The Transition from ‘A‘ā to Pāhoehoe Crust on Flows Emplaced During the Pu‘u ‘Ō‘ō-Kūpaianaha Eruption

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Abstract

The transition from pāhoehoe to ‘a‘ā along a single lava flow is widely accepted as irreversible in volcanology. However, channelized ‘a‘ā flows have been repeatedly observed changing into pāhoehoe flows during the low-effusion-rate “pāhoehoe” stage of the Pu‘u ‘Ō‘ō-Kūpaianaha eruption. This change most commonly occurs when flows move from steep slopes to level ground. The transition is marked by the change from ‘a‘ā to pāhoehoe crust on the solidified levees. It is here suggested that the definitions of pāhoehoe and ‘a‘ā be restricted to solidified crustal products, as those terms were originally defined. Active lava flows can be designated as either pāhoehoe flows or ‘a‘ā flows, depending on the type of crust that is being produced at the advancing front. Many difficulties arise from applying the terms pāhoehoe and ‘a‘ā to molten lava that has not taken a solid form. Hawaiian tholeiitic lava undergoes an irreversible transition from near-Newtonian to Bingham or other non-Newtonian behavior as it cools, crystallizes, and degases during transport away from the vent. Most Hawaiian pāhoehoe flows are fed by near-Newtonian lava, and most Hawaiian ‘a‘ā flows are fed by lava with Bingham or more complex rheologies. Within a restricted rheological range, however, Hawaiian lava can form either a pāhoehoe or an ‘a‘ā crust, depending on strain rate. This limited rheologic condition allows an ‘a‘ā flow to change into a pāhoehoe flow in response to a reduction in strain rate.

Introduction

Virtually every geologist recognizes the Hawaiian words pāhoehoe and ‘a‘ā, used to describe the two principal types of subaerial basaltic lava flows found throughout the world. Pāhoehoe has a continuous smooth, billowy, or ropy crust, and ‘a‘ā has a brecciated rough, spiny, or clinkery surface (Macdonald, 1953). Originally used to refer to the solidified lava-flow surfaces, the definitions were expanded to designate entire flows by Macdonald (1953, p. 170), who added “. . . the terms refer to the partly or completely solidified flows, though commonly the surfaces of the molten feeding rivers also are sufficiently distinct in appearance to make possible their recognition as one type or the other.” We suggest that the terms ‘a‘ā and pāhoehoe be restricted to solidified crusts and thatthe terms ‘a‘ā flow and pāhoehoe flow be used when discussing either active or inactive flows. We further recommend that none of these terms be applied to molten lava in channels, because that lava may go on to produce either pāhoehoe or ‘a‘ā.

Many Hawaiian lava flows erupt as pāhoehoe and undergo a change to ‘a‘ā as they flow away from the vent (fig. 1). This change is particularly common in short-lived flows with high effusion rates, and it is widely believed that this transition is irreversible (Macdonald, 1953; Peterson and Tilling, 1980; Kilburn, 2000). Macdonald (1953) clearly defined the transition as the place where the “solidified” part of the flow changed from predominantly pāhoehoe surfaces to ‘a‘ā surfaces. Thus, the pāhoehoe-‘a‘ā transition can be thought of as a facies change from pāhoehoe crust to ‘a‘ā crust along a single flow. Whereas Macdonald (1953) acknowledged that the material flowing in the channel generally corresponded to the material on the solidified flow levees, he noted that this did not necessarily have to be the case. Cashman and others (1999) proposed that the transition takes place in the channel, where the lava changes from a well-mixed fluid to a thermally stratified fluid with a cooler surface and edges and a hotter isothermal core. They suggested that the change related directly to an increase in yield strength resulting from an observed increase in the degree of plagioclase crystallinity (to more than 10 or 20 percent microlites).

