m is a reasonable estimate of the average. The flow traveled down slope at an angle to its source vent. Thus flow lengths vary from 1.5 to about 5 km. The total area of the flow is about 25 km^2, of which some unknown but probably small fraction is submarine [*Stearns*, 1926].

Observations of the eruption testify that lava issued from the Great Crack very rapidly in the form of a sheet that quickly flowed to the ocean, overwhelming a village and overtaking fleeing natives [*Ellis*, 1842; *Stearns*, 1926]. Although this flow contains no xenoliths, accretionary balls of lava display very thin (centimeter size or smaller) layers of lava plastering, suggesting a very fluid magma. During the eruption, flow advance was so rapid that the lava actually overrode a preexisting cone, as occurred near the vent of the 1801 Hualalai lava flow. The lava apparently went about 11 m up the side of a preflow cinder cone (called the lava-plastered cones [*Stearns*, 1926]). This evidence suggests velocities of flow of at least 15 m s^{-1}, comparable to the velocity inferred for the 1801 Hualalai lava flow [*Guest et al.*, 1995].

Elementary Flood Model

The low viscosity of these lava flows and their rapid rates of advance suggests a type of geologic mass movement more like a precipitous flooding of lava than a relatively slow-moving, viscous fluid flow. The residual deposits show that, for practical purposes, viscosity changes along the flow paths were insignificant in both cases. This contrasts with other basaltic flows, such as the 1A flow from the 1984 Mauna Loa eruption, that underwent a rheologic change of several orders of magnitude during the 5 days of advance [*Crisp et al.*, 1994; *Moore*, 1987]. Crystallization of microlites occurred during the advance of the Mauna Loa 1A flow, as well as the formation of a thick crust. As the viscosity increased toward the front of the flow, the thickness increased dramatically to 25 m in places. Although perhaps not as dramatic, many of the lobes of the Puu Oo eruption showed significant thickening with distance from the vent and a slowing of the advance rate that has been attributed to rheologic changes [*Fink and Zimbelman*, 1990]. Unlike the 1801 and 1823 eruptions, these flows were emplaced over periods, typically, of a few days.

For both the Mauna Loa 1A eruption and many of the lobate Pu Oo flows, the duration of emplacement was sufficiently long that a significant crustal component formed on top



Zone	L (km)	Slope (°)	W (m)
1	3.5	12.5	1400
2	4	9.5	1000–1300
3	5	3.8	1200
4	2.5	1.2	2000–3000

Figure 1. Sketch map of the island of Hawaii showing the 1801 Hualalai lava flow. After *Guest et al.* [1995]. Measurements of reach length (L), flow width (W), and slope (θ) were made from air photos and 7.5 min U.S. Geol. Survey topographic maps of the Kiholo Quadrangle and the Kailua Quadrangle.

of a hot inner core. Once such a flow becomes established, the streamlines within the flow tend to become parallel, although some mixing and disruption often occurs, particularly when time-dependent or intermittent effects are present. Cracking, mixing, and other processes expose the inner core causing an important and significant thermal loss [*Crisp and Baloga*, 1990; *Crisp et al.*, 1994; *Crisp and Baloga*, 1994], but for the most part the inner core of such slowly advancing, viscous flows remains highly insulated from radiative losses.

Field observations of the 1801 and the 1823 flows indicate that the emplacement was rapid enough to prevent a significant crustal component from building up. It is likely that a high degree of mixing continuously brought hot inner layers to the surface. However, emplacement was so rapid that neither the formation of a thick crustal layer or a marked change in the bulk viscosity was able to occur, except for lava from the very final stages of discharge.

Fluid dynamic effects observed in a variety of different geologic mass movements were present in these flows, however. Run-up is evident in both the 1801 and 1823 flows. The xenolithic nodules are most evident, as one would expect from a hydraulic type of flow, at breaks in slope [*Allen*, 1970]. The nodules themselves present the appearance of hydraulically worked cobbles in a lag deposit. The overall dimensions of the Hualalai flows show a corresponding widening as the underlying slope lessens. Field inspection provides the clear impression that the directions of flow and the locations of the emplaced margins were controlled largely by the preexisting topography as opposed to internal physical changes in the lava itself.

To describe the dynamics of these flows, we assume the local conservation of fluid volume

