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Some physical requirements for the emplacement of long basaltic lava flows

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Abstract. Long basaltic lava flows (over 100 km in length) require specific emplacement conditions to prevent the lava from freezing as it is transported to the flow front. The minimum dimensions of the lava transport systems (tubes, channels, or sheets) require that the flow have a volume greater than several cubic kilometers. Long lava flows are emplaced on slopes less than 10% (~5°) and the lava being transported must cool at a rate less than 0.5°C/km. We show that there are two modes by which thermally efficient, long distance lava transport can be achieved: (1) "rapid" emplacement in which the lava flows so quickly that it does not cool excessively despite large heat losses and (2) "insulated" emplacement in which heat loss is minimized. We here estimate cooling in the rapid mode using a modified version of a previously published thermal model for aa flows and find that, for a range of inputs appropriate for subaerial terrestrial condition, effusion rates of at least 3100 to 11000 m^3/s, channel flow velocities in excess of 4 - 12 m/s, and minimum channel depths of 3 - 17 m are required for basaltic flows >100 km in length. For emplacement in the insulated mode, we construct a very simple heat balance model for roofed sheet flows which shows that extremely long sheet-fed flows are possible with velocities as low as 0.2-1.4 m/s, flow thickness of 6 - 23 m, and minimum effusion rates of the order of 50 - 7100 m^3/s. Also, earlier work has suggested that tube-fed flows more than 100 km long can be produced at effusion rates as low as several tens of m3/s and with tube diameters of a few tens of meters. We argue that flows emplaced in the rapid mode should be morphologically similar to channel-fed aa flows while those emplaced in the insulated mode should be similar to tube-fed or sheet-like inflated pahoehoe flows. This leads to several field criteria for distinguishing these two modes of emplacement in ancient lava sequences. Additional constraints on the emplacement of long lava flows are expected from the continued study of the formation and evolution of lava channels, tubes, and sheets.

1. Introduction

Very long lava flows have attracted a great deal of interest over the years. The sheer size of these flows implies that they were produced by eruptions that were in some way outside the scope of those recorded in human history. However, long lava flows occur in a variety of geologic settings and throughout geologic time (Table 1), making it difficult to plead for truly extraordinary eruptive or environmental conditions.

In this paper we attempt to quantify some of the physical conditions necessary for basaltic lava to flow more than 100 km from a source vent. There have been a number of previous investigations that directly and indirectly tackle this problem [e.g., *Shaw and Swanson*, 1970; *Danes*, 1972; *Walker* 1973; *Hulme* 1974; *Malin*, 1980; *Head and Wilson*, 1986; *Pieri and Baloga*, 1986; *Pinkerton and Wilson*, 1994; *Wilson and Head*, 1994]. While we build upon the foundation of these earlier studies, we also attempt to delve into the most basic assumptions that underlie lava flow models. Our goals in this paper are to discuss and quantify (1) the most fundamental requirements for long lava flows, (2) the dynamics of basaltic lava transport systems, (3) the thermal efficiency of these transport systems, and (4) the morphologic features that can be used to identify the different modes in which long lava flows can be emplaced.

1.1. Basic Constraints

There are two extremely basic requirements for a long lava flow: (1) the lava must flow a great distance before freezing and (2) sufficient lava must be erupted to cover a large area. *Pinkerton and Wilson* [1994] used the terms "cooling-limited" and "volume-limited" to describe flows that stop by failing to meet the first or second requirement, respectively. While these requirements may be obvious, quantifying them is not trivial.

It is often assumed that the first requirement is the most difficult to meet. As such, defining the conditions under which lava can be transported a hundred or more kilometers without excessive cooling should place the most rigorous constraints on the emplacement and eruption styles of long lava flows. Typical basalts begin to behave as solids after 50-60% crystallization and ~100°C of cooling [e.g., *Shaw et al.*, 1968; *Shaw*, 1969; *Marsh*, 1981; *Hon et al.*, 1994]. However, after only about 50°C of cooling and 20-30% crystallization, the effective viscosity of the lava begins to rapidly increase and develops a complex non-Newtonian rheology [e.g., *Shaw*,

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Table 1. Examples of Intermediate and Long Lava Flows

Flow	Age	Length, (km)	Volume, (km^3)	Slope, (%)	Surface	Reference
Mauna Loa[a]	1859	51[b]	0.27	5	aa	*Rowland and Walker*, [1990]
Mauna Loa	1859	47[b]	0.11	5	phh[c]	*Rowland and Walker*, [1990]
Laki	1783	65	14.7	1-2	phh&aa (~50/50%)	*Thordarson and Self*, [1993]
Eldgja	934	70[b]	15	1	phh&aa (~70/30%)	*Miller et al.*, [1996]
Carrizozo	1-2Ka	75	4.5	0.5	phh	*Allen*, [1952]
Toomba	~13Ka	120	12	0.4	phh	*Stephenson et al.*, [1996]
Thjorsa	8.5Ka	140[b]	21	~0.7	phh-slab phh	*Vilmundardottir*, [1997]
Undara	~190Ka	160	~25	0.5	phh	*Stephenson et al.*, [1996]
Pomona	12Ma	>600[b]	760	~0.1	not preserved	*Tolan et al.*, [1989]
Rosalia	14.5Ma	550	>2000	~0.1	rubbly phh	*Tolan et al.*, [1989]
Roza	~15Ma	350	1300	~0.1	phh	*Tolan et al.*, [1989]

[a]Note that the longest pure aa flow, the 1859 Mauna Loa aa phase, is too small to be considered "long" in this paper. The Laki, Eldgja, and Carrizozo flows are also less that 100 km in length but share many characteristics of long flows (e.g., volumes of several cubic kilometers).

[b]Reported length is limited by the lava reaching the ocean.

[c]Surface abbreviation phh is pahoehoe.

1969; *McBirney and Murase*, 1984; *Chester et al.*, 1985; *Pinkerton and Stevenson*, 1992]. Because of this, we argue that after about 50°C of cooling a flow will rapidly slow and not advance much further before solidifying. This is supported by observations. The most crystalline moving lava sampled from the 1984 Mauna Loa aa lava flow has 39 vol% crystals [*Crisp et al.*, 1994]. Similarly, the most crystalline quenched pahoehoe margins from both the Carrizozo and Roza flow fields contain ~30 vol% crystals [*Keszthelyi and Pieri*, 1993; *Thordarson*, 1995]. We use the requirement of <50°C of cooling over 100 km (<0.5°C/km) as the key constraint on our subsequent modeling. In two cases when the cooling of long flows has been estimated, both show cooling rates more than an order of magnitude lower than this value [*Thordarson and Self*, 1996; *Ho and Cashman*, 1997], so we feel that our model constraints will not be excessively restrictive.

The requirement of a large erupted volume for a long lava flow also provides important constraints. Even a 1-m-thick, 100-km-long, and 10-km-wide flow has a volume of 1 km^3. The models we present in this paper suggest that the minimum thickness for a long basaltic lava flow is on the order of 10 m, and thus we predict that long lava flows should have volumes greater than 10 km^3. Not surprisingly, the volumes of long lava flows are indeed in excess of 10 km^3 (Table 1). While we do not examine magma genesis and rise in this paper, we suggest that the requirement of producing >10 km^3 of lava in a single eruption is the single most important reason why long lava flows are rare. We note that only four basaltic flows >1 km^3 in volume have been produced this century [*Simkin and Siebert*, 1994].

1.2. Empirical Constraints

In addition to these two requirements, empirical observations also suggest that (1) long lava flows traverse average slopes ≤10% and (2) long lava flows have lava transport systems that exceed 90% of the total flow length.

The statement that long lava flows must be emplaced on slopes less than 10% is a simple reflection of topography over length scales of >100 km. Even the largest volcano in our Solar System, Olympus Mons on Mars, has average slopes of only 5-6° (~10%) [e.g., *Pike*, 1978]. In fact, Table 1 shows that long terrestrial lava flows appear to be emplaced on average slopes considerably less than 0.5° (1%). It is also worth noting that the effect of lower gravity on the other terrestrial planetary bodies is the same as a shallower slope on Earth. This observation places strong constraints on the dynamics and maximum flow velocities of long lava flows.

The assertion that the lava transport system dominates long lava flows is supported by observations of both active and ancient flows. Lava flows can generally be broken into three segments: (1) vent region (ponds, fissures, cones, etc.), (2) transport system (channels, tubes, or sheets), and (3) flow front (or "snout" *Kilburn* [1996]). With the exception of fissures, the lateral scale of both basaltic vents and flow fronts are commonly on the order of tens to hundreds of meters. While fissure systems can extend over 150 km [e.g., *Swanson et al.*, 1975], in this paper a flow is only considered "long" if it has flowed >100 km from its point of eruption. Thus the lava transport system must extend >90% of the length of a long basaltic lava flow. The goal of this paper is to examine how lava tubes, channels, and sheets can transport lava >100 km with <50°C of cooling. In this paper we do not examine how these transport systems develop or evolve with time.

1.3. Thermal Efficiency

Cooling in the lava transport system can be described by

$$\rho\, C_p\, (\partial T/\partial t) = Q_{out} - Q_{in} \tag{1}$$

where ρ is density, C_p is heat capacity, T is temperature, t is time, Q_{out} is the heat lost from a control volume of the transport system, and Q_{in} is the heat added to the same control volume. To convert equation (1) to the cooling per unit length of the transport system, one can divide by the velocity, producing

$$\partial T/\partial x = (Q_{out} - Q_{in}) \,/\, \mathbf{v}_x\, \rho\, C_p \tag{2}$$

where x is the flow direction and $\mathbf{v}_x$ is the downflow velocity. If we assume that the properties of the lava (i.e., ρ and C_p) are relatively immutable, then there are only two ways in which to reduce $\partial T/\partial x$: increase $\mathbf{v}_x$ or decrease (Q_{out} - Q_{in}). We will call these end-member cases "rapid" and "insulated" emplacement, respectively.

There is an important caveat to using equations (1) and (2); they only apply to isothermal parts of the lava flow. In the strictest sense, these equations are only valid for infinitesimal control volumes, making them the basis for the partial differential equation governing heat transfer in a moving fluid [e.g., *Bird et al.*, 1960]. However, we agree with *Danes* [1972] that in most lava transport systems it is appropriate, at the first order, to treat the fluid lava as if it were isothermal. A simple Prantl number analysis shows that thermal boundary layers require more that 10 km to become fully established inside typical flows [*Kauahikaua et al.*, this issue]. However, most lava transport systems can be expected to encounter mixing events much more often than once every 10 km. Mixing can take place during turbulent flow or laminar flow past sudden changes in conduit geometry such as lava falls, sharp bends or constrictions, submerged obstacles, or the confluence of two streams. These kinds of features seem to be sufficiently common in channels and tubes to justify the assumption that the fluid lava is well-mixed and thus essentially isothermal within the transport system. Mixing is less likely in slow-moving roofed sheet flows and therefore we will only use this assumption when we can demonstrate that the heat flux out of the sheet is almost exactly matched by the heat generated inside it. As the above discussion shows, we must understand the dynamics of lava flows before we can calculate their thermal efficiency.

2. Lava Flow Dynamics

2.1. Navier-Stokes

Fluid motions are described by the Navier-Stokes equation which can be written as

$$\partial/\partial t\, \rho \mathbf{v} = -[\nabla \bullet \rho \mathbf{v}\mathbf{v}] - [\nabla \bullet \tau] - \nabla p + \rho g \qquad (3)$$

where t is time, ρ is density $\mathbf{v}$ is the velocity vector, p is pressure, τ is stress, and g is gravitational acceleration [e.g., *Bird et al.* 1960]. Equation (3) applies to any moving fluid, and tractable solutions are found by making assumptions about the fluid rheology, flow regime, boundary conditions, etc. We here go through each term of the Navier-Stokes equation, describing and justifying the simplifying assumptions we use in our subsequent modeling.

The left-hand side of equation (3) contains the time dependence of the flow. When working with a well-established lava transport system, it is reasonable to assume that the system is in steady state and set this term equal to zero. This ignores the fascinating (and complicated) details of how the lava transport system evolves with time and how the lava flow front advances.

It is important to note the distinction between flow front velocity and the velocity of the fluid lava. The velocity within a channel or tube can be as much as 3 orders of magnitude higher than the speed at which the flow front advances [cf., *Finch and Macdonald*, 1953; *Booth and Self*, 1973; *Lipman and Banks*, 1987; *Rowland and Walker*, 1990; *Mattox et al.*, 1993; *Hon et al.*, 1994; *Kauahikaua et al.*, this issue; etc.]. This is possible because the cross-sectional area of the entire lava flow front is much greater than the cross-sectional area of the fluid in the transport system. Unless otherwise stated, all velocities we refer to in this paper are velocities within the transport system, not rates of advance of the flow front.

The first term on the right-hand side of equation (3) describes the advection of momentum. For simple geometries with straight, laminar flow lines, this term is negligible. However, this term becomes very important in more complex geometries such as a bend in the stream, flow around obstacles, the confluence of two lava streams, or the sudden change in depth or width of a channel. The classic solutions for fluid flow in rectangular channels or circular pipes ignore this term and often have very limited applicability in nature [e.g., *Goncharov*, 1964]. It is also the dominant term in turbulent flow.

The second right-hand side term describes the viscous stresses retarding the motion of the flow. In order to determine the stresses within the flow, it is first necessary to select boundary conditions. It is customary to assume that there is a "no-slip" condition at the boundary between the fluid lava and the solid walls and a "free-slip" condition at lava-atmosphere boundaries. We make these assumptions throughout this paper but note that thermal and/or mechanical erosion can make defining the "solid" boundary difficult. As an aside, we note that in very unusual circumstances there can be significant drag along the lava-atmosphere interface (e.g., dust-devils that picked lava out of an open channel on Kilauea during the night of February 18-19, 1992).