A typical transitional crustal sequence along a single Mauna Loa flow goes from pāhoehoe to slabby pāhoehoe to scoriaceous-spinose ‘a‘ā and, finally, to clinker-dominated ‘a‘ā at the distal parts of the flows (Lipman and Banks, 1987). Wolfe and others (1988) reported similar transitions for the large ‘a‘ā flows from the early stages of Kīlauea’s Pu‘u ‘Ō‘ō eruption. In the large flows they described from Mauna Loa and Kīlauea, the surface textures of the active channels were roughly the same as those of the solidified levees, and the position of the pāhoehoe to ‘a‘ā transition remained in a relatively fixed position during the life of the flow (Lipman and Banks, 1987, p. 1529; Wolfe and others, 1988, p. 30). Across the pāhoehoe-to-‘a‘ā transition, vesicularity decreases, deformation of vesicles is greater, and plagioclase microlite crystallinity increases in the solidified lava (Polacci and others, 1999: Cashman and others, 1999).

During the past decade of the Pu‘u ‘Ō‘ō-Kūpaianaha eruption, we have observed numerous examples of the transition of active ‘a‘ā flows into active pāhoehoe flows. This transition is marked by a “reverse” facies change from clinkery ‘a‘ā to spiny ‘a‘ā to slabby pāhoehoe, to pāhoehoe. It is concluded that this process is relatively common, particularly during long-lived eruptions of low and/or variable effusion rates.

The Nature of the Pāhoehoe-‘A‘ā Transition

The original definition of the pāhoehoe-‘a‘ā transition was explicit in specifying it as being a change in flow types away from the vent. “A flow of pāhoehoe may change downslope to ‘a‘ā, but the change of ‘a‘ā to pāhoehoe has never been observed” (Macdonald, 1953, p. 183). The reasons given for this transition (Macdonald, 1953) are an increase in viscosity due to cooling and gas loss during transport away from the vent and the promotion of crystallization by mechanical stirring of the lava. Cashman and others (1999) provided an excellent review of the historical development of ideas regarding the pāhoehoe-‘a‘ā transition.

Peterson and Tilling (1980) elegantly demonstrated that the transition from pāhoehoe to ‘a‘ā is related not only to apparent viscosity, but also to applied shear stress and the rate of shear strain. The factors listed by Macdonald (1953), such as composition, temperature, crystallinity, melt polymerization, volatile content, and vesicularity, affect the apparent viscosity but are not the sole controls on the pāhoehoe-‘a‘ā transition in Hawai‘i. Peterson and Tilling (1980) showed that other variables, such as effusion rate, channel configuration, flow velocity and duration, and ground slope, are directly related to the applied shear stress and affect the shear-strain rate.

Peterson and Tilling (1980) analyzed the pāhoehoe-‘a‘ā transition by examining the change in shear-strain rate and apparent viscosity to which an individual “element” (infinitesimal part) of molten lava is subjected during the transition from pāhoehoe to ‘a‘ā. Because completely molten lava has the properties of neither pāhoehoe nor ‘a‘ā, determining the type of crust that will form upon cooling can be difficult. Furthermore, once a pāhoehoe crust forms, that “element” of lava is no longer capable of changing to ‘a‘ā. Instead, the change in an “element” of lava is really from a nearly Newtonian rheology to either a Bingham or other non-Newtonian rheology. This rheological change is irreversible, as alluded to by Kilburn (1981), but does not necessarily coincide with the pāhoehoe-to-‘a‘ā transition, which is a change in crustal state (Kilburn, 1990). The transition from a pāhoehoe flow to an ‘a‘ā flow, or the reverse, generally takes place at the front of the advancing flow, where most of the crust is being produced. However, the change in rheological behavior of an individual “element” of molten lava as it moves to the flow front may occur anywhere in the transport system, depending upon cooling, gas loss, and other factors. We retain here the original definitions of pāhoehoe and ‘a‘ā as crusts or flows and of the transition from pāhoehoe to ‘a‘ā as a change in flow type (Macdonald, 1953), which is similar to a facies change.

Figure 1. Pāhoehoe-to-‘a‘ā transition occurring in many small flows as they cascade down a steep slope at the base of Pūlama pali on the west side of the Pu‘u ‘Ō‘ō flow field in May 2002.

The pāhoehoe-to-ʻaʻā transition has been widely accepted as irreversible (Macdonald, 1953; Peterson and Tilling, 1980; Kilburn, 1981). Specific examples of misinterpreted reversals of this transition (apparent changes from ʻaʻā to pāhoehoe) were given by Macdonald and others (1983, p. 21) and by Kilburn (2000, p. 299). Kilburn (2000) also summarized the main objection by pointing out that broken ʻaʻā surfaces cannot change into unbroken pāhoehoe surfaces.