$$w\frac{\partial h}{\partial t} + \frac{\partial Q}{\partial x} = 0 \tag{1}$$

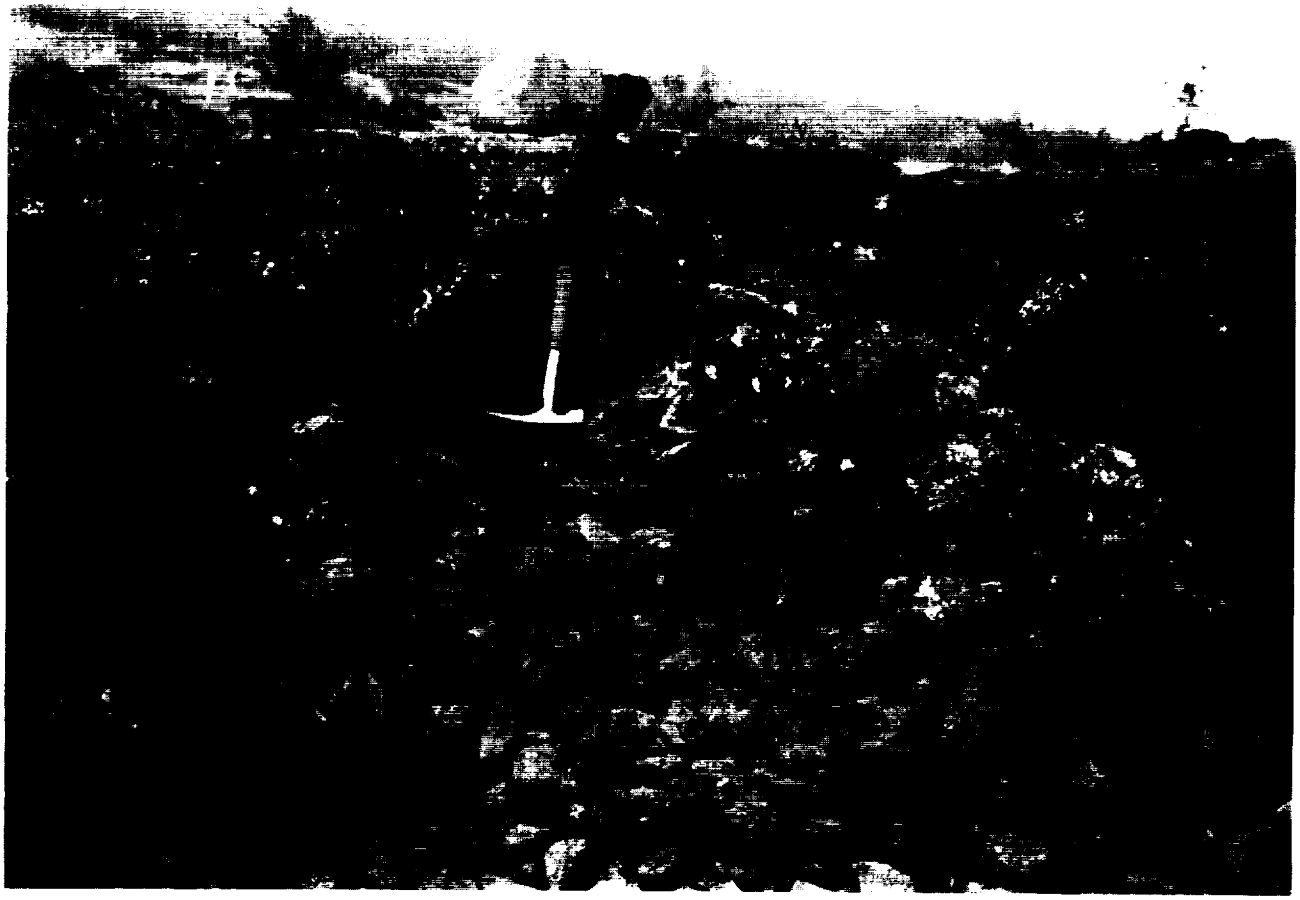


Figure 2. Xenolith nodules found in the 1801 Hualalai lava flow. Thin rinds (about 1 mm or less) of host lava on the nodules suggests an extremely fluid, low-viscosity lava.

where h is the depth of the flow, w is the width, Q is the local volumetric flow rate, x represents the distance along the path of the flow, and t is time. The volumetric flow rate is the product of the cross-sectional area of the flow times the mean velocity. Equation (1) simply requires any spatial change in the local flow rate to be compensated by a local thickening of the flow with time.

The generic statement of volume conservation in (1) provides little practical value until the form of the volumetric flow rate is explicitly determined by theoretical or empirical means. Generally, a flowrate prescription features a dependence on the underlying slope, the depth of the flow, gravity, and parameters that characterize the resistance of the medium to flow (e.g., viscosity and yield strength).

In spite of the field evidence for a highly fluid lava, one might conjecture an average flow velocity of the form

$$u = \frac{1}{3}\frac{\rho g \sin \theta h^2}{\mu} \tag{2}$$

characteristic of a laminar Newtonian flow, where ρ is the density, g is gravity, θ is the slope, and μ is the viscosity. Equation (2) in effect defines a local flowrate needed to make (1) an explicit differential equation for h. Applications of (2) and (1) to lava flows have been presented in the literature [e.g., *Baloga and Pieri*, 1986].

Fluid flow governed by (2) may be characterized by the dimensionless Reynolds number

$$Re = \frac{\rho u h}{\mu}. \tag{3}$$

As a reference, laboratory studies of fluid flow in pipes have shown that the transition from laminar flow to classical turbulence occurs for Reynolds numbers in the range from 2000 to 3000. For higher values, the onset of classical turbulence is difficult to prevent, regardless of the geometry. For lower values, the flow has insufficient energy to maintain turbulence.

We can examine the plausibility of (2) for the 1801 flow using the slope measurements that appear in Table 1. Field observations suggest a minimum thickness in the central part of the flow of at least 5 m. *McGetchin and Eichelberger* [1975] and *McGetchin et al.* [1976] arrived at an eruption viscosity of 10^1–10^2 Pa s using petrologic constraints and a range of plausible eruption temperatures and volatile contents.

Table 2 shows hypothetical flow velocities given by (2) and corresponding Reynolds numbers for the 1801 flow using a conservative estimate of 5 m for the flow depth, 10^2 Pa s for the viscosity, and a density of 2.6×10^3 kg/m^3. These elementary computations clearly indicate that (2) is a poor choice. The extremely high flow velocities and Reynolds numbers are grossly inconsistent with the concept of a laminar Newtonian fluid flow, at least for describing the large-scale features of the 1801 flow.