The third term on the right-hand side of equation (3) incorporates the change in pressure along the flow. Since the change in atmospheric pressure along the length of a lava flow is negligible, this term is only important if there is significant overpressurization. However, overpressurization of a fluid is only possible in an enclosed conduit and cannot happen in open channels and partly drained lava tubes. We suggest that significant overpressurization does not occur even in inflating sheet flows because (1) lava flows gently out of rupturing tumuli, (2) pahoehoe lobes emanating from the base of even large tumuli are not observed to climb above their breakout points, (3) lava does not squirt out of holes drilled into slowly inflating active flows (C. Thornber, U.S. Geological Survey, personal communication, 1998), and (4) volcanic gasses appear to readily escape from inflating flows. However, our assessment that flow overpressurization is negligible is not shared by *Hon et al.* [1994], *Cashman and Kauahikaua* [1997], or *Sakimoto et al.* [1997a,b].

The last term in equation (3) includes the gravitational and buoyancy forces. Since we have argued that flows are not pressure driven, gravity provides the force driving the flow forward. As noted earlier, the lower gravity on the other planetary bodies has the same effect as shallower slope on the Earth. Only when lava flows through a medium of similar density (water or mud) are there significant buoyancy forces.

2.2. Rheology

To translate the stresses in the second term on the right-hand side of equation (3) into velocities, it is necessary to make assumptions about the rheology of the fluid. Throughout this paper, we assume a Newtonian rheology for the fluid lava. This is probably adequate, though not completely correct, for lava within the transport system. As we noted earlier, for long lava flows, we assume that lava cools no more than 50°C in the transport system and most of the known non-Newtonian behavior of lava is found after greater cooling and associated crystallization [e.g., *Shaw*, 1969; *Marsh*, 1981; *Chester et al.*, 1985; *Pinkerton and Stevenson*, 1992].

However, the rheological effect of the bubbles in lava is still not properly quantified. Under low strain rates, when the bubbles remain undeformed, they behave like rigid spheres. Such spheres increase the viscosity and introduce a yield strength [e.g., *Pinkerton and Stevenson*, 1992]. This is demonstrated by slow moving pahoehoe lobes on Kilauea where we have observed that typical lobes with ~50 vol% bubbles are 15-30 cm thick while dense lobes with 5-20 vol% bubbles can form lobes as thin as 2 cm. However, at higher strain rates, the bubble walls tear and the effective viscosity of the fluid drops. This can be directly experienced by dipping a hammer into a foamy flow and then into a dense lobe. After the hammer breaks through the outer crust, the flows with >50% bubbles offer minimal resistance to stirring motion and the hiss of gas escaping from burst bubbles is often clearly audible. Flows with ≤20% bubbles, on the other hand, require significant effort to stir. The differences in viscosity cannot be attributed to differences in temperature, which were only ~10°C as determined by in situ thermocouple measurements [*Keszthelyi*, 1996]. Since there is currently no accepted method to calculate the effect of bubbles on lava rheology, we do not include this phenomenon in our models. However, we do correct for the effect of bubbles on the density and other thermal properties of lava. We assume 20 vol% bubbles in our subsequent modeling, a number that is in accord with observations from larger basaltic lava flows [e.g., *Keszthelyi and Pieri*, 1993; *Thordarson*, 1995].

Ignoring the effect of bubbles, lava viscosity can be estimated given the melt composition, lava temperature, and crystal content [e.g., *Shaw*, 1972; *Pinkerton and Stevenson*, 1992]. For typical basaltic compositions (49-51 wt % SiO_2), lava temperatures (1050-1200°C), and crystal contents (0-25%), calculated lava viscosities are usually in the range of 100-1000 Pa s [e.g., *Keszthelyi and Pieri*, 1993; *Crisp et al.*, 1994; *Thordarson*, 1995; *Ho and Cashman*, 1997]. Viscosities in this range have been measured, directly and indirectly, in the field from basaltic lavas [e.g., *Shaw et al.*, 1968; *Pinkerton and Sparks*, 1978; *Heslop et al.*, 1989; *Pinkerton et al.*, 1996].

However, it is important to note that many estimates of the bulk viscosity of lava flows are much higher than the values we use in our models. This is particularly true for aa flows: estimated bulk viscosities for both the 1984 Mauna Loa and 1983 Kilauea flows range from 10^5 to 10^7 Pa s [*Moore*, 1987; *Fink and Zimbelman*, 1986]. The bulk properties of aa flows also include a significant yield strength [e.g., *Hulme* 1974; *Chester et al.* 1985; *Dragoni*, 1993; *Pinkerton and Wilson*, 1994]. Both the enhanced bulk viscosity and the apparent yield strength are probably caused by the disrupted crust. In fact, *Booth and Self* [1973] found that the apparent viscosity of the 1971 Etna flow was less than 1000 Pa s near the vent where essentially no crust had yet formed but rapidly increased to 10^7 Pa s down flow with the growth of a clinkery aa crust. Similarly, *Fink and Zimbelman* [1986] found a 30-fold increase in yield strength down the length of the 1983 Kilauea aa flows which they attributed to "increased brecciation and thickening" of the rubbly aa crust. However, for flow within the lava transport system we contend that it is the rheology of the fluid lava, not the bulk rheology of the entire flow, that is of interest and that the lower viscosity estimates are appropriate.

2.3. Laminar Versus Turbulent Flow

Fluid flow should be treated in different ways depending on whether the flow is in the laminar or turbulent regime. For laminar flow in simple geometries the average flow velocity ($\langle v \rangle$) can be calculated using Jeffrey's equation. For open channel flow of a Newtonian fluid on a shallow slope, this can be expressed as

$$\langle v \rangle = \rho\, g\, \theta\, H^2 / 3\, \eta \tag{4}$$

where θ is slope, H is flow thickness, and η is viscosity. For a sheet flow between two parallel plates, the 3 in the denominator is replaced by an 8 [e.g., *Bird et al.*, 1960].

Turbulent flow is more difficult to define, much less quantify. There are a multitude of definitions of turbulent flow, but they all hinge on the concept of chaotic flow in which small perturbations grow down flow. In order for this to occur, the inertial forces (first term on the right-hand side of equation 3) must dominate over the viscous forces (second right-hand side term in equation 3). The ratio of these two terms is expressed by the non-dimensional Reynolds number (*Re*) [e.g., *Reynolds*, 1974].

Definitions of *Re* vary from reference to reference and according to the geometry of the flow. Perhaps the most general definitions for *Re* is

$$Re = \rho\, r_h \langle v \rangle / \eta \tag{5}$$

where ρ is density, r_h is an appropriate length scale, $\langle v \rangle$ is the average velocity, and η is the viscosity [e.g., *Whitaker*, 1981]. The parameter r_h is the length scale of the largest perturbation that is able to propagate down flow (smaller perturbations are more easily damped out by viscous forces). As such, the use of the length of the entire flow [e.g., *Ho and Cashman*, 1997] is inappropriate. Instead, the greater of the width or thickness of the flow would seem to be the most appropriate dimension. However, for wide sheet flows it is the well-established procedure to use the thickness of the flow instead of the width [e.g., *Bird et al.*, 1960; *Reynolds*, 1974; *Whitaker*, 1981]. Also, in the case of obstacles in the flow, the greater of the dimension of the obstacles or the dimension of the spacing between the obstacles should be used.

Flow usually becomes turbulent for Re above 2000 for a pipe ($r_h = D$) or 500 for broad sheet-like floods ($r_h = H$) and for narrow channels whose width and depth are equal ($r_h = H/3 = W/3$) [*Bird et al.*, 1960]. However, in the laboratory it is possible to maintain laminar flow in pipes at Re >50,000 [*Reynolds*, 1974] and much caution needs to be used in applying these criteria to lava flows. The most serious caveat in using equation (5) to determine if a lava flow is turbulent or laminar is that flow dimension, flow velocity, lava viscosity, and lava density can all vary greatly in both time and space in a lava flow. Selecting values of these parameters that do not reflect the bulk properties of the fluid lava may lead to very misleading results.

Given the difficulties in computing *Re* for real lava flows, it is important to examine field observations of active lava flows. *Kauahikaua et al.* [this issue] contains a more complete description of non-laminar flow within tubes, but we add some additional general observations here. It is clear that the flow of small pahoehoe lobes, though often unpredictable, is in the laminar regime. This is true for lobes moving up to a few centimeters per second with dimensions of 0.5-1 m. Assuming a viscosity around 100-1000 Pa s, and a bubble corrected density of about 1300 kg/m^3, these lobes should have $Re < 10$. However a network of 3-5 m wide, ~1-m-deep channels active on August 20, 1994 on Kilauea with flow velocities of 2-4 m/s demonstrated early signs of turbulence such as shedding of vortices. These flows still have $Re < 300$ but suggest that transition to turbulence may indeed take place at a Re of about 500 in open channels. For basaltic lava flows with typical viscosities of 100-1000 Pa s and r_h of 5-20 m, the transition to turbulent flow is predicted to take place at velocities between 2.5 and 100 m/s.

If a flow is found to be turbulent, there are a number of expressions for computing flow velocities. The simplest of these is the Chezy formula given by

$$\langle v \rangle^2 = g\, H\, \theta \,/\, C_f \tag{6}$$

where C_f is the friction coefficient [e.g., *Jeffreys*, 1925]. The difficulty in using this formula lies in selecting a value for C_f. Figure 1 shows a typical plot of C_f versus Re. For highly turbulent flow ($Re \gg 10^5$), C_f appears to be relatively constant and a function of bottom roughness instead of fluid properties. Values for C_f vary between 0.0025 for smooth water channels to ~1 for lahars and debris flows [e.g., *Jeffreys*, 1925; *Weir*, 1982]. *Baloga et al.* [1995] estimated values for C_f between 0.05 and 0.11 for the 1801 Hualalai flows and 0.0057-0.0128 for the 1823 Kilauea eruption.

The dependence of C_f on bottom roughness can be empirically quantified. For example, the most popular formulation of turbulent flow in civil engineering applications uses the Manning friction factor which is defined as

$$f = 116\, n^2 \,/\, r_h^{1/3} \tag{7a}$$

$$C_f = f/8 \tag{7b}$$

with n being the "roughness factor," which varies from 0.007 to 0.03 $m^{1/6}$ for various natural surfaces [e.g., *Marks and Baumeister*, 1958; *Whitaker*, 1981]. *Booth and Self* [1973] estimated n^2 to be between 0.07 and 2.20 $m^{1/3}$ for a lava channel in the 1971 Etna eruption.

However, both the Chezy and Manning formulations are strictly valid only for very high Re where C_f becomes insensitive to changes in Re (Figure 1). The most appropriate method of calculating C_f for a moderately turbulent flow (Re ~10^3-10^4) is probably the one used by *Shaw and Swanson* [1970] when they examined the emplacement of the Columbia River flood basalts. They used the work of *Goncharov* [1964] for turbulent sheet flow over smooth surfaces which allows the average flow velocity to be iteratively calculated from

$$C_f = \lambda/2 \tag{8a}$$

$$\lambda = \{4 \log_{10}[6.15((Re'+800)/41)^{0.92}]\}^{-2} \tag{8b}$$

$$Re' = 2Re \tag{8c}$$

In this case C_f (and flow velocity) is dependent on both the dimensions of the channel and the properties of the fluid. This equation is applicable where bottom roughness is small compared to the flow thickness and $Re < 10^5$. We prefer the Goncharov formulation over the Chezy and Manning formulations because (1) we expect lava flows to indeed have $Re < 10^5$, (2) the Chezy and Manning formulas depend on empirical factors that can only be determined after obtaining independent estimates of flow velocity, and (3) the Goncharov formulation explicitly includes the viscosity of the lava via the Re.

Regardless of the exact formulation of the turbulent flow, the difference between laminar and turbulent flow is extremely important for thick, fast moving lavas. For laminar flow, velocity increases as the flow thickness squared. For turbulent flow, velocity increases only as the square root of the flow thickness. As pointed out by *Baloga et al.* [1995], the failure to recognize that the flow has undergone a transition to the turbulent flow regime has led to very large over-estimates of flow velocity [cf. *Keszthelyi and Pieri*, 1993; *Miyamoto and Sasaki*, 1996]. Having discussed our basic assumptions of flow dynamics, we can proceed to model the thermal efficiencies of the rapid and insulated modes of emplacement.

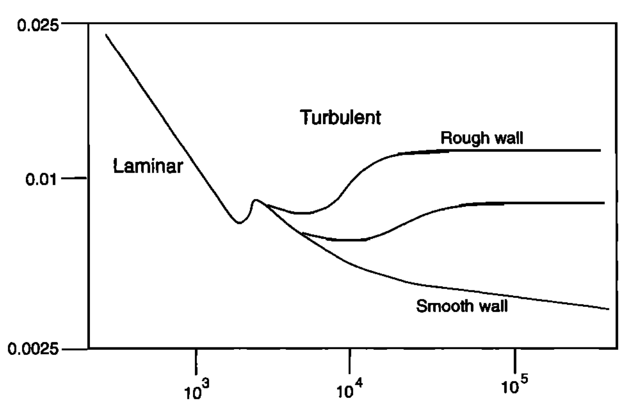


Figure 1. Schematic sketch of friction factor as a function of Reynolds number. After *Reynolds* [1974] and *Goncharov* [1964]. Note that the curve is linear for laminar flow and at very high Re. The Chezy and Manning formulas are designed for the high Re case, but the Goncharov formulation is more appropriate at intermediate Re. Locations of the laminar to turbulent transition is for pipe flow and would be shifted to lower Re for open channel flow.