During the course of the Puʻu ʻŌʻō-Kūpaianaha eruption, however, we have repeatedly observed ʻaʻā flows changing downslope into pāhoehoe flows. We refer to this change as the ʻaʻā-to-pāhoehoe flow transition and suggest that the pāhoehoe-to-ʻaʻā transition is in fact reversible. Lava that produces most pāhoehoe flows more closely approximates Newtonian rheology, whereas lava that produces ʻaʻā flows has either a Bingham rheology, with a significant yield strength, or demonstrates other non-Newtonian rheological behavior (Shaw, 1969; Pinkerton and Sparks, 1978; Moore, 1987; Pinkerton and Norton, 1995).

We suggest that there is a limited rheological interval in which lava with a yield strength may produce either pāhoehoe or ʻaʻā crust, depending on strain rate. ʻAʻā surfaces do not form by simple rupturing of pāhoehoe surfaces but, rather, by the tearing of clinker from the stiff, viscoelastic molten core of an ʻaʻā flow. An individual piece of ʻaʻā crust can certainly not change to a pāhoehoe crust, but neither can a piece of smooth pāhoehoe crust change to ʻaʻā clinker. Kilburn (1990) suggested that the transition threshold between pāhoehoe and ʻaʻā is a failure envelope, where the crust begins to fragment, instead of being a simple rheological boundary. Cashman and others (1999) proposed a more complex model using crustal stability and rheology to define the transition. These models of the transition are based upon changes in the type of crust and are compatible with the original definition of the pāhoehoe-to-ʻaʻā transition.

Examples of ʻAʻā-to-Pāhoehoe Flow Transitions

Since the onset of nearly continuous effusion from Kīlauea Volcano in July 1986, a number of types of transitions from ʻaʻā to pāhoehoe flows have been observed. Lava from the Puʻu ʻŌʻō and Kūpaianaha vents flowed over gently sloping ground (1–5°), then down a much steeper slope (5–20°), and, finally, onto the flat (<2°) coastal plain (Kauahikaua and others, this volume). Most of the transitions from ʻaʻā flows to pāhoehoe flows occur at the change from steeper to flatter slopes. The transition is, in general, reflected in the types of solidified lava exposed in these two areas. A mixture of ʻaʻā and pāhoehoe flows exists on the steeper slope, whereas on the flat coastal plain virtually no ʻaʻā can be seen.

Owing to a combination of circumstances, the transition from ʻaʻā to pāhoehoe crust along a single flow has gone largely unreported. Macdonald (1953) made many of his observations on active eruptions of Mauna Loa, whose high effusion rates and short durations strongly favor ʻaʻā formation. Peterson and Tilling (1980), who made their observations during the prolonged Mauna Ulu eruption, had easy access to the vent area but had difficulty observing the transition from steep to flat slopes near the coast owing to the lack of roads. In contrast, access to the vent areas has been difficult during the ongoing Puʻu ʻŌʻō eruption, whereas access to lava flows on the flat coastal plain has been relatively easy. In addition, transitions from pāhoehoe to ʻaʻā along single Mauna Loa flows tend to be well preserved, but transitions fromʻaʻā to pāhoehoe flows are rapidly overplated by subsequent pāhoehoe flows during long-lived eruptions.

Channelized ʻAʻā Flows Changing to Pāhoehoe Flows

The most convincing examples of the ʻaʻā-to-pāhoehoe flow transition occur when a channelized ʻaʻā flow advances from the steeper slope onto the coastal flat without stagnating. The distribution of solidified levee morphologies is roughly the reverse of that reported by Lipman and Banks (1987). ʻAʻā grades downflow into slabby pāhoehoe, and then into pāhoehoe (fig. 2). On the morning of August 17, 1998, we arrived at the front of an ʻaʻā flow just as it was changing to a slabby pāhoehoe flow on the coastal flats. The progress and speed of this flow was captured on video (Hon and Gansecki, 2002). By midafternoon (about 6 hours later), the flow front had slowed significantly and consisted of smooth pāhoehoe lobes and toes. The ʻaʻā, slabby pāhoehoe, and pāhoehoe portions of this flow were all fed by an open channel that extended from the flow front approximately 200 m upslope to where it was crusted over (fig. 2).