Many geologic mass movements and fast-moving fluids bearing various kinds of loadings or suspensions exhibit a non-

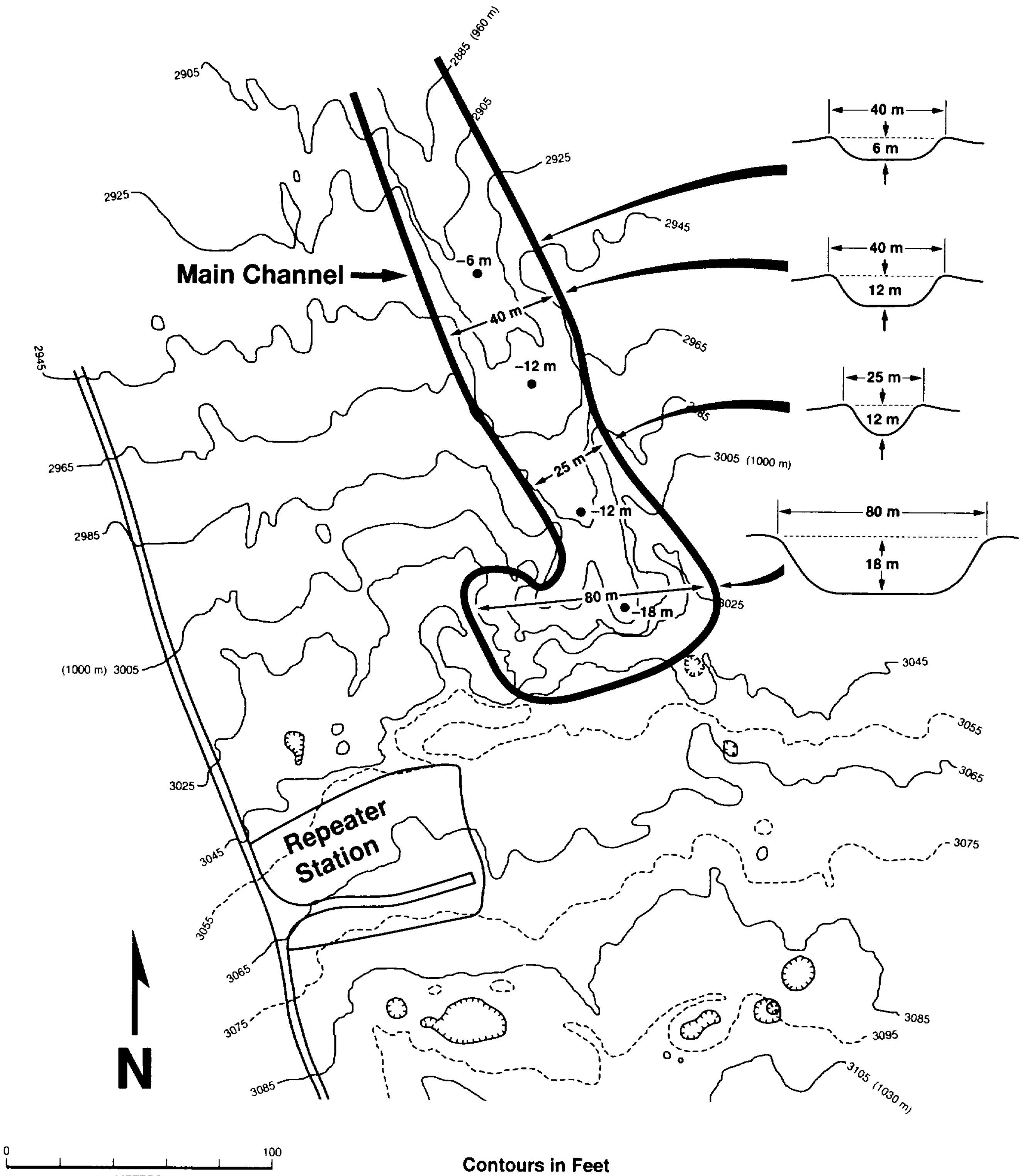


Figure 3. Topographic map of the area near the repeater station showing the dimensions of a huge lava channel in which bedded layers of xenolith nodules are found. Channel dimensions taken from map, supplemented by field measurements. Topography from *Jackson and Clague* [1982].

Newtonian or quasi-Newtonian behavior with a flow velocity given approximately by

$$u = \sqrt{\frac{hg \sin \theta}{C_f}} \tag{4}$$

where C_f is a dimensionless friction coefficient that is determined empirically, or quasi-empirically, and may depend somewhat on the nature of the suspended material and the roughness of the underlying flowbed or the lateral confinement.