3. Thermal Models

3.1. Rapid Emplacement

Fast moving lava flows have invariably had open channels connecting the vents to the flow fronts. This assertion is supported by a wealth of empirical observations [e.g., *Einarsson*, 1949; *Finch and Macdonald*, 1953; *Lipman and Banks*, 1987; *Rowland and Walker*, 1990; *Pinkerton and Wilson*, 1994; *Barnard*, 1995] and laboratory experiments [e.g., *Hallworth et al.*, 1987; *Griffiths and Fink*, 1992; *Gregg and Fink*, 1996].

The dynamics and thermal budget of open channels are investigated and described in detail by *Crisp and Baloga* [1994]. We adopt their model with only minor modifications. *Crisp and Baloga* [1994] assume that an open channel flow has a well-mixed, isothermal core that is only partially covered by a crust. This crust is approximately isothermal and is continually created and destroyed by entrainment into the core. *Crisp and Baloga* [1994] quantified the thermal budget for such a flow with

$$\frac{\partial T}{\partial t} = \frac{L}{C_p}\frac{\partial \Phi}{\partial t} - \frac{\varepsilon \sigma f T^4}{\rho C_p H} + \frac{(T_e - T)}{\tau_e} + \frac{L}{C_p}\frac{(\Phi - \Phi_e)}{\tau_e} \tag{9}$$

where L is the latent heat of crystallization, C_p is heat capacity, Φ is the vol% crystals in the fluid lava, ε is the emissivity, σ is the Stephan-Boltzman constant, f is the fraction of the flow not covered by crust, H is channel depth, T_e is the temperature of the crust, τ_e is the timescale for entrainment of the crust back into the fluid core of the flow, and Φ_e is the crystallinity of the crust that is being remelted.

We make several minor modifications. First, we convert to the cooling rate per unit distance of flow by dividing by the average velocity. Then, we incorporated latent heat into our expression for heat capacity using $C_p^* = C_p + L\,\partial\Phi/\partial T$.

While τ_e was the natural way for *Crisp and Baloga* [1994] to describe the rate of entrainment given the field data collected from the 1984 Mauna Loa eruption, it is difficult to generalize to other flows. Thus we revised the expressions for entrainment, using a timescale for the survival of a piece of crust on the surface of the flow (τ) [*Keszthelyi and McEwen*, 1997]. The two are related via $\tau_e = \tau\,(H/H_c)$ where H is flow thickness and H_c is the average thickness of the crust.

Finally, we add atmospheric convective cooling and heating via viscous dissipation (shear heating). We have not included conductive heat losses from the base of the flow, even though recent work suggests that it may be comparable to the other heat loss mechanisms [*Harris et al.*, 1998; L. Hultgrien, written communication, 1997]. Our modified version of the *Crisp and Baloga* [1994] model can be written as

$$\partial T/\partial x = \{Q_{visc}-Q_{rad}-Q_{atm}-Q_{entr}\}/\{\langle v\rangle\,\rho\,C_p^*\,H\} \tag{10}$$

Where Q_{visc} is the viscous heating, Q_{rad} is radiative heat losses, Q_{atm} is atmospheric heat loss, Q_{entr} is the heat lost from the core of the flow via entrainment of the crust, and the other terms are as defined earlier. If the flow is not accelerating, the viscous heating matches the loss of potential energy and

$$Q_{visc} = \rho\, g\, H \langle v\rangle\, \theta \tag{11}$$

[e.g., *Keszthelyi*, 1995]. Radiative heat losses can be expressed as

$$Q_{rad} = \varepsilon\sigma f\,(T^4 - T_a^4) \tag{12}$$

where ε is emissivity (~0.95), σ is the Stephan-Boltzman factor, f is the fraction of core exposed, T is the core temperature, and T_a is the ambient temperature [*Crisp and Baloga*, 1990]. The atmospheric heat loss has a similar expression

$$Q_{atm} = \mathbf{h}\, f\,(T - T_a) \tag{13}$$

where **h** is the atmospheric heat transfer coefficient, which has a nominal value of 70 W m^{-2} K^{-1} [*Keszthelyi and Denlinger*, 1996]. Substituting our expressions for τ and C_p^* simplifies the entrainment term to

$$Q_{entr} = \rho\, C_p^*\, H_c\,(T - T_c)\,/\,\tau. \tag{14}$$

Before determining the conditions required for an open channel flow to cool less than 0.5°C/km with this model, it is useful to first investigate the sensitivity of the model to the different input parameters. Figure 2 plots the cooling rate as a function of each of the inputs. The nominal case has $\langle v\rangle$ = 10 m/s, H = 10 m, θ = 1%, f = 10%, T_c = 700°C, T = 1100°C, T_a = 30°C, τ = 300 s, H_c = 2 cm, ε = 0.95, ρ = 2080 kg/m^3, and C_p^* = 1575 J/kg °C. We did not investigate the effects of varying the intrinsic properties of the lava (ρ, C_p^*, ε, and T) and varying T_a has negligible effect and is not plotted. Figure 2 shows that none of the remaining parameters can be considered negligible, although slope, flow thickness, and τ have the most dramatic effects.

Figure 3 plots flow velocity as a function of flow thickness, lava viscosity, and slope using equations (4) and (8). The sharp kink in the curves marks the transition from laminar to turbulent flow. While the effect of viscosity is significant, it is completely overshadowed by the effect of slope. On a 10% slope, the Goncharov formulation predicts that the channel velocity will exceed 100 m/s for a 30 m thick flow. The same channel on a 0.1% gradient would be moving at less than 10 m/s. This has notable effects on the thermal efficiency of the flow.

In principle, the properties of the upper crust (τ, f, H_c, T_c) might be expressed as functions of flow velocity or Reynolds number. However, the make-up of the upper crust can be expected to vary significantly along the length of the flow and with time. Entrainment (and cooling) are likely to be concentrated at locations where the geometry of the channel changes suddenly (lava falls, sharp bends, etc.). Instead of creating unconstrained relationships between the properties of the upper crust and the flow regime, we investigate three cases: thick, thin, and no upper crust. Typical slow (≤ 1 m/s) aa flows would roughly fall in the "thick crust" case (f = 1%, τ = 1 day, H_c = 1 m, T_c = 400°C) [cf. *Crisp and Baloga,* 1990, 1994]. We expect faster flows (≥ 5 m/s) to be more akin to the "thin crust" case (τ = 300 s, f = 10%, H_c = 2 cm, T_c = 700°C). These values are in general accord with the qualitative observations of the most rapid flows on Mauna Loa [e.g., *Finch and Macdonald*, 1953; *Lipman and Banks*, 1987]. Unfortunately, there are few field observations from channels with velocities >10 m/s. However, as an extreme case, we calculated the cooling rate for no crust (f = 100%, H_c = 0).

Figure 4 plots the cooling rate for these three cases for a slope of 1% (0.57°) and Table 2 lists the conditions necessary to achieve cooling rates <0.5°C/km for a range of crusts, viscosities, and slopes. For open channel flows with a thick crust, flows >100 km long require minimum flow velocities of 1.6 - 5.6 m/s, minimum flow thicknesses of 2 - 11 m, and effusion rates greater than 200 - 1900 m^3/s. These values are within the range of large historical eruptions. The 1984 Mauna Loa eruption had a maximum sustained effusion rates of 140 m^3/s and the historical eruption with the highest effusion rates, the 1783-1784 Laki eruption, was able to sustain 2500 m^3/s for about 4 months [*Lipman and Banks*, 1987; *Thordarson and Self*, 1993]. However, we doubt that a thick crust could be maintained on a flow moving at ~5 m/s.

For the thin crust and no crust cases, long flows require velocities greater than 3-15 m/s, flows thicker than 3.7-19 m, and effusion rates in excess of 1100 - 1.7x10^4 m^3/s. This suggests that eruptions similar to Laki might produce long lava flows, but only if they are emplaced on slopes ≥10%. The long lava flows in Table 1 were all emplaced on slopes ≤1% and would have required effusion rates at least several times larger than any observed eruption. Increased viscous dissipation is the primary reason why flows emplaced in the rapid mode are more thermally efficient on higher slopes.

The idea that very high effusion rates are necessary for long lava flows has been reached by many previous workers using several different lines of reasoning [e.g., *Shaw and Swanson*, 1970; *Walker*, 1973; *Kilburn and Lopes*, 1991; *Komatsu et al.*, 1992; *Pinkerton and Wilson,* 1994].

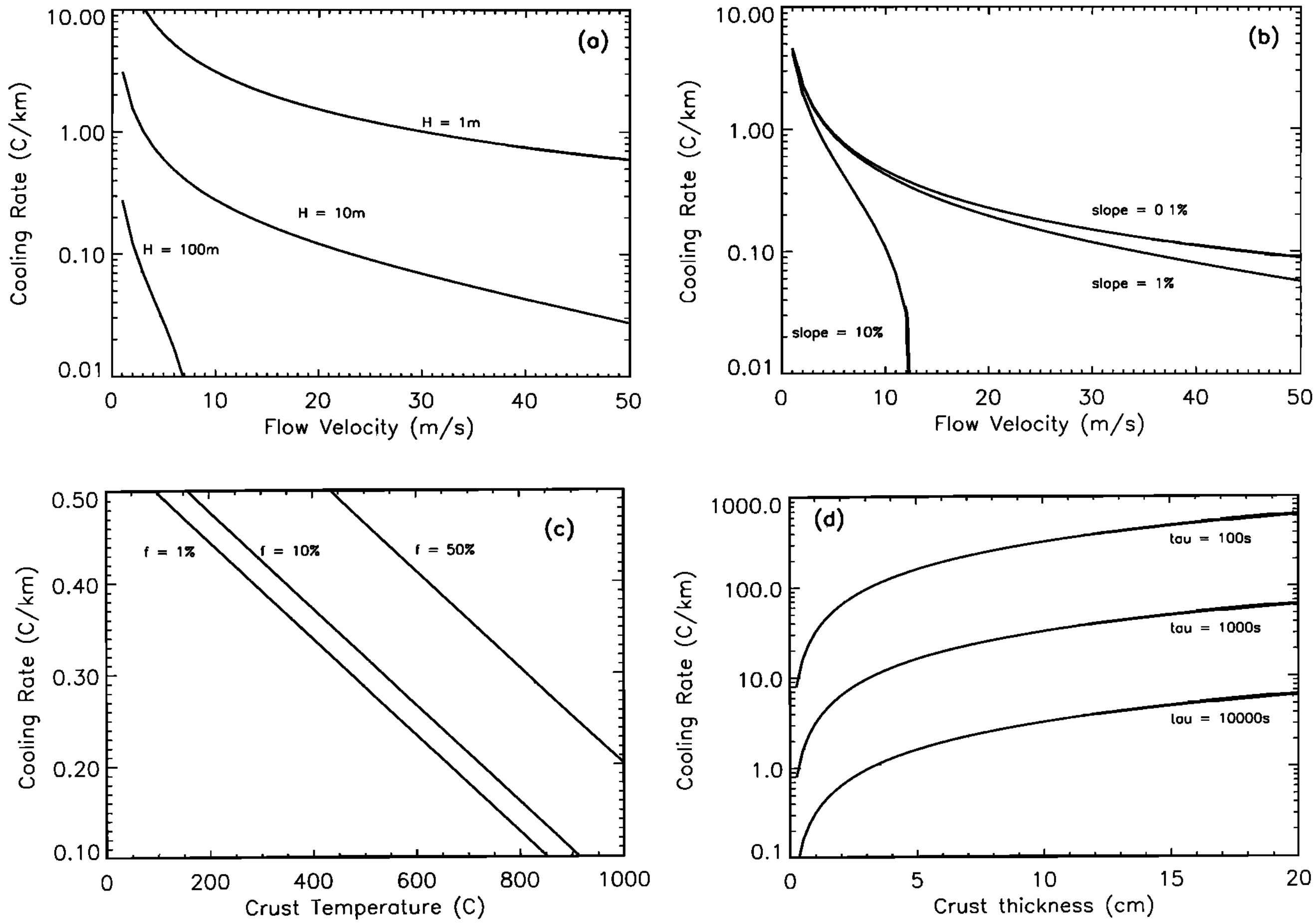


Figure 2. Cooling rate of channelized flows. Long lava flows are possible for cooling rates below ~0.5°C/km. Baseline input parameters given in text. (a) Cooling rate as a function of flow velocity and channel depth (H). Note that flows should be at least several meters thick to have cooling rates below 0.5°C/km. (b) Cooling rate as a function of flow velocity and slope. The difference between the curves is largely due to changes in the heat input by viscous dissipation. (c) Cooling rate as a function of crust temperature and fraction of core exposed (f). (d) Cooling rate as a function of crust thickness and crust survival time scale (τ). Note that cooling rate is very sensitive to τ.

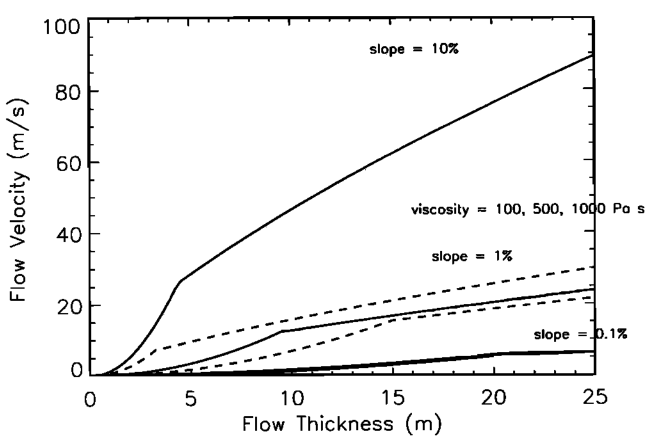


Figure 3. Channel velocities as a function of flow thickness, viscosity, and slope. Dashed lines are flow velocities for viscosities of 100 and 1000 Pa s and 1% slope. Solid lines are for 0.1, 1, and 10% slopes and 500 Pa s viscosity. The kinks in the curves are at the transition from laminar to turbulent flow.