Another excellent example was observed on September 8–9, 2000, in which a similar transition from ʻaʻā flow to pāhoehoe flow occurred over a 24-hour period. When seen in the late afternoon, the flow front was undergoing a transition from stiff, clinkery ʻaʻā to less viscous, spiny ʻaʻā. By midafternoon of the next day, the flow was in transition from ʻaʻā to slabby pāhoehoe as it began to spread out on the coastal flats. By early evening, the front of the flow had completely changed to a channel-fed pāhoehoe flow.

In both of these cases, the lava channels feeding the flow fronts were crusting over several hundred meters behind the active flow front (fig. 2). The formation of tubes insulates the lava delivered to the flow front and keeps the temperature from decreasing very much as flows lengthen (Helz and others, 1993; Helz and others, this volume). The relative position of the interface between crusted channel and open channel remained at a fairly constant distance behind the flow front as the flow underwent the transition from ʻaʻā to pāhoehoe (that is, lava-tube formation was keeping pace with the advance of the flow front). Once the transition to a pāhoehoe flow is completed, the channel commonly crusts over completely within a few hours to a day, and the lava then advances as a tube-fed sheet flow.

A slightly different variation of the ‘a‘ā-to-pāhoehoe flow transition occurs when an ‘a‘ā flow reaches a flat area and the front stops advancing. The rubbly snout of the ‘a‘ā flow confines additional lava coming into the flow and causes the front to inflate (see, for example, Calvari and Pinkerton, 1998). The front then ruptures, and the fluid interior pours out and forms a pāhoehoe flow (fig. 3*A*). The ‘a‘ā flow seen in figure 3*B* was still moving slowly when a lava stream broke out of the flow front and a pāhoehoe lobe spread rapidly in front of the ‘a‘ā flow. This pāhoehoe lobe then inflated and formed a sheet flow.

Surges and Overflows from ‘A‘ā Channels and Levees

More localized examples of the ‘a‘ā-to-pāhoehoe flow transition have been observed along the margins of channels in ‘a‘ā flows. Either channel blockages or surges in lava supply can cause lateral overflows from ‘a‘ā levees (Lipman and Banks, 1987). In both cases, the overflows at a lower velocity form pāhoehoe crusts rather than ‘a‘ā crusts. Overflows generally represent a small volume of the material in the channel and tend to spread laterally rather than downhill,