For simplicity, we have expressed the mean velocity in a

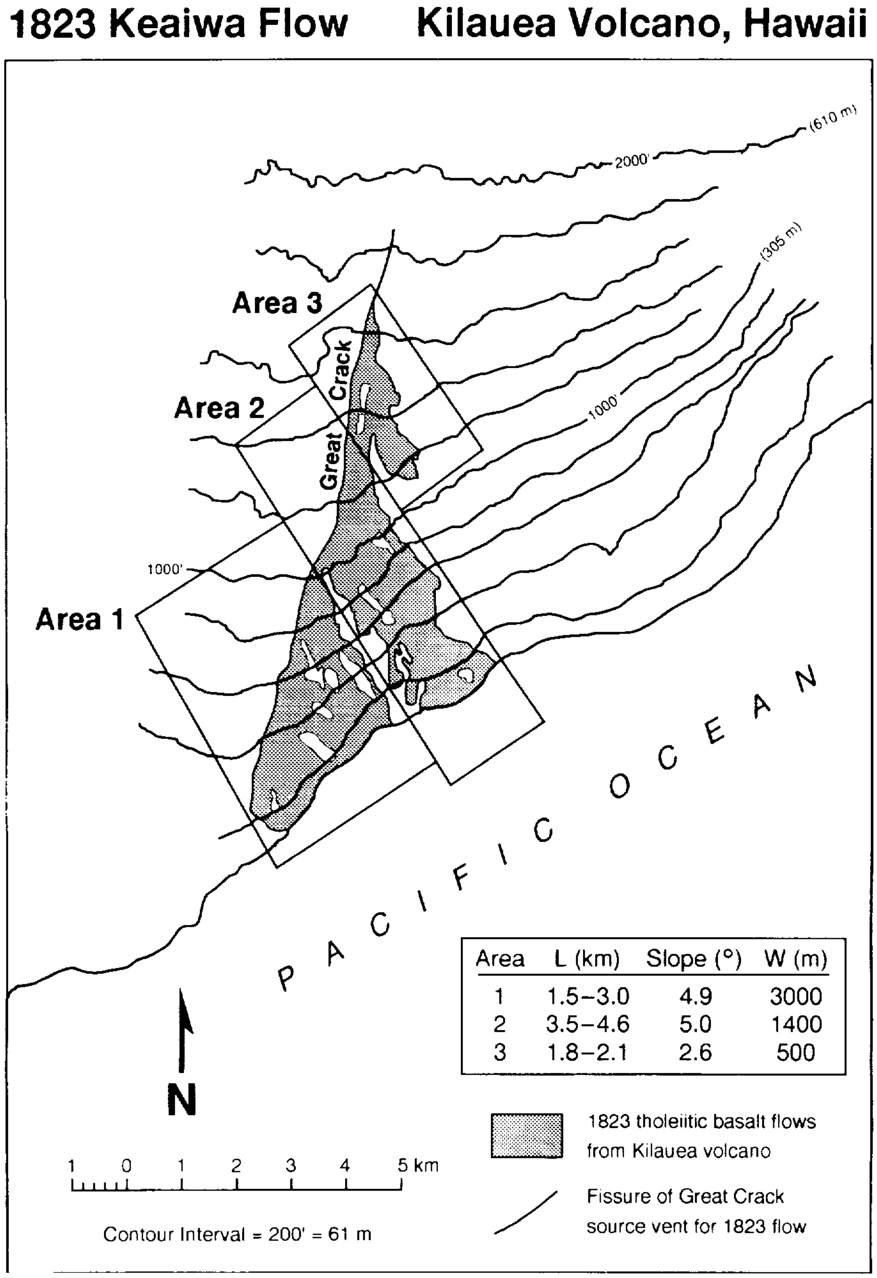


Figure 4. Sketch map of the 1823 Kilauea lava flow. Flow erupted from long fissure vent (Great Crack) and was very fluid. Dimensional measurements were made from the 1:100,000 U.S. Geological Survey topographic map: "Hawaii Volcanoes National Park and Vicinity" (1986). Estimates of the flow velocity reported by *Guest et al.* [1995] were made near the northernmost kipuka in area 2.

conventional form [see *Komar*, 1980, equation (6)] that makes C_f dimensionless. Some authors [e.g., *Daugherty and Franzini*, 1965] use a slightly different convention featuring a numerical prefactor, usually 2, in the numerator of (4), but C_f remains dimensionless. Other specifications like (4) feature explicitly physical variables related to the resistance to flow, such as the average density or radius of transported particles, thus making C_f a dimensioned constant. Consequently, some care is required in comparing the flow resistance parameters of different geologic mass movements.

This general form of the flow rate has appeared in the scientific literature for over 200 years in connection with

Table 1. Dimensional Data and Computed Changes for the 1801 Hualalai Lava Flow

Zone	L, km	W, m	Slope, deg	h/h_r	u/u_r
1	3.5	1400	12.5	1	1
2	4	1150	9.5	1.25	0.98
3	5	1200	3.8	1.64	0.71
4	2.5	2500	1.2	1.48	0.38

See Figure 1 for definition of zones. Ratios of flow thickness (h/h_r) and flow velocity (u/u_r), relative to reference values for Zone 1, are computed from the lava flood model.

Table 2. Hypothetical Flow Velocities and Reynolds Numbers for 1801 Hualalai Flow

Zone	u, m/s	Re
1	467	6×10^4
2	358	5×10^4
3	144	2×10^4
4	45	6×10^3

Flow velocities and Reynolds numbers are inconsistent with the assumption of a laminar Newtonian flow. See text for method of computation and parameters.

Chezy's formulation of flooding by water in a confined channel. Nevertheless, modern treatments of the complex geologic materials in lahars, debris flows, and mud flows, still embrace a formulation like (4) and try to refine the description with advances in theory or the use of empirical data. Frequently, these studies focus on adjustments to the power law dependence of the flow rate on the depth of the flow, the appropriate value for the friction coefficient, and the dependence of the friction coefficient on the dimensions of entrained particles, and similar considerations.

A small sampling of literature results provides an indication of the magnitude of the friction coefficients that might be anticipated for different types of geologic mass movements that have been modeled by a form as shown in (4). In a comparable formulation for the ascent dynamics of maars and diatremes, *McGetchin and Ullrich* [1973] have used a friction coefficient of 0.1. *Komar* [1980] cited a value of 0.005 for sediment-laden turbulent water flooding. For water flowing in smooth cement channels, *Jeffreys* [1925] obtained the empirical value of 0.0025, a remarkable result which has withstood the test of time [*Whitham*, 1974] in spite of significant advances in both hydrology and metrology. With a comparable formulation, *Weir* [1982] (see equation (10) and Table 8 in that reference, noting that C_f is approximately g/b^2) has obtained values approximately equivalent to the range 0.01 to 1 from a kinematic wave theory for the movement of lahars. For certain types of debris flows, *Takahashi* [1980] (see equation (19) and the subsequent discussion in that reference) has obtained 0.577 from empirical studies.