3.2. Insulating Emplacement

To significantly reduce heat loss, the flow must develop an undisrupted crust. Such a stationary crust will effectively insulate the core of the flow from direct radiative and convective cooling as well as eliminate the entrainment of cold crust. The dynamics of such a flow are different enough from open channel aa flows to preclude the use of models closely related to that of *Crisp and Baloga* [1994]. The thermal models for insulated emplacement must also take into account differences in the geometry of the lava transport system (i.e., generally cylindrical tubes vs. sheets that are much wider than they are thick).

3.2.1. Lava tubes. Lava tubes have been subjected to increasingly quantitative observations over the years [*Kauahikaua et al.*, this issue; *Greeley et al.*, this issue; and references therein]. The theoretical modeling of tubes has lagged behind but two sets of thermal models for lava tubes have been recently published [*Keszthelyi* 1992, 1995; *Sakimoto and Baloga*, 1995; *Sakimoto et al.*, 1997a,b; and *Sakimoto and Zuber*, this issue]. Despite different initial assumptions, both sets of studies reach essentially identical

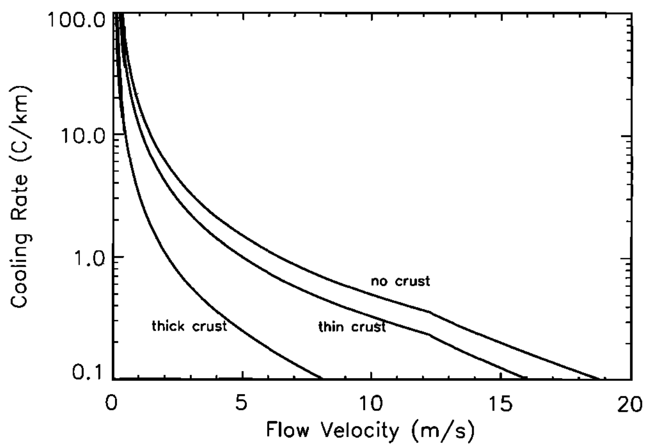


Figure 4. Cooling rate as a function of flow velocity for the thick crust, thin crust, and no crust cases. The kink in the curve is at the transition from laminar to turbulent flow. Parameters for the three cases are listed in the text, curves plotted for 1% slope and viscosity of 500 Pa s.

conclusions: lava tubes in excess of 100 km in length (and cooling rates <0.5°C/km) are readily achieved with tube diameters of 10-50 m and effusion rates of a few tens of m^3/s. This conclusion seems to hold for tubes on other planets, in the submarine environment, and for a range of slopes and lava rheologies. Both sets of studies also agree that the key to producing a long tube-fed lava flow is a steady, uninterrupted eruption. In the absence of significant downward thermal and/or mechanical erosion, lava tubes with diameters ≥ 10 m require flows at least as thick. Since erosion is relatively uncommon for lava flows emplaced on shallow slopes [*Greeley et al.*, this issue; *Kauahikaua et al.*, this issue], long, tube-fed lava flows should be greater than 10 m thick.

The fundamental difference between the two sets of models is that *Sakimoto and Baloga* [1995], *Sakimoto et al.*, [1997a,b], and *Sakimoto and Zuber*, [this issue] assume that the rate limiting step for cooling in a lava tube is the ability of heat to be transported through the fluid lava to the tube wall. *Keszthelyi* [1992, 1995] assume that it is the ability of heat to escape through the walls and roof of the tube that limits heat loss. The assumption that it is difficult for heat to be transported within the liquid lava is reasonable for completely laminar flow within a filled tube. However, field observations suggest that significant mixing takes place within a lava tube, even when the flow has low to moderate Reynolds numbers [*Kauahikaua et al.* this issue]. Also, long-lived tubes are rarely filled [*Kauahikaua et al.*, this issue, *Greeley et al.*, this issue]. Since the typical heat flux escaping the fluid lava can be sustained by a ~10°C temperature difference between the liquid lava and its roof [*Keszthelyi*, 1995], radiative heat transfer can easily transport heat from the liquid lava across the air gap to the roof and walls of the tube. The different assumptions do not alter the final conclusions because both models agree that lava tubes are highly effective insulators.

3.2.2. Sheet flows. Sheet flows are the one case in which it is likely that the fluid lava will not be well-mixed. As such, a full solution of the coupled differential equations governing conductive heat loss and fluid flow, such as that of *Sakimoto and Zuber* [this issue] may be both valid and useful. As an alternative, we use a simple comparison of the heat lost via conduction to the heat added by viscous dissipation in order to constrain the conditions under which sheet flows might feed lava 100 km or more. We can assume that the core of the flow is isothermal as long as viscous heating and conductive cooling are nearly matched because both heating and cooling are concentrated at the roof and floor of the flow, leaving the core temperature largely unaffected.

The simple thermal balance for a sheet flow is the same as equation (10) except Q_{rad}, Q_{atm}, and Q_{entr} are assumed to be zero and we add steady state conductive heat loss (Q_{cond}):

$$Q_{cond} = k\,(T - T_a)\,/\,H_c \tag{15}$$

where k is thermal conductivity, T is the lava temperature, T_a is the ambient temperature, and H_c is the thickness of the upper crust [e.g., *Bird et al.*, 1960]. Thermal conductivity is a complex function of temperature and vesicularity, but we use a

Table 2. Constraints on Long Channel-fed Lava Flows

Viscosity, Pa s	Slope, %	Minimum Flow Thickness, m	Minimum Average Velocity, m/s	Minimum Flux, m^2/s	Minimum Effusion Rate[a], m^3/s
			Thick Crust		
100	1	2.9	5.6	16	470
500	10	2.0	5.0	10	200
500	1	4.8	3.1	15	710
500	0.1	11	1.6	18	1900
1000	1	6.2	2.5	16	960
			Thin Crust		
100	1	5.6	10	56	3100
500	10	3.0	12	36	1100
500	1	7.5	7.7	58	4300
500	0.1	17	3.7	63	1.1×10^4
1000	1	9.1	5.6	51	4600
			No Crust		
100	1	7.0	12	84	5900
500	10	3.4	15	51	1700
500	1	8.6	10	86	7400
500	0.1	19	4.8	91	1.7×10^4
1000	1	11	8.0	88	9700

[a] Assumes channel width equals 10 times channel depth.

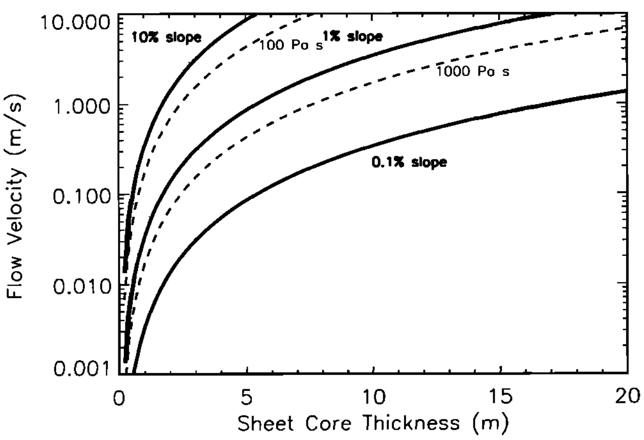


Figure 5. Sheet flow velocities as a function of core thickness, viscosity, and slope. Dashed lines are flow velocities for viscosities of 100 and 1000 Pa s and 1% slope. Solid lines are for 0.1, 1, and 10% slopes and 500 Pa s viscosity. Velocities above 10 m/s were not considered since a stationary roof probably would not form at such velocities.

value of 1 W/m °C in the following rough calculations. We also assume that the difference between the lava and the ambient temperature is about 1000°C.

This reduces $\partial T/\partial x$ to a function of flow velocity and crust thickness. We calculate the flow velocity as a function of fluid core thickness, slope and lava viscosity using equation (4) with the appropriate denominator. Figures 5 and 6 plot velocity and cooling rate as functions of these inputs. Figure 6 is strictly valid only when the conductive heat loss is exactly matched by the viscous heating but remains accurate as long as $\partial T/\partial x$ is small. We do not plot sheet flow velocities >10 m/s because it seems unlikely that an undisrupted, stationary crust can form at such velocities. Figure 6 shows that sheet flows can be incredibly thermally efficient for a wide range of viscosities and crust thicknesses. Table 3 lists the conditions under which there is no net cooling of the lava within the sheet flow, and hypothetically, the flow could reach infinite length. Note that with an insulating crust 1-10 m thick, flow velocities of only 20 cm/s to 1.4 m/s are needed to completely overcome the conductive heat loss. While this is relatively slow, even at 20 cm/s lava will travel 100 km in less than 6 days. The volumetric flux through these sheets depends critically on the width of the flow. If we assume that the sheet is at least 10 times wider than its total thickness, the minimum effusion rates are ~8-11 m^3/s on a 10% slope, ~50-300 m^3/s at 1%, and ~920-7100 m^3/s at 0.1%. It is interesting that all these values are within the range of historical eruptions except for the sheet on a 0.1% slope with only a 1-m-thick crust.

4. Expected Flow Morphology

Given the fundamental difference in the behavior of lava flows emplaced in the rapid and insulated modes, we expect significant differences in the final morphology of the resulting flow. Thus it should be possible to determine the mode of emplacement of ancient long lava flows by careful examination of their morphology. In this section we describe key diagnostic morphologic features as well as some of the caveats in their use.

4.1. Aa Versus Pahoehoe

The dynamics we have assumed for the rapid mode of emplacement (i.e., channelized flow with a well-mixed interior and a disrupted, mixed crust) are the dynamics of a large aa flow [e.g., *Kilburn*, 1990]. Similarly, the insulated case with tube

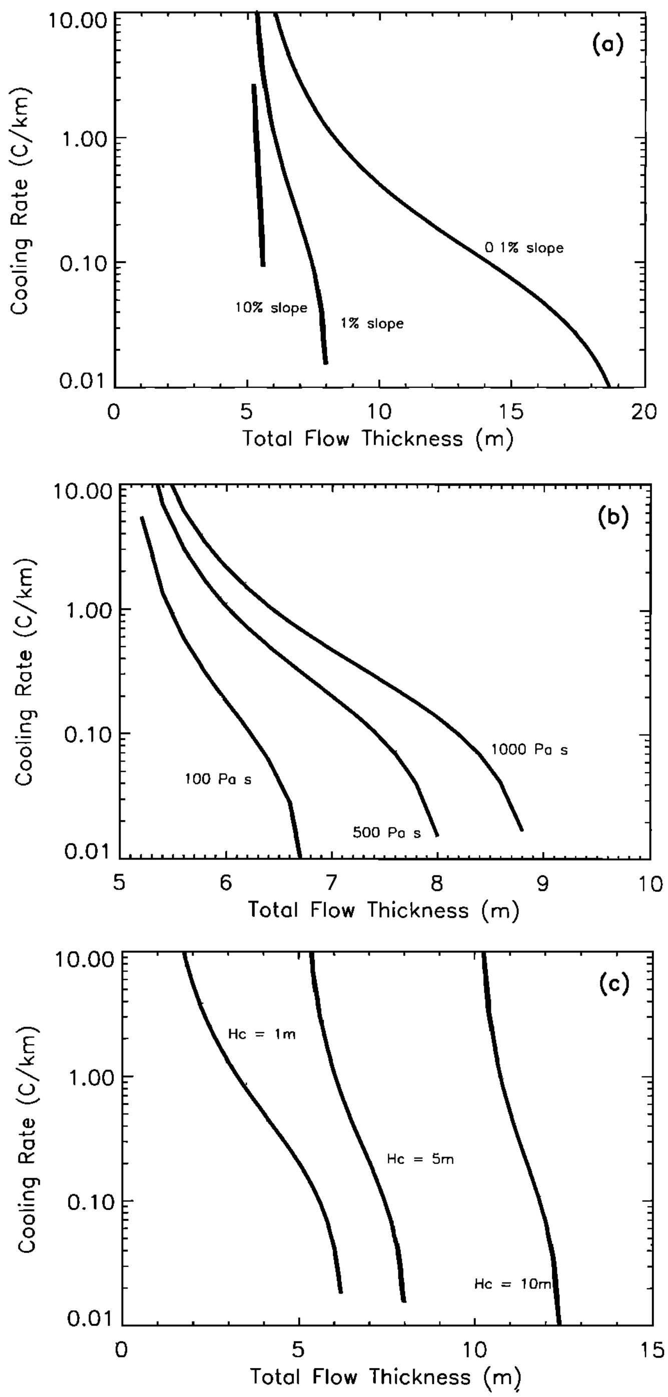


Figure 6. Cooling rate of sheet flows. (a) Cooling rate as a function of total flow thickness (core plus crust) and slope for 5-m-thick upper crust and 500 Pa s viscosity; 10% slope curve terminates because flow velocities exceeded 10 m/s at total flow thicknesses >6 m. (b) Cooling rate as a function of total flow thickness and lava viscosity. Slope is 1%, and upper crust is 5 m thick. (c) Cooling rate as a function of total flow thickness and upper crust thickness. Slope is 1%, and lava viscosity is 500 Pa s.

Table 3. Conditions for Zero Cooling Inside a Sheet Flow

Slope, %	Upper Crust Thickness, m	Total Thickness, m	Average Velocity, m/s	Minimum Flux, m^2/s	Minimum Effusion Rate,[a] m^3/s
10	1	2	0.4	0.4	8
10	5	6	0.2	0.2	12
10	10	11	0.1	0.1	11
1	1	6	1.0	5	300
1	5	8	0.3	0.9	72
1	10	12	0.2	0.4	48
0.1	1	23	1.4	31	7100
0.1	5	20	0.7	11	2200
0.1	10	21	0.4	4.4	920

[a] Assumes sheet width equals 10 times the total flow thickness. Changing viscosity between 100 and 1000 Pa s leads to less than 50% changes in velocities and effusion rates.

or sheet flow below a stationary, undisrupted crust typifies inflated pahoehoe flows. Thus the simple observation that a long lava flow is pahoehoe or aa should be a strong indicator of the mode of emplacement.