A

B

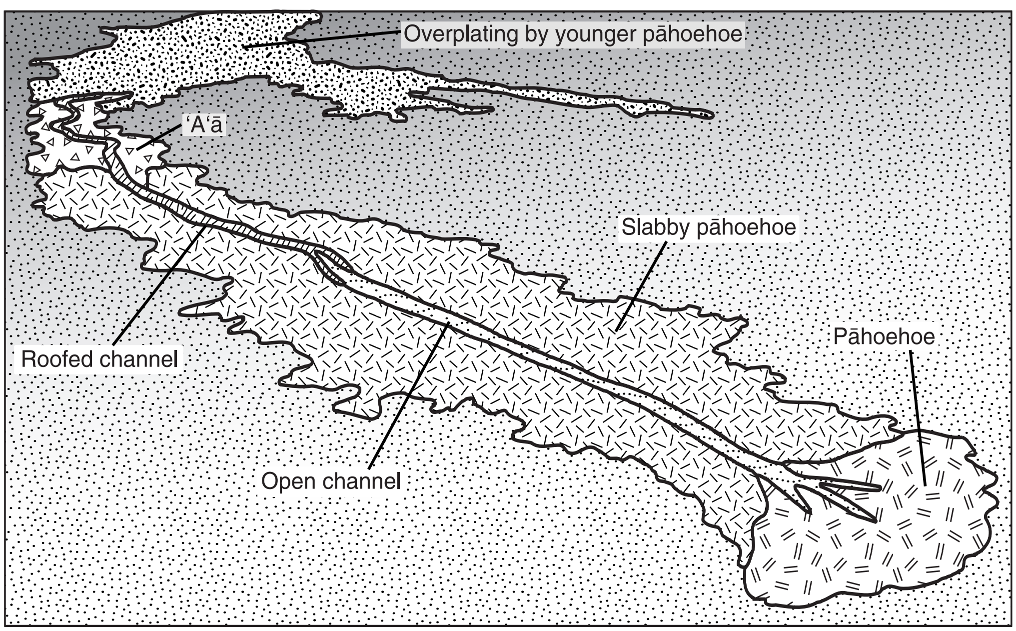

Figure 2. Transition from ‘a‘ā to pāhoehoe along a single flow. *A*, Video image of the August 17, 1998, flow (Hon and Gansecki, 2002), showing the ‘a‘ā-to-slabby-pāhoehoe part of the transition. The flow is about 75 m wide. *B*, Drawing of the entire ‘a‘ā-to-pāhoehoe flow transition on August 17, 1998, showing the pāhoehoe snout of the flow.

reducing both velocity and shear strain on the flow and thereby allowing pāhoehoe to form.

Lipman and Banks (1987) reported pāhoehoe overflows from the channel about 3 km beyond the ʻaʻā transition during the early stages of the 1984 Mauna Loa eruption. Cashman and others (1999) sampled a transitional slabby pāhoehoe flow that broke out about 300 m downstream of the pāhoehoe-to-ʻaʻā transition during the current eruption of Kīlauea. Another excellent example can be seen at Muliwai a Pele in the Mauna Ulu flow field along the Chain of Craters Road. Here, overflows of smooth pāhoehoe deposited pāhoehoe-crusted lava balls on ʻaʻā levees.

Pāhoehoe Leaks from ʻAʻā Flows

Small oozes or leaks of pāhoehoe from dying ʻaʻā fronts have been observed repeatedly during Kīlauea's current eruption and are well documented in other studies (Camp and others, 1987; Wolfe and others, 1988; Jurado-Chichay and Rowland, 1995; Calvari and Pinkerton, 1998). Reduced lava supply to the flow front provides insufficient force to move the front. The small amount of incoming lava pools behind the snout of the ʻaʻā flow and can produce lateral injections (Lipman and Banks, 1987) that "leak" from the levees or the snout. On September 8, 2000, small pāhoehoe flows were observed oozing from the side of an ʻaʻā flow (fig. 4) that had stagnated on flat ground. The lava channel was open for several kilometers upslope, and lava tubes were not present. Wolfe and others (1988) reported similar small pāhoehoe toes oozing from the stagnated fronts of large ʻaʻā flows during the early high fountaining stages of Puʻu ʻŌʻō. Macdonald also noticed pāhoehoe toes leaking from ʻaʻā flows but ascribed them to younger pāhoehoe flows burrowing beneath older ʻaʻā flows, thereby "giving the false appearance of flow changing from ʻaʻā to pāhoehoe" (Macdonald and others, 1983, p. 21).

Jurado-Chichay and Rowland (1995) described a complex example in a 1,300-year-old Mauna Loa flow, where an overflow from an ʻaʻā channel emerged as pāhoehoe and changed to ʻaʻā as it moved away. The ʻaʻā flow appeared to have stopped advancing, inflated, and finally leaked pāhoehoe before stagnating. They interpreted the formation of pāhoehoe at this locality as the result of the crusting-over of earlier flows.

Breakouts from Inflated Pāhoehoe Flows

Large breakouts from the fronts of inflated pāhoehoe flows can show an interesting variation on the ʻaʻā-to-pāhoehoe flow transition. As the flow front breaks open, lava is pushed through a relatively small opening (1–5 m wide) and begins to spread rapidly. Initially, the flow moves very fast, owing to a combination of narrow width and high supply from stored lava (Hon and others, 1994), commonly forming slabby pāhoehoe flows. As the flow spreads out and the lava supply decreases, the velocity of the flow also decreases, and normal pāhoehoe forms at the edge of the flow. We have seen this happen repeatedly on Kīlauea; transitions between slabby pāhoehoe and normal pāhoehoe can also be seen on the inflated McCartys pāhoehoe flow in the Zuni Bandera field of New Mexico (K.A. Hon, unpublished data, 1995).