We propose the flow rate form shown in (4) for lava flows that were emplaced by rapid flooding, rather than laminar, viscous fluid flow. The governing time-dependent equation for the flooding of lava becomes

$$w\frac{\partial h}{\partial t} + \frac{\partial}{\partial x}\left(hw\sqrt{\frac{hg\sin\theta}{C_f}}\right) = 0. \qquad (5)$$

In general, this equation must be supplemented by an appropriate boundary condition to evaluate the influence of time-dependent changes in the effusion rate on the evolving downstream flow thickness. When detailed information about vent conditions is not available, steady state solutions may be assumed as an approximation. This makes the first term in (5) irrelevant, but we lose the ability to explain thickness changes with time at fixed stations along the path of the flow.

At present, there is no experience with the application of such a formulation to lava flows. Specifically, the friction coefficient appearing in the denominator is completely unknown and the power law describing the response of the flow rate to the depth and underlying slope is postulated. Consequently, our approach is to apply this model to the flows and compare the results with independent field checks for consistency. We will also compare the results with the experience documented in the geologic literature for other types of mass movements with similar governing equations.

Dynamics of the 1801 Flow

The 1801 eruption produced lava flows that overran four relatively distinct zones on the volcano during transit to the ocean. These zones are indicated on the sketch map appearing as Figure 1. Measurements of the length of each reach (L), the width (W), and slope (θ) were made from air photos superimposed on U.S. Geologic Survey topographic maps. As seen in Figure 1 and Table 1, the width and slope are roughly constant in each zone. We assume that the outer margins of the flow were emplaced during an early phase of the eruption and that the channels found in all reaches were formed subsequently.

Because no other information is available, the eruption rate is considered to be constant during the early phase of the eruption. With no time-dependent effects in (5), the flow rate becomes a conserved quantity ($= Q_0$) along the flow path. Thus

$$Q_0 = u_i h_i w_i \qquad (6)$$

where the subscript i indicates the value of the variable in zone i.

Equations (4) and (6) provide an immediate constraint on the ratio of the thickness of the flow in each zone given by

$$\frac{h_i}{h_r} = \left(\frac{w_r}{w_i}\right)^{2/3}\left(\frac{\sin\theta_r}{\sin\theta_i}\right)^{1/3} \qquad (7)$$

where i denotes the zone of interest and r indicates a reference zone. The reference values are taken as those of zone 1, although any zone could be chosen.

Table 1 shows the relative thicknesses in each of the zones based on the measured values of the flow widths and slopes. This type of flow model produces a longitudinal thickness profile that is relatively flat in spite of the significant changes in slope and flow width in the different zones. When the underlying slope variations and lateral spreading of the flow are taken into account, we expect only a modest thickening along the flow path according to the results in Table 1.

We can also derive the relative velocity of the flow in each of the zones by substituting (7) into (4).

$$\frac{u_i}{u_r} = \frac{h_r w_r}{h_i w_i} = (w_r \sin\theta_i / w_i \sin\theta_r)^{1/3}. \qquad (8)$$

The relative flow velocities are also shown in Table 1.

The predictions of flow thickness and velocity in Table 1 are only relative values based on volume conservation, the form of the flow rate, and geometric considerations. In effect, these predictions must be calibrated by independent information on reference values in one of the zones. If we had independent experience and knowledge of the friction coefficient, this would not be the case.

Field observations in zone 1 indicate that the flow must have been at least 5 m deep, presumably during the early stage of the eruption. By using this estimate as the reference value, the flow thicknesses in the other zones are given by (7). The numerical thickness values are shown in Table 3. Although it is difficult to establish flow thicknesses from postemplacement

Table 3. Reconstructed Dynamics for the 1801 Hualalai Lava Flow

	Zone 1 (Estimated)			
	h = 5 m	u = 5 m/s	u = 10 m/s	u = 15 m/s
Zone 2 (Computed)	6.2	4.9	9.8	14.6
Zone 3 (Computed)	8.2	3.5	7.1	10.6
Zone 4 (Computed)	7.4	1.9	3.8	5.7
Transit time, min		71	35	24
Flow rate, m^3/s		35,000	70,000	105,000
C_f		0.42	0.11	0.05

Zone 1 values for thickness and velocity are estimated as described in text. All other values are computed from the lava flood model.

conditions, we consider the values in Table 3 to be consistent with field observations.

Guest et al. [1995] have estimated a minimum local flow velocity in zone 1 to be at least 10 m/s. Using this as the reference value for (8) gives the zonal flow velocities shown in Table 3. There are uncertainties in such field estimates, so Table 3 includes alternative reference choices (5 and 15 m/s) for the flow velocity in zone 1 and the corresponding downstream velocities. Based on the observations of *Guest et al.* [1995], we consider the middle value (10 m/s) to be the best estimate of the three alternatives shown in Table 3.