There are many ways in which to identify an aa flow. One of the diagnostic features the formation of rubbly zones at both the flow top and flow bottom [*Macdonald*, 1953]. The core of the flow is typically quite dense and the contact between the core and the rubbly flow top and bottom is very irregular, with fingers of dense core reaching far into the rubbly zones and chunks of entrained rubble preserved within the core (Figure 7). Fresh aa flows also have distinctive radar backscattering properties and fractal dimensions of their flow margins when compared to pahoehoe flows [*Campbell and Campbell*, 1992; *Bruno et al.*, 1992; *Bruno*, 1994; *Bruno and Taylor*, 1995]. Aa flows also have distinctive elongate, angular vesicles, and are more finely crystalline [*Macdonald*, 1953]. Aa and pahoehoe can also be distinguished in thin section: aa flows typically have 10-25 vol% 10-50 μm microphenocrysts embedded in a matrix of <5 μm groundmass grains while pahoehoe flows typically have an open matrix of 40-100 μm microphenocrysts with little interstitial groundmass (<10 vol%) [*Friedman*, 1998].

Inflation features, which are largely absent on aa flows, can help identify long pahoehoe flows. A long pahoehoe flow must be extensively inflated because typical advancing pahoehoe lobes are only 15-50 cm thick [*Walker*, 1996; *Keszthelyi and Denlinger*, 1996] but our modeling indicates that long sheet flows need to be ≥5 m thick. In Hawaii, inflation has been observed to thicken flows that were initially about 30 cm thick to a thickness of 3-7 m over a period of weeks [*Hon et al.*, 1994]. The same process has been proposed to produce the 20-50 m thick lava flows in the Columbia River Basalts [*Self et al.*, 1996a,b, 1997], other flood basalt provinces [*Keszthelyi et al.*, 1998; *Self et al.*, 1998], and other long lava flows [e.g., *Keszthelyi and Pieri*, 1993; *Hill and Perring*, 1996; *Stephenson et al.*, 1996]. The inflation process produces a plethora of distinctive surface features. These features include flat-topped sheet flows, inflation ridges, tumuli, inflation pits, and raised tree molds [*Theilig and Greeley*, 1986; *Walker*, 1991; *Keszthelyi and Pieri*, 1993; *Hon et al.*, 1994; *Self et al.*, 1996a; *Stephenson and Whitehead*, 1996; *Whitehead and Stephenson*, 1996]. Commonly these inflation features are on the order of 1-100 m in scale, although some inflation features are tens of kilometers in length (e.g., the 40-km-long Mount Surprise Wall of the Undara flow field in Queensland, Australia [*Stephenson*, 1996]).

Inflated pahoehoe flows also have a distinctive three-part internal structure (Figure 8) [*Aubele et al.*, 1988; *Thordarson*, 1995; *Thordarson and Self*, 1996; *Spry and Stephenson*, 1996; *Self et al.*, 1996a,b, 1997, 1998; *Cashman and Kauahikaua*, 1997]. The upper crust is vesicular because of the continuous flux of bubble laden lava running through the sheet. Field observations show that this upper vesicular crust usually makes up ~50% of the total thickness of sheet flows [*Thordarson*, 1995; *Self et al.*, 1997; *Cashman and Kauahikaua*, 1997]. By comparison, the thickness of the vesicular crust on stagnant lava ponds [*Peck*, 1978; *Wright and Okamura*, 1977; *Mangan and Helz*, 1986] is proportionally much less [*Self et al.*, 1998].

While the distinction between pahoehoe and aa should be generally useful in distinguishing the rapid and insulated modes of emplacement, lava flows are notorious for violating any generalization. Tube-fed aa flows [e.g., *Calvari and Pinkerton*, this issue] are an example of flows that undergo a transition from the rapid to the insulated mode of emplacement as their channels crust over to form lava tubes. Also, there are transitional lava types (e.g., slab pahoehoe) that are labeled as "pahoehoe" but in fact demonstrate dynamics more akin to aa flows. Furthermore, it is very important to note that aa flows often have small pahoehoe spill overs or breakouts [e.g., *Jurado-Chichay and Rowland*, 1995; *Kilburn*, 1996] and large inflated pahoehoe flows often have small patches of slab pahoehoe or even aa [e.g., *Keszthelyi and Pieri*, 1993, *Kilburn* 1996]. Thus it is impossible to deduce the dominant style of emplacement by observing a single locality or using a single isolated morphologic criterion.

We also note that *Baloga et al.* [1995] and *Guest et al.* [1995, 1996] argue that the vigorous entrainment associated with rapid emplacement will lead to the formation of an extremely smooth final surface. We disagree. While very rapid emplacement can lead to an insubstantial crust at the top of an active lava channel, the surface morphology of the moving lava is not a good indicator of the final surface of the frozen lava. Before the flow freezes, it must slow and the entrainment rates must decrease and the disrupted crust should be preserved in the final frozen surface, at the very least along the channel levees and the flow front.

We would be more confident in using aa morphology as a criterion for rapid emplacement of a long lava flow if it could be demonstrated that the flow velocities, viscosities, etc., required for the rapid mode of emplacement would also require the formation of an aa flow. Unfortunately quantitative constraints on the pahoehoe to aa transition are still largely lacking. While *Peterson and Tilling* [1980] clearly show that

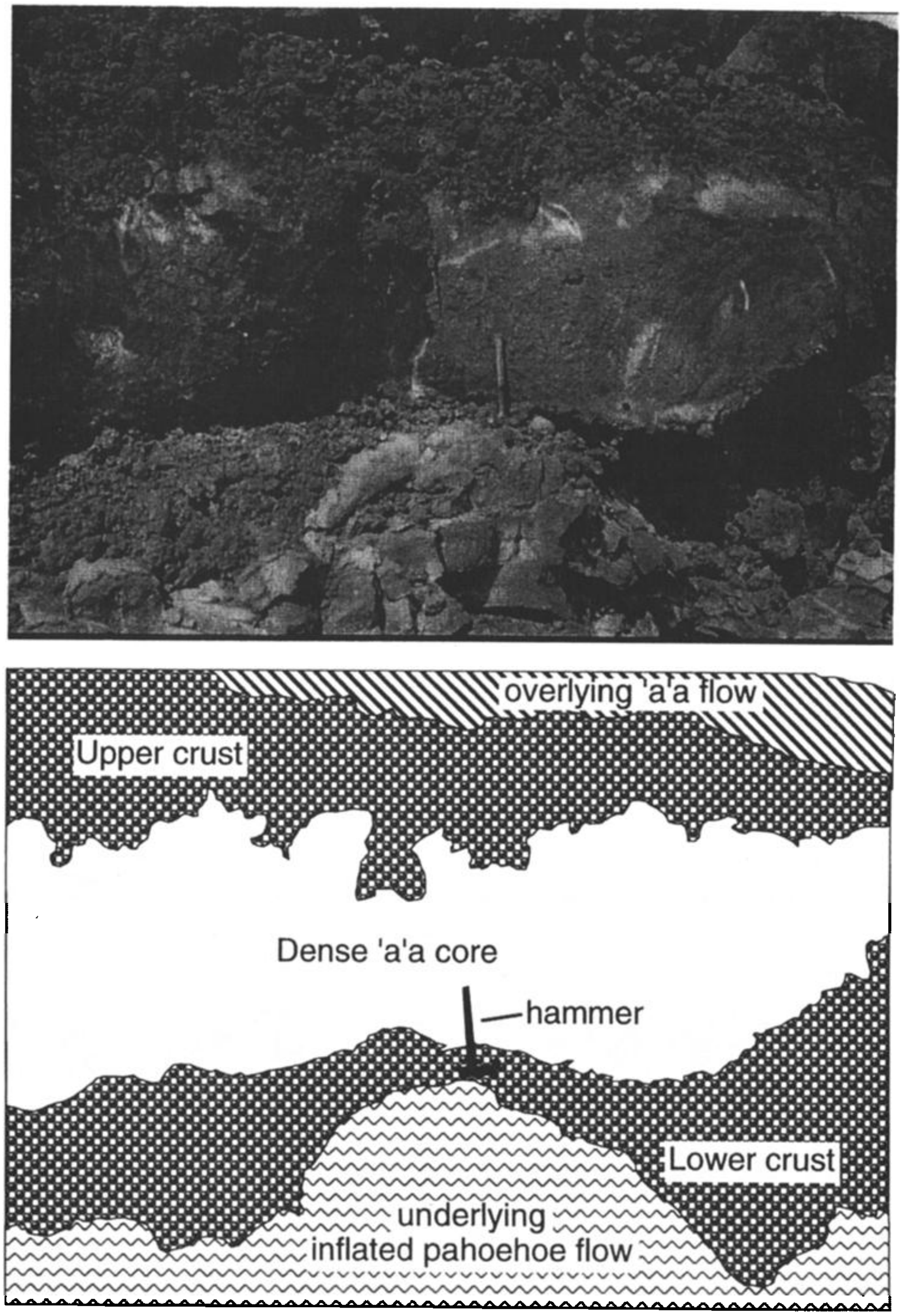
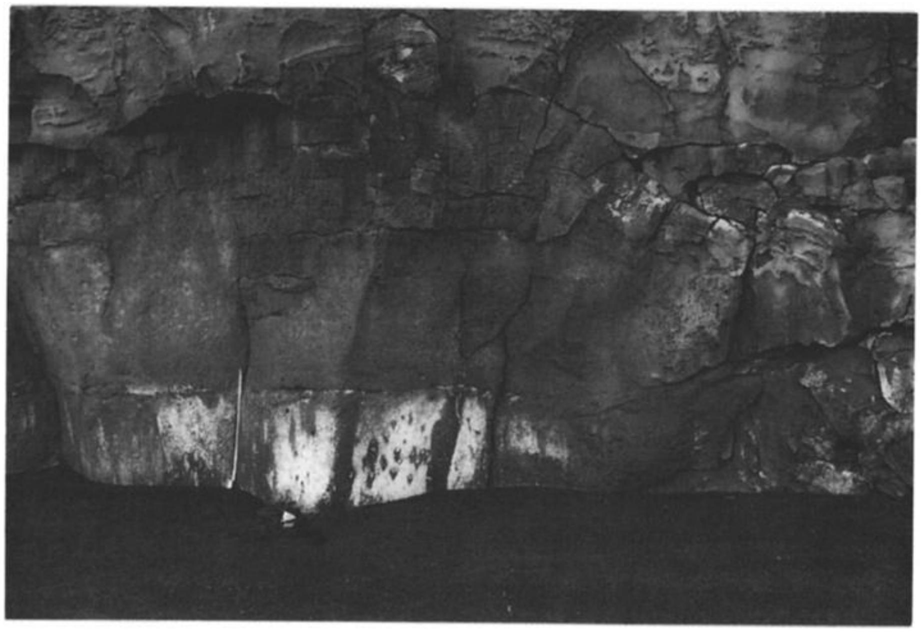


Figure 7. Photo and sketch of the cross section of a typical aa flow. Road cut through a 1972 aa flow within the 1969-1974 Mauna Ulu flow field, Kilauea Volcano, at approximately mile marker 11 on Chain of Craters Highway. Note the clinkery rubble at the flow top and flow base and the irregular contact with the dense core. This irregular contact is largely the result of the mixing between the outer crust and the inner core.

the transition is controlled by a combination of lava viscosity and strain rate (with high viscosities and strain rates both favoring aa formation), they do not place any quantitative constraints on the location of this transition. *Kilburn* [1981] calculated the strain rate versus effective viscosity of a single pahoehoe and single aa flow for an assumed Bingham lava rheology. However, his results are difficult to generalize because they rely on this assumed rheology and have only one example of each flow type. *Rowland and Walker* [1990] suggested that the pahoehoe to aa transition takes place at a volumetric effusion rate of 15-20 m^3/s for Hawaiian flows. However, they did not directly examine strain rate or viscosity. The fact that the transition between Hawaiian pahoehoe and aa can be defined by a single parameter is probably an artifact of the relatively narrow range of slopes, eruption styles, and viscosities of the historical flows examined by *Rowland and Walker* [1990]. For example, the 1783-1784 Laki eruption in Iceland is largely pahoehoe but



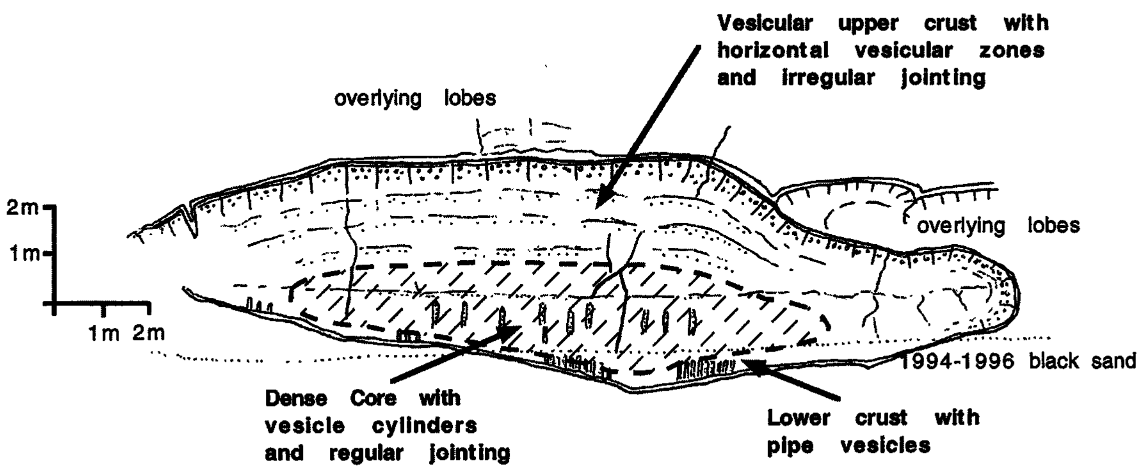


Figure 8. Photograph and sketch of the cross section of a typical inflated pahoehoe flow lobe. Lobe is in ~500-year-old Kane Nui o Hamo flow at the base of a seacliff near Lae'apuki, Kilauea Volcano, Hawaii. Sketch shows the partition of the flow into three sections: the upper crust, core, and lower crust. These same partitions exist in inflated pahoehoe lobes ranging from < 1 m to > 30 m in thickness. Note that the upper crust and the core each make up approximately half the flow thickness and the lower crust is typically 10-30 cm thick. Detailed descriptions of each section and interpretations of their formation are given in *Thordarson* [1995], *Self and Thordarson*, [1996], or *Self et al.* [1997, 1998].

had an average eruption rate of about 2500 m^3/s for 4 months [*Thordarson and Self*, 1993]. We conclude that demonstrating that a long lava flow is dominantly pahoehoe or aa is highly suggestive of its mode of emplacement but, by itself, is far from a definitive criterion.