Changes from ʻAʻā to Pāhoehoe That Are Not Transitions

During some Hawaiian eruptions, a widespread change occurs from ʻaʻā flows to pāhoehoe flows that is related to a drop in effusion rate (Macdonald, 1953; Rowland and Walker, 1990, Kilburn, 2000). The effusion rate may wane during an eruption because of a change in flux or in vent geometry and storage. Rowland and Walker (1990) clearly document these changes for the paired eruptions of Mauna Loa (1859 and 1880–81) and for the Puʻu ʻŌʻō-Kūpaianaha eruption, as well as for a number of other Hawaiian eruptions. High effusion rates are accompanied by relatively high strain rates; cooling rates are also high because tube formation is inhibited in large channels. This change from ʻaʻā to pāhoehoe during the course of an eruption does not represent an example of the ʻaʻā-to-pāhoehoe flow transition. The change generally does not occur along a single flow but, rather, represents a change in style of flow-field development.

Another type of change from ʻaʻā to pāhoehoe may take place when lava tubes form by the crusting-over of channels in ʻaʻā flows (Peterson and Swanson, 1974; Peterson and Tilling, 1980; Calvari and Pinkerton, 1999). As the crust of the tube thickens, it reduces the cross-sectional area of the nascent lava tube. This causes a constriction, which frequently results in breakouts of slow-moving pāhoehoe from the new tube well upslope of the flow front. These pāhoehoe flows, along with other pāhoehoe flows that commonly overplate the earlier ʻaʻā flows, give the appearance of an ʻaʻā-to-pāhoehoe flow transition. The pāhoehoe facies, however, is upslope of the ʻaʻā facies during formation.

Rheological and Crustal Transitions in Hawaiian Lavas

Peterson and Tilling (1980) and Kilburn (1981) altered the discussion about the pāhoehoe-to-ʻaʻā transition from the change in crust morphology along a flow to the change that a discrete "element" of molten lava undergoes as it is transported to the flow front. The pāhoehoe-to-ʻaʻā transition was sketched as the transition threshold zone (TTZ) on a dimensionless graph of apparent viscosity versus shear strain (fig. 9 of Peterson and Tilling, 1980). Within the TTZ, slabby pāhoehoe forms at high strain rates, and viscous spiny pāhoehoe forms at lower strain rates. Peterson and Tilling (1980) gave examples to demonstrate how changes in

A

B

Figure 3. *A*, Pāhoehoe flow breaking out from the front of a stagnated ‘a‘ā flow at the base of Pūlama pali on September 10, 2000. *B*, Pāhoehoe flow breaking out from the front of an ‘a‘ā flow as it reaches the flat just outside Kalapana (photograph taken April 3, 1990). Notice the incandescent material in the ‘a‘ā flow, which was still active as the pāhoehoe flow emerged.

Figure 4. Pāhoehoe lobes oozing from an ‘a‘ā levee as the ‘a‘ā flow stagnates (photograph by A.M. Burt, September 8, 2000). The active ‘a‘ā levee is seen in the picture above the pāhoehoe lobe, which is flowing on older ‘a‘ā crust.

shear strain and apparent viscosity could force lava from the pāhoehoe field into the ‘a‘ā field.

Kilburn (1981) made some simplifying assumptions, based upon the rheological measurements of Shaw (1969) and Pinkerton and Sparks (1978), to show that the range of apparent Newtonian viscosities (η_a) at a given Bingham viscosity (η_b) and yield strength (σ_y) defines a distinctive curve (fig. 5). Bingham fluids differ from Newtonian fluids in having a yield strength (Shaw and others, 1968; Peterson and Tilling, 1980; Kilburn, 1981). The values of η_a for Bingham lavas asymptotically approach η_b (fig. 5) at high shear-strain rates. As the shear-strain rate decreases, η_a increases dramatically. If there is no change in η_b (for example, no change in temperature, volatile, or crystal content), the fluid follows a η_a curve (white arrows in figure 5) as the shear-strain rate is reduced (for example, the flow slows). If the fluid cools, crystallizes, or otherwise increases in apparent viscosity, the path will shift to the right (black arrows) in figure 5. A black vertical dashed line shows the Newtonian viscosity (η_N) for Hawaiian lava with a temperature of 1,150°C (Shaw, 1969), a temperature common for lavas within a few kilometers of the Pu‘u ‘Ō‘ō vent (Helz and others, this volume; Cashman and others, 1999).