Three important characteristics of the emplacement can be reconstructed from the formulation above, namely, total transit time from the vent to the ocean, the volumetric flow rate, and the friction coefficient. Values for each of these quantities with the three alternative reference velocities appear in Table 3. The transit time is computed from the relation

$$T = \sum_{i=1}^{4} \frac{L_i}{u_i}. \tag{9}$$

The best estimate of the transit time from the crater to the ocean is about 35 min, which is consistent with the recorded observations [*Ellis*, 1842]. Such a transit time is also consistent with the lack of a significant crust formed on the surface of the flow. For times much longer than an hour, we would expect to find some evidence for a significant crust in the distal parts of the flow, and this was not observed in the field.

The best estimate of the flow rate is 7×10^4 m^3/s. Even the lowest value in Table 3 is more than a full order of magnitude higher than the estimated peak of 3000 m^3 s^{-1} for the 1950 Mauna Loa eruption [*Malin*, 1980]. The volcanology literature is replete with estimates of eruption rates. However, these estimates are usually time-averaged over the entire duration of eruption, so one rarely finds values approaching the 1950 Mauna Loa value or the result we have obtained for Hualalai.

The heretofore unknown friction coefficient was determined by the equation

$$C_f = \frac{g \sin \theta h}{u^2}. \tag{10}$$

It may be recalled that the range of values shown in Table 3 for the friction coefficient is common in the literature for lahars, debris flows, and sediment-laden water floods, where similar formulations of the flow rate have been used.

Perhaps the most convincing test of this formulation concerns the unusually large channels of the Hualalai flow complex. Near the repeater station, the largest channel is approximately rectangular in shape with a width that narrows to about 40 m and a depth of 18 m (See Figure 3). Field observations indicate that this channel must have been completely filled at some point during the eruption. Moreover, there is no field evidence for a ponding or staging area along the path of the flow where lava could have accumulated prior to a catastrophic release.

We now examine the depth of flow in such a channel with the inferred eruption rates and friction coefficients. As a minimum, the channel depth must satisfy the constraint,

$$h_c = \left(\frac{Q}{w}\right)^{2/3} \left(\frac{C_f}{g \sin \theta}\right)^{1/3} \tag{11}$$

where h_c is the depth of the channel, w is the measured width of the channel, and Q and C_f are the inferred values from the previous analysis. If the entire flow rate was confined to such a channel, the computed depth of filling would be more than twice the measured depth for each of the flow rate/friction coefficient pairs shown in Table 3.

An alternative approach is to compute the flow rate and flow velocity necessary to fill such a channel. The results of these computations are shown in Table 4. The flow rate in the channel is approximately one seventh of the previous estimates, although the velocity of flow in the channel remains very high.

We might suspect the flow in the channel to be laminar late in the eruption and attempt to abandon the flood model at this stage. Using the familiar formula for the average flow velocity shown in (2) with $\rho = 2.6 \times 10^3$ kg/m^3, $g = 10$ m/s^2, $\theta = 8$ degrees, $h = 18$ m, and the conservative estimate $\mu = 100$ Pa s (1000 P), we obtain $u = 3.9 \times 10^3$ m/s and $Re = 2 \times 10^6$. For this conjecture to be remotely viable, there would have to be new evidence for a viscosity at least 2 orders of magnitude higher than the *McGetchin et al.* [1976] estimate.

We consider this to be an important check on the consistency of the previous flow rate estimates. Basaltic eruptions often wane significantly in effusion rate from relatively high peak values established early in the eruption. Channels typically form after the early phase of the eruption and it is unlikely that the entire flow rate was conducted through this solitary channel at the repeater station. Our flow rate estimate in the channel is consistent with the hypothesis of a high effusion rate early in the eruption followed by a subsequent waning.

Given the total volume of lava that can be identified and associated with this eruption, it is not likely that high flow rates (approaching 10^5 m^3/s) were maintained continuously for days

Table 4. Flow Rate and Velocity in Repeater Station Channel

C_f	Q, m^3/s	u, m/s
0.42	5.5×10^3	7.6
0.11	1.1×10^4	15.2
0.05	1.6×10^4	22.8

Using the friction coefficient of previous computations and dimensional data for channel, the flow rates (Q) and velocities (u) are required to fill the channel.

or weeks. Otherwise, a much more massive edifice would have been constructed from the 1801 eruption. The flow rate in the channel at the repeater station supports a significant drop in the flow rate from the peak value established early in the eruption.

The relatively high velocities in the channel argue strongly for a highly agitated and forceful flow that would be necessary to suspend and transport the nodule load. Evidently, this occurred with such a highly mixed and agitated character that significant internal temperature gradients were prevented during the rapid delivery to the distal reaches of the flow. Indeed, the deposition of nodules in the deep channels supports the continued drop in the effusion rate as the eruption waned.

The cause of the unusually high volumetric flow rate featured by this eruption remains somewhat unclear at present. *Guest et al.* [1995] have attributed the high rate of discharge to the near-Newtonian rheologic properties of the host lava during both eruption and emplacement. Alternatively, a summit crater about 250 m in diameter may have served as a reservoir for molten lava. Field evidence clearly suggests that crater wall collapse supplied at least some of exposed lava flows.

The overall dimensions of the flow and the behavior of the eruption are consistent with a dynamic model of lava flooding. The volumetric effusion rate during the early stage of eruption appears to have been between 10^4 and 10^5 m^3/s. This rate is almost 2 orders of magnitude higher than the peak of the 1950 Mauna Loa eruption and is comparable to eruption rates conjectured in the literature for large planetary lava flows. The transit times from the vent to the ocean are also consistent with the lack of a surficial crust, the low viscosity of the host magma in the distal reaches, and the historical comments on the speed of advance of the flow. It is clear that such high effusion rates were not maintained for a long duration. The drop from the high values shown in Table 3 is supported by the significantly lower volumetric flow rates in the channel, as shown in Table 4. The waning character of the effusion rate is also supported by the deposition of nodules exposed in the channels.