4.2. Other Morphologic Criteria

There are distinctive morphologic features resulting from rapid emplacement, whether the flow be aa or otherwise. As noted earlier, both historical observations of active flows and laboratory experiments suggest that rapid flows form open channels. When the surface of a flow is well preserved, it is almost trivial to identify lava channels. Channels are also relatively straightforward to identify in cross-section, though they can be confused with dikes and faults [*Linneman and Borgia*, 1993].

Channels laterally confine a lava flow and the difference in planform between channelized flows and tube-fed or sheet-fed flows can be very dramatic, especially on shallow slopes (Figure 9). Channelized flows often fail to spread out even on very shallow slopes, forming long, narrow, finger-like lobes [cf. *Swanson et al.*, 1979]. In contrast, on shallow slopes, sheet and tube-fed flows have a remarkable tendency to wander about, efficiently filling in an entire area [e.g., *Mattox et al.*,

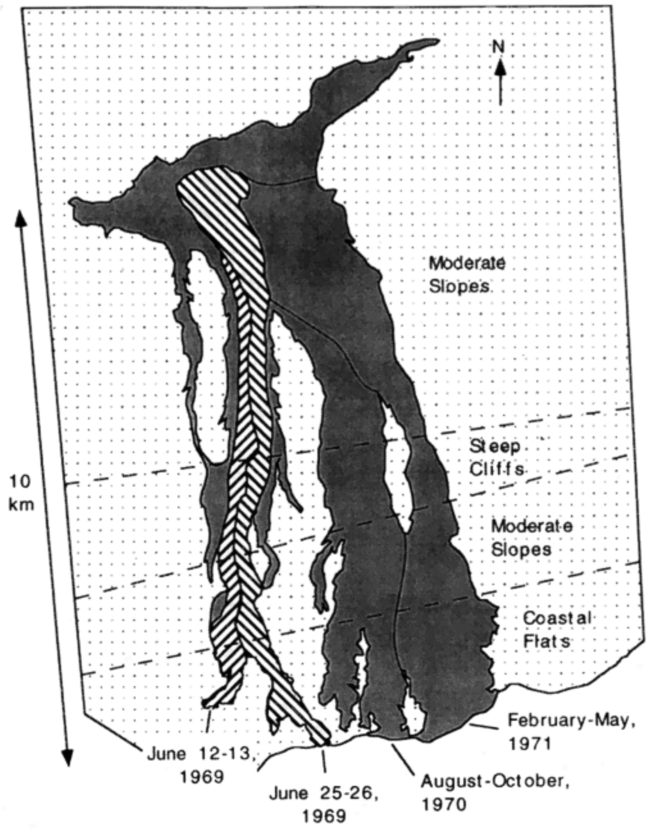


Figure 9. Map of 1969-1972 Mauna Ulu flow field, Kilauea Volcano, after *Swanson et al.* [1979]. The two striped flows on the west are channelized aa flows emplaced June 12-13, 1969, and June 25-26, 1969. The area to the east is dominantly tube-fed pahoehoe, mostly emplaced during August-October 1970 and February-May 1971. Note how the channelized aa failed to spread out even on the shallow slopes (<1%) of the coastal flats. In contrast, the tube-fed pahoehoe covers a broader area, especially at the coast. This is a reflection of the differences in both the dynamics and typical eruption durations of channelized versus tube-fed flows. We expect rapidly emplaced flows to resemble the channelized aa.

1993]. This difference in planform is also partly the result of the generally shorter duration of channelized aa flows relative to tube-fed pahoehoe flow fields.

If the channelized flow is rapid and turbulent, thermal erosion may be expected to deepen the channel [e.g., *Hulme*, 1973; *Huppert et al.*, 1984; *Jarvis*, 1995; *Greeley et al.*, this issue]. Thermally eroded channels are predicted to be somewhat triangular in cross section, with a wide base, since the flow will erode the sides as well as the base of the channel [*Jarvis*, 1995]. Extensive erosion of the substrate should also lead to detectable chemical contamination of the lava flows [e.g., *Huppert et al.*, 1984].

Lava tubes and sheets do not leave surface expressions that are as easy to identify as channels. Only occasionally are lava tubes marked by elongate tumuli [*Keszthelyi and Pieri*, 1993; *Hon et al.*, 1994; *Stephenson*, 1996] and sheets are, by definition, characterized by a relatively featureless surface. Both lava tubes and sheets are best identified in cross section. Drained lava tubes leave unmistakable caves, but lava tubes often do not drain on the shallow slopes crossed by long lava flows. However, the caves of the Undara tube system in a 160-km-long flow [*Atkinson and Atkinson*, 1995] provide an excellent exception. Given cross-sectional exposures, filled tubes can be recognized by concentric variations in vesicularity and crystal sizes. Filled tubes on shallow slopes in Hawaii and elsewhere tend to be lenticular or elliptical in cross section, with the long axis of the ellipse horizontal (Figure 10a). In contrast, where the tube is able to erode its base, the cross section becomes more elongate vertically (Figure 10b). The temporal evolution of lava tubes from being elongate in the horizontal dimension to becoming elongate in the vertical dimension is described by *Kauahikaua et al.* [this issue].

Historically, effusion rates approaching those required to form a long lava flow in the rapid mode have only been produced while curtains or fountains of lava were active. We cannot demonstrate that all rapidly emplaced lava flows must have been fed by vents with extensive pyroclastic activity, but it does seem to be a reasonable suggestion. It is interesting to note that the very high effusion rates required to emplace a long lava flow in the rapid mode might not require fountains much taller than those in historical eruptions. *Wilson and Head* [1981] show that fountain height depends on effusion rate raised to only the 1/4 power and thus effusion rates a hundred times greater than in historical eruptions would feed fountains only about 3 times taller.

For the insulated mode of emplacement to produce a long lava flow, a long uninterrupted effusive period is required. Pauses in the eruption of only a few days are often not sufficient to cause the tube system to break down on Kilauea Volcano [*Mattox et al.*, 1993], but longer-scale fluctuations in effusion rate can destroy even well-established tube systems [*Kauahikaua et al.*, 1996]. Broad sheet flows are probably also robust to minor pauses, but major fluctuations in the flux of lava will cause the crust to break up and lava to break out of the sheet. Historically, long-lived steady eruptions are more commonly fed from a lava pond or other point source rather than directly from a fissure [e.g., *Swanson et al.*, 1979; *Heliker and Wright*, 1991]. While the bulk of a long-lived eruption is expected to feed through a relatively quiescent point source, the early and terminal phases can be quite different, confusing the final near-vent morphology. Furthermore, during an eruption that lasts many months or years, the vent can migrate [e.g., *Heliker and Wright*, 1991; *Thordarson and Self*, 1993]. While the eruption must have long uninterrupted periods in order to form a long, stable lava tube system, each vent could have several episodes of activity. Thus a long-lived lava flow emplaced in the insulated mode may become a complex flow field, made up of many different individual flows from multiple episodes of activity from a number of vents [*Self et al.*, 1997, 1998].

5. Discussion and Conclusions

We have investigated two modes in which long lava flows can be produced and term these "rapid" and "insulated" emplacement. The rapid mode can produce lava flows >100 km long with less than 0.5°C/km of cooling given flow velocities >2-15 m/s, channel depths >2-19 m, and effusion rates >200-17,000 m^3/s. The insulated mode requires flow velocities >0.1-1.4 m/s, flows thicker than 2-23 m, and effusion rates >8-7100 m^3/s. The range in these constraints comes from.

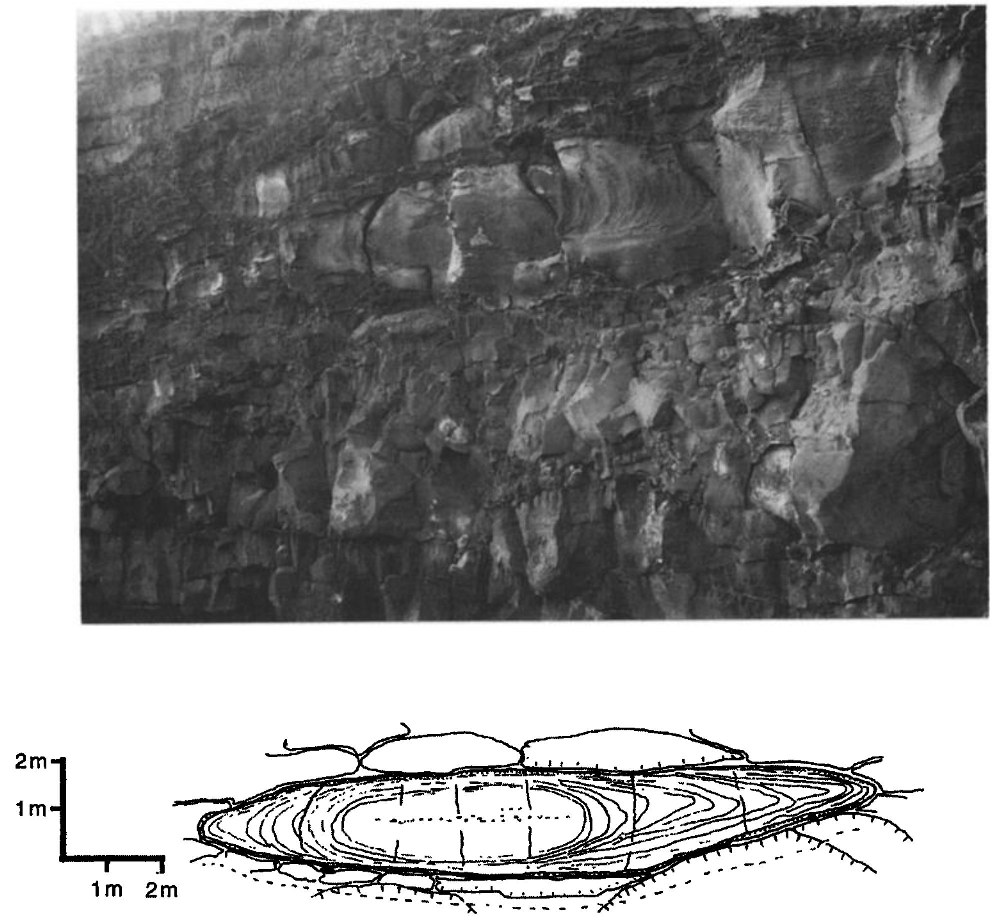


Figure 10a. Photo and sketch of the cross-section of a lava tube in the ~500-year-old Kane Nui o Hamo flows in a sea cliff near Lae'apuki, Kilauea Volcano. Note the broad, elliptical, shape of the filled tube emplaced on the shallow slopes (<1%) of the coastal flats. The tube is delineated by concentric vesicle and crystallization fronts that formed as the tube gradually froze shut.

many model parameters, but slope is the most important Steeper slopes produce higher velocities but thinner flows, requiring lower minimum effusion rates. These extremes in emplacement style should be readily distinguishable in the field. Lava flows emplaced in the rapid mode are expected to be channel fed, probably with aa surface and internal textures, evidence for thermal erosion along the base and walls of the channel(s), and extensive fire fountaining at a relatively short lived vent. We expect most long lava flows emplaced in the insulated mode to produce a complex flow field of tube-fed and/or sheet-like inflated pahoehoe flows, supplied from a number of relatively long-lived lava ponds or shields.

These results bring us one step closer to understanding why and how long basaltic lava flows form. However, many additional (and tighter) constraints on the emplacement of these flows can be found by investigating the formation and evolution of lava transport systems. For example, tighter constraints on eruption duration of pahoehoe sheet flows can come from the thickness of the insulating crust. *Hon et al.* [1994] show that a 1 m thick crust will take a week to form, a 5 m thick crust about 6 months and a 10 m thick crust almost 2 years. Thus, it is possible to use the thickness of the upper vesicular crust of a sheet flow to estimate eruption duration [*Thordarson*, 1995; *Thordarson and Self*, 1996; *Self et al.*, 1997; *Cashman and Kauahikaua*, 1997]. When this technique is applied to the 1300 km^3 pahoehoe sheet flows of the Roza Member of the Columbia River Basalt Group [*Tolan et al.*, 1989], a minimum eruption duration of about a decade and a maximum effusion rate of about 4000 m^3/s are derived [*Thordarson*, 1995; *Self et al.*, 1996b]. Unfortunately, since the widths of individual sheets in the Roza flows cannot easily be measured, it is not currently possible to directly compare this number to our model results. Other constraints may arise from the fact that the thermal efficiency of a sheet flow should increase as the crust thickens with time. If a sheet flow advances too rapidly, then a large portion of the sheet is too young and thermally inefficient to feed a long flow. As shown by *Kauahikaua et al.* [this issue], the formation of a roughly cylindrical lava tube also takes time and quantifying this could lead to additional constraints on the duration over which tube-fed flows were emplaced.

We end by noting that this study raises more questions than it has answered. We have suggested that historical eruptions could have formed transport systems efficient enough for lava to travel >100 km before freezing. Yet such long flows have not formed in recorded human history and appear rather rare even in the geologic record. Why? We hypothesize that scarcity of long flows is related to the paucity of effusive basaltic eruptions with volumes greater than 10 km^3.

It is also intriguing that all the well-preserved long lava

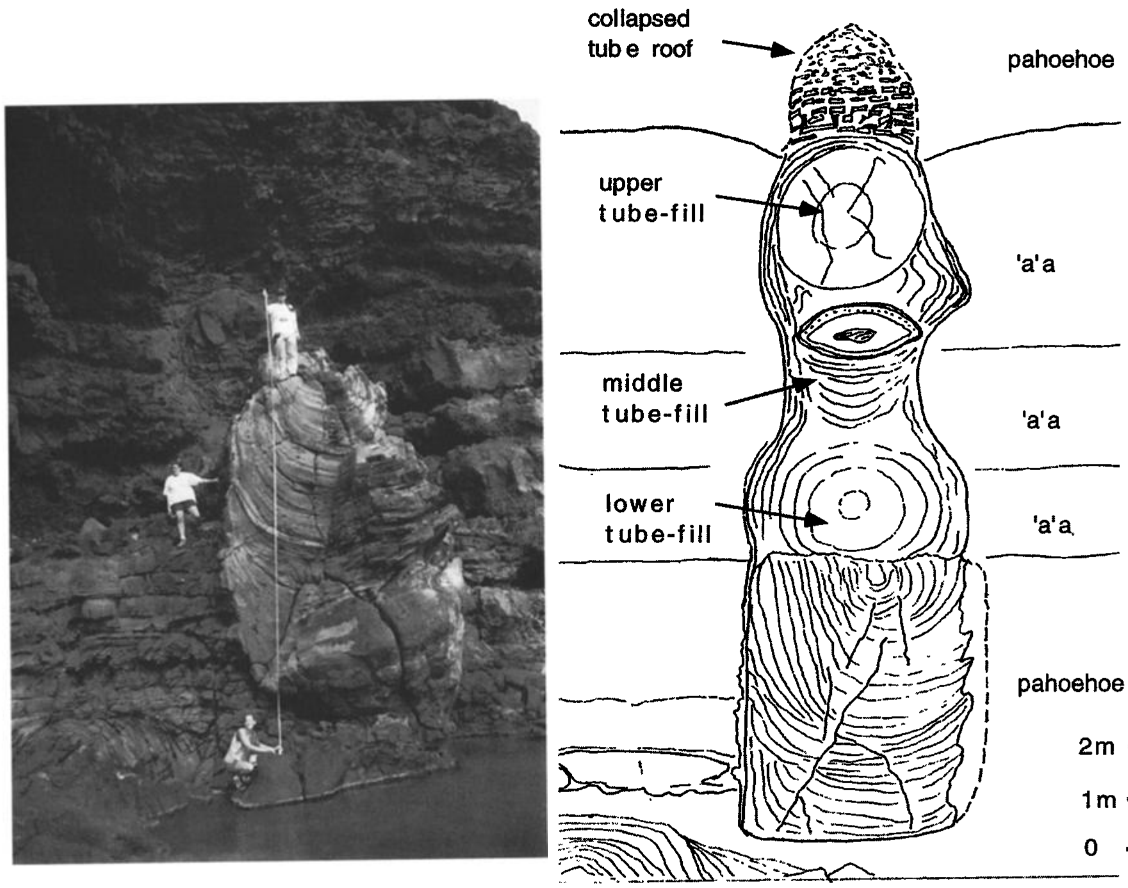


Figure 10b Photo and sketch of the cross section of filled tube emplaced on a steep slope (>10%) at Makapu`u, O`ahu. The tube is thought to have formed in the uppermost pahoehoe flows and eroded down >12 m. Note how the thermal and mechanical erosion has formed a tube elongate in the vertical direction and somewhat flared outward at the base. We expect lava tubes in most long lava flows to more closely resemble the Kilauea example, not the Makapu`u tube.

flows on the Earth appear to be inflated pahoehoe emplaced in the insulated mode. Have there never been large volume eruptions with sustained effusion rates substantially higher than those seen historically? Or are aa-like flows erased more quickly from the geologic record than pahoehoe-like sheet flows? While long flows emplaced in the rapid mode might not exist on the Earth, a silicate lava flow on Io was observed to cover approximately 10^5 km^2 in 5 hours [*Blaney et al.*, 1995; *Davies*, 1996]. We expect additional interesting results when the effects of planetary environments and nonbasaltic compositions (e.g., komatiites or andesites) are investigated.

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References

Allen, J. E., The Carrizozo Malpais (abstract), *Roswell Geol. Soc. 5th Field Conf. Guidebook*, pp. 9-11, Roswell Geol. Soc., Roswell, N. M., 1952.

Atkinson, A., and V. Atkinson, Undara volcano and its lava tubes, 86 pp., Vernon and Anne Atkinson, Brisbane, Australia, 1995.

Aubele, J. C., L. S. Crumpler, and W. E. Elston, Vesicle zonation and vertical structure of basalt flows, *J. Volcanol. Geotherm. Res.*, *35*, 349-374, 1988.

Baloga, S., P. D. Spudis, and J. E. Guest, The dynamics of rapidly emplaced terrestrial lava flows and implications for planetary volcanism. *J. Geophys. Res.*, *100*, 24,509-24,519, 1995.

Barnard, W. M., Mauna Loa Volcano: Historical eruptions, exploration, and observations (1779-1910), in *Mauna Loa Revealed, Geophys. Monogr. Ser.*, vol. 92, edited by J. M. Rhodes and J. P. Lockwood, pp. 1-19, AGU, Washington, D.C., 1995.

Bird, R. B., W. E. Stewart, and E. N. Lightfoot, *Transport Phenomena*, 780 pp., John Wiley, New York, 1960.

Blaney, D. L., T. V. Johnson, D. L. Matson, and G. J. Veeder, Volcanic eruptions on Io: Heat-flow, resurfacing, and lava composition, *Icarus*, *113*, 220-225, 1995.

Booth, B., and S. Self, Rheological features of the 1971 Mount Etna lavas, *Philos. Trans. R. Soc. London, Ser. A*, *274*, 99-106, 1973.

Bruno, B. C., Lava flow dynamics: Clues from fractal analysis, Ph.D. thesis, 247 pp., Univ. of Hawai'i at Manoa, Honolulu, 1994.

Bruno, B. C., and G. J. Taylor, Morphologic identification of Venusian lavas, *Geophys. Res. Lett.*, *22*, 1897-1900, 1995.

Bruno, B. C., G. J . Taylor, S. K. Rowland, P. G. Lucey, and S. Self, Lava flows are fractals, *Geophys. Res. Lett.*, *19*, 305-308, 1992.

Calvari, S., and H. Pinkerton, Formation of lava tubes and extensive flow field during the 1991-1993 eruption of Mount Etna, *J. Geophys. Res.*, this issue.

Campbell, B. A., and D. B. Campbell, Analysis of volcanic surface morphology on Venus from comparison of Arecibo, Magellan, and terrestrial airborne radar data, *J. Geophys. Res., 97*, 16,293-16,314, 1992.

Cashman, K. V., and J. P. Kauahikaua, Re-evaluation of vesicle distributions in basaltic lava flows, *Geology, 25*, 419-422, 1997.

Chester, D. K., A. M. Duncan, J. E. Guest, and C. R. J. Kilburn, *Mount Etna: The Anatomy of a Volcano*, 404 pp., Stanford Univ. Press, Stanford, Calif., 1985.

Crisp, J., and S. Baloga, A model for lava flows with two thermal components, *J. Geophys. Res.*, 95, 1255-1270, 1990.

Crisp, J., and S. M. Baloga, Influence of crystallization and entrainment of cooler material on the emplacement of basaltic aa flows, *J. Geophys. Res., 99*, 11,819-11,831, 1994.

Crisp J., K. V. Cashman, J. A. Bonini, S. B. Hougen, and D. C. Pieri, Crystallization history of the 1984 Mauna Loa lava flow, *J. Geophys. Res., 99*, 7177-7198, 1994.

Danes, Z. F., Dynamics of lava flows, *J. Geophys. Res., 77*, 1420-1432, 1972.

Davies, A. G., Io's volcanism: Thermo-physical models of silicate lava compared with observations of thermal emission, *Icarus, 124*, 45-61, 1996.

Dragoni, M., Modeling the rheology and cooling of lava flows, in *Active lavas: Monitoring and Modeling*, edited by C. R. J. Kilburn and G. Luongo, pp. 235-262, UCL Press, London, 1993.

Einarsson, T., The flowing lava: Studies of its main physical and chemical properties, in *The Eruption of Hekla 1947-1948*, vol. IV(3), edited by T. Einarsson, G. Kjartansson, and S. Thorarinsson, pp 1-71., Visindafelg Islendinga, Reykjavik, Iceland, 1949.

Finch, R. H., and G. A. Macdonald, Hawaiian volcanoes during 1950, *U.S. Geol. Surv. Bull., 996-B*, 27-89, 1953.

Fink, J. H., and J. R. Zimbelman, Rheology of the 1983 Royal Gardens basalt flows, Kilauea Volcano, Hawaii, *Bull. Volcanol., 48*, 87-96, 1986.

Friedman, R. C., Petrologic clues to lava flow emplacement and post-emplacement processes, Ph.D. thesis, 260 pp., Univ. of Hawai'i at Manoa, 1998.

Goncharov, V. N., *Dynamics of Channel Flow*, 317 pp., translated from Russian by Israel Program Sci. Transl., U.S. Dep. of Commer., Off. of Tech. Serv., Washington, D. C., 1964.

Greeley R., S. A. Fagents, R. S. Harris, S. D. Kadel, D. A. Williams, and J. E. Guest, Erosion by flowing lava: Field evidence, *J. Geophys. Res.*, this issue.

Gregg, T. K. P., and J. H. Fink, Quantification of extraterrestrial lava flow effusion rates through laboratory simulations, *J. Geophys. Res., 101*, 16,891-16,900, 1996.

Griffiths, R. W., and J. H. Fink, The morphology of lava flows in planetary environments: Predictions from analog experiments, *J. Geophys. Res., 97*, 19,739-19,748, 1992.

Guest, J. E., P. D. Spudis, R. Greeley, G. J . Taylor, and S. M. Baloga, Emplacement of xenolith nodules in the Kaupulehu lava flow, Hualalai Volcano, Hawaii, *Bull. Volcanol., 57*, 179-184, 1995.

Guest, J. E., M. G. Lancaster, D. L. Copp, and E. R. Stofan, Emplacement mechanisms for great lava flows on the plains of Venus: Slow vs. fast (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 16-17, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Hallworth, M. A., H. E. Huppert, and R. S. J. Sparks, A laboratory simulation of basaltic lava flows, *Mod. Geol., 11*, 93-107, 1987.

Harris, A. J. L., L. P. Flynn, L. Keszthelyi, P. J. Mouginis-Mark, S. K. Rowland and J. A. Resing, Calculation of lava effusion rates from LANDSAT TM data, *Bull. Volcanol.*, in press, 1998.

Head, J. W., and L. Wilson, Volcanic processes and landforms on Venus: Theory, predictions, and observations, *J. Geophys. Res., 91*, 9407-9446, 1986.

Heliker, C., and T. L. Wright, The Pu'u `O'o - Kupaianaha eruption of Kilauea, *Eos Trans. AGU, 72*, 521, 526, 530, 1991.

Heslop, S. E., L. Wilson, H. Pinkerton, and J. W. Head, Dynamics of a confined lava flow on Kilauea Volcano, Hawaii, *Bull. Volcanol., 51*, 415-432, 1989.

Hill, R. E. T., and C. S. Perring, The evolution of Archean Komatiite flow fields: Are they inflationary sheet flows (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 18-21, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Ho, A. M., and K. V. Cashman, Temperature constraints on the Ginko flow of the Columbia River Basalt Group, *Geology, 25*, 403-406, 1997.

Hon, K., J. Kauahikaua, R. Denlinger, and K. Mackay, Emplacement and inflation of pahoehoe sheet flows: Observations and measurements of active lava flows on Kilauea Volcano, Hawaii, *Geol. Soc. Am. Bull., 106*, 351-370, 1994.

Hulme, G., Turbulent lava flow and the formation of lunar sinuous rilles, *Mod. Geol., 4*, 107-117, 1973.

Hulme, G., Interpretation of lava flow morphology, *Geophys. J. R. Astron. Soc., 39*, 361-383, 1974.

Huppert, H. E., R. S. J. Sparks, J. S. Turner, and N. T. Arndt, Emplacement and cooling of komatiite lavas, *Nature, 309*, 19-22, 1984.

Jarvis, R. A., On the cross-sectional geometry of thermal erosion channels formed by turbulent lava flows, *J. Geophys. Res., 100*, 10,127-10,140, 1995.

Jeffreys, H., The flow of water in an inclined channel of rectangular section, *Philos. Mag., 49*, 793-807, 1925.