Dynamics of the 1823 Flow

The 1823 flow can be analyzed in a similar manner. Figure 4 shows a sketch map of the flow complex and suggests treating the emplacement as three separate and distinct flows emerging from different parts of the fissure. In each area the flow deposits each have a single width and slope along the path of the flow from the Great Crack to the ocean. Flow continuity between the three areas cannot be applied as was done for the 1801 eruption because the lava did not advance from one area to the next. Consequently, the dynamics of emplacement for the 1823 flow cannot be reconstructed in as great detail.

Guest et al. [1995] have estimated a flow velocity of at least 15 m/s in area 2 about a kilometer from the Great Crack. First, we compute the friction coefficient from the formula

$$C_f = \frac{g \sin \theta h}{u^2} \tag{12}$$

with the mean flow thickness taken as 1.5 m and the average slope as shown in Figure 4. This gives $C_f = 0.0057$, a value similar to that of sediment-laden flooding by water [*Komar*, 1980]. If the estimated flow velocity is conservatively decreased to 10 m/s, the friction coefficient becomes 0.0128, which is more typical of water.

The two velocity estimates (15 and 10 m/s) for area 2 correspond to transit times of 4.4 min and 6.7 min to the ocean. The flow velocity must have remained essentially constant along the flow path, as far as we know. The rapid emplacement must have resembled an advancing, highly turbulent, floodlike sheet. There was very little time for a crust or skin to form. The thinness and fluidity of this flow suggests that it would naturally follow the small-scale topography, thus mixing any cooler skin into the interior.

For area 2 we find the eruption rates of 21,000 and 14,000 m^3/s for these two velocities estimates (15 and 10 m/s). These rates seem very high by present-day experience, but this is primarily due to the extent of the fissure (e.g., 1400 m parallel to the direction of flow) and the lateral width of the flow.

In area 1 the slope is only marginally different from that of area 2. The same flow velocity estimates are likely to be reasonable for this area as well. This part of the flow complex probably added another 45,000 m^3/s or 30,000 m^3/s to the eruption rate for estimated flow velocities of 15 m/s and 10 m/s, respectively.

Toward the upper part of the Great Crack in area 3, the slope is about a factor of 2 lower. Here the flood model can be used to estimate the flow velocity of area 3. Based on the two friction coefficients derived above, we obtain estimates of 10.8 and 7.2 m/s, respectively, in area 3. These results are also based on an average depth of 1.5 m. These velocity estimates correspond to flow rates of 8100 and 5400 m^3/s for area 3.

The flood model is hardly essential for these computations. They are based primarily on the mapping of the flow, geometric considerations, and the field estimate of the velocity in one of the areas. We have used the model only to compute the friction coefficient and the velocity in area 3, which is a rather minor contributor to the eruption rate.

The model does provide one very important aspect of support for these considerations. Specifically, the friction coefficient is about 10 or 20 times less than that obtained for the 1801 flow. Conceptually, this is qualitatively consistent with the lack of an appreciable load in the 1823 magma and an emplacement that is even more rapid than that of the Hualalai flows.

The conclusion we obtain for the 1823 flow is that it also had a very high eruption rate, probably at least 21,000 m^3/s. If the three segments of the fissure were fed simultaneously, which appears likely, the eruption rate could have been roughly 3 times as high. As with the 1801 flow, this high eruption rate probably persisted only for a short duration, perhaps on the order of tens of minutes or less.

Scaling the Model for Planetary Applications

We now consider some implications of the lava flooding model for volcanism in a planetary setting. As a basis of comparison, we consider the volumetric flow rates of a laminar Newtonian fluid (equation (13)) and the lava flood model (equation (14)):

$$Q = \frac{g \sin \theta h^3 W}{3\nu} \tag{13}$$

$$Q = \sqrt{\frac{g \sin \theta h}{C_f}}\, hW. \tag{14}$$

To compare terrestrial analogs with lava flows in a planetary setting, we suppose that the eruption rate, preexisting topography, and rheological properties remain the same in both settings. Our chief aim is thus to examine the influence due solely to changes in gravity.

The following example provides a comparison between a lava flow on the Earth and the Moon. For either of the models shown in (13) and (14), we obtain the relation

$$g_e h_e^3 = g_m h_m^3. \tag{15}$$

A priori, it is somewhat surprising that (15) holds for both the Newtonian flow rate based on a laminar flow and the flood model that is more characteristic of disrupted, possibly suspension-laden, turbulent flow. By rearranging (15), and substituting values for the gravity of the Earth and Moon, we find

$$h_m = (g_e/g_m)^{1/3} h_e = 1.82 h_e. \tag{16}$$

A comparable flow on the Moon has a thickness almost twice that of its terrestrial counterpart.

Why lower gravity produces systematically thicker flows on planetary bodies with lesser gravity is evident from the corresponding velocity scaling:

$$u_m = (g_m/g_e)^{1/3} u_e = \frac{u_e}{1.82}. \tag{17}$$

Due to the lesser gravity on the Moon, a comparable flow would advance almost twice as slowly. For the same eruption rate and preexisting topography, the slower rate of advance is the cause of the corresponding thickness increase. Equation (17) is valid for both of the flow rate models under consideration.

These considerations suggest that the 1801 Hualalai flow could indeed be quantitatively similar to lunar eruptions that produced the lava flows at Mare Imbrium and other lunar locations. Doubling the margin thicknesses of the 1801 flows to account for the lower gravity produces thicknesses comparable to those of certain flow lobes in Mare Imbrium and elsewhere on the lunar maria [*Schaber et al.*, 1976].

At one point during the Hualalai eruption, the central depth in channels was at least 18 m. Comparable, but sustained, eruption rates on the Moon or Mars could produce flows with thicknesses between 10 and 30 m, just due to the lower value of gravity. Flow thicknesses in this range are abundant on the plains and shield volcanos on Mars and the lava flow lobes in Mare Imbrium. Unlike the Hualalai eruption, however, the volumes of many of the planetary flows are at least an order of magnitude larger than those of the Hualalai flow complex. The lava flood model suggests that one primary difference is that the duration of planetary eruptions of this style persisted for orders of magnitude longer than their terrestrial analogs.

The high fluidities and high effusion rates of these terrestrial eruptions might be analogous to conditions that once existed on Mars, the Moon, and perhaps Io and Venus. Most lava flows that have been identified on these surfaces are probably 1 or 2 orders of magnitude thicker than the Keiawa lava flows, although some authors have argued for similar eruption rates. Our knowledge of planetary volcanism is limited by the resolution of available images. Flows analogous to those at Keiawa may indeed exist in numerous planetary settings, but they would not be discernable with existing data. Flows with a relief comparable to the 1823 flows, even with adjustments for gravitational scaling, would go largely undetected, even on the Moon. However, some thin lunar flows with vertical relief of the order of meters have been suggested by *Greeley* [1975]. This consideration has been thought to be the basis for the scarcity of mappable flow lobes on the young lunar maria for some time [*Schaber et al.*, 1976].

When plan form, vertical relief, and topographic slope could be distinguished for lava flows on planetary surfaces, many authors have used models based on laminar flow to infer eruption, emplacement, and rheologic conditions [e.g., *Moore and Schaber*, 1975; *Zimbelman*, 1985; *Cattermole*, 1990; *Lopes-Gautier*; 1993, and references therein]. For the rapidly emplaced terrestrial analogs considered in this work, we have the benefit of knowing that they were rapidly emplaced. Plan form and vertical relief comparable to available planetary data would be of little assistance in choosing between a more conventional laminar model and the rapid emplacement model we have employed. However, future studies will investigate the existence of morphologic diagnostics indicative of a rapid emplacement phase in the early stage of a planetary eruption.

Conclusions

The planimetric form, dimensions, and general morphologic character of basaltic flows do not guarantee that viscous fluid dynamics for laminar flow are appropriate for describing the essential character of emplacement. We have evoked an elementary lava flooding model, similar to ones used for a wide variety of geologic mass movements, to explain the emplacement of low-viscosity lavas from eruptions with high effusion rates. Application of this approach to the 1801 flow of Hualalai and the 1823 flow of Kilauea produces quantitative results that are consistent with several types of field observations. The inferred eruption rates for these flows are rather high by present-day experience, ranging from 10^3 to 10^5 m^3/s. The friction coefficients obtained by this analysis are of the order of 10^{-2} to 10^{-1}, generally in the range anticipated from other types of geologic mass movements.

The 1801 Hualalai and 1823 Kilauea eruptions may indeed be terrestrial analogs for many planetary eruptions due to the high fluidity and eruption rates. Our analysis suggests that the eruptions of these terrestrial analogs must have been very brief by planetary standards. Scaling of the terrestrial analogs for gravitational changes indicates that flow thicknesses would roughly double on the Moon, Mars, Mercury, and Io, with corresponding decreases in flow advance velocities. Judging by the large volumes of discernable lava flows on planetary surfaces, the durations of eruptions must have been orders of magnitude longer than these unique terrestrial counterparts.

The transport model appearing in (1) relies on the assumption that the volume of a fluid element is a conserved quantity during emplacement. Various field and theoretical studies indicate that volume conservation may be only an approximation during the emplacement of a flow. Different aspects of this issue are given by, for example, *Moore* [1987], *Moore* [1982], *Stasiuk et al.* [1993], and *Wadge* [1981]. One natural extension of the model presented here is the addition of a rate term to (1) to describe differences between the erupted and postemplacement volumes. We will reserve this extension for future studies.

Rapidly moving, highly fluid terrestrial lava flows may be less rare than is commonly thought. An unusually fast-moving, highly fluid flow with yet another composition was associated with the 1977 eruption of the Mount Niragongo volcano in Zaire [*Tazieff*, 1977]. A perched lava lake of nephelinite composition was very rapidly drained in less than an hour. This massive effusion resulted in flows with average velocity estimates of about 10–15 m/s and velocities early in the drainage may have been significantly higher. This eruption may have been analogous to the 1800–1801 Kaupulehu eruption in that lava flows appear to have been fed, at least in part, by the rapid draining of a topographically perched lava lake.

We expect the lava flooding approach to be appropriate for other fluid lava flows that are emplaced too rapidly for laminar flow conditions to prevail. When a transit time or an independent flow velocity estimate can be obtained, morphologic data can be used to reconstruct the dynamics of the flow, as shown in this work. When a transit time or an independent velocity estimate cannot be obtained, a crude reconstruction of flow dynamics may be conjectured by using the friction coefficients obtained here and in future applications.

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S. Baloga (corresponding author), Proxemy Research, 20528 Farcroft Lane, Laytonsville, MD 20882.

J. E. Guest, University of London Observatory, Mill Hill Road, London NW7 4SD, England.

P. D. Spudis, Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston, TX 77058.

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