Jurado-Chichay, Z., and S. K. Rowland, Channel overflows of the Pohue Bay flow, Mauna Loa, Hawai'i: Examples of the contrast between surface and interior lava, *Bull. Volcanol., 57*, 117-126, 1995.

Kauahikaua, J., M. Mangan, C. Heliker, and T. Mattox, A quantitative look at the demise of a basaltic vent: The death of Kupaianaha, Kilauea Volcano, Hawai'i, *Bull. Volcancl., 57*, 641-648, 1996.

Kauahikaua, J., K. V. Cashman, K. A. Hon, T. N. Mattox, C. C. Heliker, M. T. Mangan, and C. R. Thornber, Observations on basaltic lava stream dynamics in tubes from Kilauea Volcano, island of Hawai'i, *J. Geophys. Res.*, this issue.

Keszthelyi, L., Are lava tubes air-cooled? (abstract) *Eos Trans. AGU, 73*(43), Fall Meet. Suppl., 648, 1992.

Keszthelyi, L., A preliminary thermal budget for lava tubes on the Earth and planets, *J. Geophys. Res. 100*, 20,411-20,420, 1995.

Keszthelyi, L., Measurements of the cooling of pahoehoe lava flows (abstract), *Eos Trans. AGU, 77*(46), Fall Meet. Suppl., F807, 1996.

Keszthelyi, L., and R. Denlinger, The initial cooling of pahoehoe lava flow lobes, *Bull. Volcanol., 58*, 5-18, 1996.

Keszthelyi, L., and A. S. McEwen, Thermal models for basaltic volcanism on Io, *Geophys. Res. Lett., 24*, 2463-2466, 1997.

Keszthelyi, L. P., and D. C. Pieri, Emplacement of the 75-km-long Carrizozo lava flow field, south-central New Mexico, *J. Volcanol. Geotherm. Res., 59*, 59-75, 1993.

Keszthelyi, L., S. Self, and T. Thordarson, Application of recent studies on the emplacement of basaltic lava flows to the Deccan Traps, *Mem. Geol. Soc. India*, in press, 1998.

Kilburn, C. R. J., Pahoehoe and aa lavas: A discussion and continuation of the model of Peterson and Tilling. *J. Volcanol. Geotherm. Res., 11*, 373-382, 1981.

Kilburn, C. R. J., Surfaces of aa flow fields on Mount Etna, Sicily: Morphology, rheology, crystallization and scaling phenomena. in *Lava Flows and Domes*, edited by J. H. Fink, *IAVCEI Proc. Volcanol., 2*, 129-156, 1990.

Kilburn, C. R. J., Patterns and predictability in the emplacement of subaerial lava flow and flow fields, in *Monitoring and Mitigation of Volcano Hazards*, edited by R. Scarpa and R. I. Tilling, pp. 491-537, Springer-Verlag, New York, 1996.

Kilburn, C. R. J., and R. M. C. Lopes, General patterns of flow field growth: Aa and blocky lavas, *J. Geophys. Res., 96*, 19,721-19,732, 1991.

Komatsu, G., J. S. Kargel, and V. R. Baker, Canali-type channels on Venus: Some genetic constraints. *Geophys. Res. Lett., 19*, 1415-1418, 1992.

Linneman, S. R., and A. Borgia, The blocky andesite lava flows of Arenal Volcano, Costa Rica, in *Active Lavas*, edited by C. J. Kilburn and G. Luongo, pp. 25-72, UCL Press, London, 1993.

Lipman, P. W., and N. G. Banks, Aa flow dynamics, Mauna Loa 1984, *U.S. Geol. Surv. Prof. Pap., 1350*, 1527-1565, 1987.

Macdonald, G. A., Pahoehoe, aa, and block lava, *Am. J. Sci., 251*, 169-191, 1953.

Malin, M. C., Lengths of Hawaiian lava flows, *Geology, 8*, 306-308, 1980.

Mangan, M. T., and R. T. Helz, The distribution of vesicles and olivine phenocrysts in samples from drill hole KI 79-3, Kilauea Iki lava lake, Hawaii. *U.S. Geol. Surv. Open File Rep., 86-424*, 25 pp., 1986.

Marks, L., and T. Baumeister (Eds.) *Marks' Mechanical Engineers' Handbook*, 6th ed., pp. 3-81, McGraw Hill, New York, 1958.

Marsh, B. D., On the crystallinity, probability of occurrence, and rheology of lava and magma, *Contrib. Mineral. Petrol., 78*, 85-98, 1981.

Mattox, T. N., C. Heliker, J. Kauahikaua, and K. Hon, Development of the 1990 Kalapana flow field, Kilauea Volcano, Hawaii, *Bull. Volcanol., 55*, 407-413, 1993.

McBirney, A. R., and T. Murase, Rheological properties of magmas, *Annu. Rev. Earth Planet. Sci., 12*, 337-357, 1984.

Miller, J., T. Thordarson, and G. Larsen, Sulfur degassing and nature of eruptive activity during the 935AD Eldgja eruption, S-Iceland, *Eos Trans. AGU (abstracts), 77*, F803, 1996.

Miyamoto, H., and S. Sasaki, Numerical study of lava flow: Factors controlling flow morphology abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 42-43, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996..

Moore, H. J., Preliminary estimates of the rheological properties of the 1984 Mauna Loa lava, *U.S. Geol. Surv. Prof. Pap., 1350*, 1569-1588, 1987.

Peck D. L., Cooling and vesiculation of Alae lava lake, Hawaii. *U.S. Geol. Surv. Prof. Pap., 935B*, B148-B151, 1978.

Peterson, D. W., and R. I. Tilling, Transition of basaltic lava from pahoehoe to aa, Kilauea Volcano, Hawaii: Field observations and key factors. *J. Volcanol. Geotherm. Res., 7*, 271-293, 1980.

Pieri, D. C., and S. M. Baloga, Eruption rate, area, and length relationships for some Hawaiian lava flows, *J. Volcanol. Geotherm. Res., 30*, 29-45, 1986.

Pike, R. J., Volcanoes on the inner planets: Some preliminary comparisons of gross topography, *Proc. Lunar Planet. Sci. Conf., 9th*, 3239-3273, 1978.

Pinkerton, H., and R. S. J. Sparks, Field measurements of the rheology of lava, *Nature. 276*, 383-384, 1978.

Pinkerton, H., and R. Stevenson, Methods of determining the rheological properties of lavas from their physico-chemical properties, *J. Volcanol. Geotherm. Res., 53*, 47-66, 1992.

Pinkerton, H., and L. Wilson, Factors controlling the lengths of channel-fed lava flows, *Bull. Volcanol., 56*, 108-120, 1994.

Pinkerton, H., R. A. Herd, R. M. Kent, Field measurements of the rheological properties of basaltic lavas on Kilauea, Hawaii (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 52-53, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Reynolds, A. J., *Turbulent Flow in Engineering*, 452 pp., John Wiley, New York, 1974.

Rowland, S. K., and G. P. L. Walker, Pahoehoe and aa in Hawaii: Volumetric flow rate controls the lava structure, *Bull. Volcanol., 52*, 615-628, 1990.

Sakimoto, S. E. H., and S. M. Baloga, Thermal controls on tube-fed planetary lava flow lengths (abstract), *Lunar Planet. Sci. Conf., XXIV*, 1217-1218, 1995

Sakimoto, S. E. H., and M. T. Zuber, Flow and convective cooling in lava tubes, *J. Geophys. Res.*, this issue.

Sakimoto, S. E. H., S. Baloga, and J. Crisp (abstract), Viscosity and flow rate constraints for tube-fed planetary lava flows, *Lunar Planet. Sci. Conf., XXVIII*, 1231-1232, 1997a.

Sakimoto, S. E. H., J. Crisp, and S. M. Baloga, Eruption constraints on tube-fed planetary lava flows, *J. Geophys. Res., 102*, 6597-6613, 1997b.

Self, S., T. Thordarson, and L. Keszthelyi, Inflation features as keys to the emplacement of pahoehoe lava flows fields in the Columbia River Basalt Group (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 72-73, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Self, S., T. Thordarson, L. Keszthelyi, G. P. L. Walker, K. Hon, M. T. Murphy, P. Long, and S. Finnemore, A new model for the emplacement of the Columbia River Basalt as large, inflated pahoehoe sheet lava flow fields, *Geophys. Res. Lett., 23*, 2689-2692, 1996b.

Self, S., T. Thordarson, and L. Keszthelyi, Emplacement of continental flood basalt lava flows, in *Large Igneous Provinces, Geophys. Monogr. Ser.*, vol. 100, edited by J. J. Mahoney and M. Coffin, pp. 381-410, AGU, Washington, D.C., 1997.

Self, S., L. Keszthelyi, and T. Thordarson, The importance of Pahoehoe, *Ann. Rev. Earth Planet. Sci., 26*, 81-110, 1998.

Shaw, H., Rheology of basalt in the melting range, *J. Petrol., 10*, 510-535, 1969.

Shaw, H., Viscosities of magmatic liquids: An empirical method of prediction, *Am. J. Sci., 272*, 870-893, 1972.

Shaw, H., and D. Swanson, Eruption and flow rates of flood basalts, in *Proceedings of the Second Columbia River Basalt Symposium*, edited by E. Gilmour and D. Stradling, editors, pp. 271-299, East. Wash. State College Press, Cheney, 1970.

Shaw, H. R., T. L. Wright, D. L. Peck, and R. Okamura, The viscosity of basaltic magma: An analysis of field measurements in Makaopuhi lava lake, Hawaii, *Am. J. Sci., 266*, 225-264, 1968.

Simkin, T., and L. Siebert, *Volcanoes of the World*, 349 pp., Geosci. Press, Tucson, Ariz., 1994.

Spry, M., and P. J. Stephenson, Segregation structures in the Toomba basalt, north Queensland (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 78-79, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Stephenson, P. J., Ridges and depressions in the Undara lava flow, N. Queensland: Alternative mechanisms for their formation (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 82-83, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Stephenson, P. J., and P. W. Whitehead, Lava pits in the Toomba basalt, north Queensland (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 80-81, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Stephenson, P. J., A. T. Burch-Johnson, and D. Stanton, Long lava flows in north Queensland: Context, characteristics, emplacement (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 86-87, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Swanson, D., T. Wright, and R. Helz, Linear vent systems and estimated rates of magma production and eruption for the Yakima basalt on the Columbia Plateau, *Am. J. Sci., 275*, 877-905, 1975.

Swanson, D. A., W. A. Duffield, D. B. Jackson, and D. W. Peterson, Chronological narrative of the 1969-1971 Mauna Ulu eruption of Kilauea Volcano, Hawaii, *U.S. Geol. Surv. Prof. Pap., 1056*, 55 pp., 1979.

Theilig, E., and R. Greeley, Lava flows on Mars: Analysis of small surface features and comparisons with terrestrial analogs, Part 1, *Proc. Lunar Planet. Sci. Conf. 17th, J. Geophys. Res., 92*, suppl., E193-E206, 1986.

Thordarson, T., Volatile release and atmospheric effects of basaltic fissure eruptions, Ph.D. thesis, 446 pp., Univ. of Hawai'i at Manoa, Honolulu, 1995.

Thordarson, T., and S. Self, The Laki (Skaftár Fires) and Grímsvötn eruptions in 1783-1785, *Bull. Volcanol., 55*, 233-263, 1993.

Thordarson, T., and S. Self, Sulfur, chlorine, and fluorine degassing and atmospheric loading by the Roza eruption, Columbia River Basalt Group, Washington, USA, *J. Volcanol. Geotherm. Res., 74*, 49-73, 1996.

Tolan, T. L., S. P. Reidel, M. H. Beeson, J. L. Anderson, K. R. Fecht, and D. A. Swanson, Revisions to the estimates of the areal extent and volume of the Columbia River Basalt Group, in *Volcanism and Tectonism in the Columbia River Flood-Basalt Province*, edited by S. P. Reidel and P. R. Hooper, *Spec. Pap. Geol. Soc. Am. 239*, 1-20, 1989.

Vilmundardottir, E. G., Tungnarhraun, Orustofnun geologic report, *Rep. OS ROD 7702*, 156 pp., Reykjavik, 1977.

Walker, G. P. L., Lengths of lava flows, *Philos. Trans. R. Soc. London, Ser. A, 274*, 107-118, 1973.

Walker, G. P. L., Structure, and origin by injection under surface crust, of tumuli, "lava rises," "lava-rise pits," and "lava inflation clefts" in Hawaii, *Bull. Volcanol., 53*, 546-558, 1991.

Walker, G. P. L., Morphometric study of pahoehoe lava flows (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 92-93, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Weir, G. J., Kinematic wave theory for Ruapehu lahars, New Zealand, *J. Sci., 25*, 197-203, 1982.

Wilson, L., and J. W. Head, Ascent and eruption of basaltic magma on the Earth and Moon, *J. Geophys. Res., 86*, 2971-3001, 1981.

Wilson, L., and J. W. Head, Mars: Review and analysis of volcanic eruption theory and relationship to observed landforms, *Rev. Geophys., 32*, 221-263, 1994.

Whitaker, S., *Introduction to Fluid Mechanics*, 457 pp., Robert E. Krieger, Malabar, Fla., 1981.

Whitehead, P. W., and P. J. Stephenson, Lava rise ridges of the Toomba basalt flow, north Queensland (abstract), in *Chapman Conference on Long Lava Flows: Conference Abstracts Long Lava Flows, Econ. Geol. Res. Unit Contrib. 56*, edited by P. W. Whitehead, pp. 94-95, Dep. of Earth Sci., James Cook Univ. of North Queensl., Townsville, Australia, 1996.

Wright T. L. , and R. T. Okamura, Cooling and crystallization of tholeiitic basalt, 1965 Makaopuhi lava lake, Hawaii. *U.S. Geol. Surv. Prof. Pap., 1004*, 78 pp., 1977.

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