Kilburn (1981) proposed that the TTZ is represented by a family of η_a curves for lava with Bingham rheology. According to this model, lava that passes into the TTZ cannot return to a previous state; the transition is irreversible. Kilburn (1990) later modified this argument to suggest that the TTZ represents a crustal-failure envelope rather than a simple rheological boundary. We have followed this model in constructing a semiquantitative graph of the transition for Hawaiian lava (fig. 6*A*).

Refining the Graphical Transition for Hawaiian Lavas

Apparent Newtonian viscosity η_a curves L, 1, 4, and 8 (table 1, fig. 6*A, B*) have been constructed from data for Kīlauea and Mauna Loa lava with Bingham viscosities (Shaw and others, 1968; Shaw, 1969; Moore, 1987), using equation 3 of Kilburn (1981).

$$\eta_a = \eta_b + (\sigma_y / \acute{\varepsilon}),$$

where σ_y is the yield strength of the lava and $\acute{\varepsilon}$ is the shear-strain rate.

Kilburn (1981) had previously plotted the data of Shaw and others (1968) from Hawai‘i and the single curve from an Etna ‘a‘ā flow (Pinkerton and Sparks, 1978) on a similar plot. The η_a curves from the two volcanoes differ significantly in shape, owing to the very different temperatures and compositions of the lava (Kilburn, 1981). Plotting the viscosity

Table 1. Viscosity data for Hawaiian tholeiite lava with Bingham rheology used to construct figures 5 and 6.

Curve number	Bingham viscosity η_b (Pa·s)	Yield strength σ_y (Pa)	Temperature (°C)	Crystal content (percent)	Principal reference
Curve L	700	95	1130–1135	25	Shaw and others (1968)[1]
Curve 8	1100	150	1128; 1139	22; ---	Moore (1987)[2]
Curve 4	3700	930	1126; 1133	40; 24	Moore (1987)[3]
Curve 1	81000	2225	1125	>40?	Moore (1987)[4]

[1] Average of the two measurements taken from Makaopuhi lava lake.

[2] Viscosity data from station 8 at 2,500-m elevation on Mauna Loa (Moore, 1987, p. 1581). Location given on figure 57.1 and referred to as 1852 vent in table 57.1 of Lipman and Banks (1987). Temperatures taken from table 57.1 for two samples from 2,500-m elevat ion (NER-12/57 and unmarked sample). Crystallinity estimated from table 3 of Crisp and others (1994) for sample NER-12/57 taken at 2,250-m elevation.

[3] Viscosity data from station 4 at 1,900-m elevation on Mauna Loa (Moore, 1987, p. 1582). Yield strength is average of two estimates given. Temperatures are from table 57.1 of Lipman and Banks (1987) for samples NER-12/48, taken at 1,940-m elevation, and NER-12/28, taken at 1,730-m elevation near station 4. Crystallinity (40 percent for NER-12/48 and 24 percent for NER-12/28) is taken from table 3 of Crisp and others (1994) for same samples. Note that Crisp and others (1994) may have mislabeled sample NER-12/28 as NER-12/27 throughout their paper. The sample they label as NER-12/28 is a near-vent sample.

[4] Viscosity data from station 1 at 1,600-m elevation on Mauna Loa (Moore, 1987, p. 1582). Yield strength is lower of two estimates given and was used by Moore (1987). Temperature is from table 57.1 of Lipman and Banks (1987) for a measurement taken in 1984 on 3/31. There are no data on crystallinity for this site, which is even farther from the vent than sample NER-12/28 and is probably at least as crystal rich. This lava did not fit a Bingham rheological model and was thought to exhibit pseudoplastic or other complex behavior by Moore (1987).

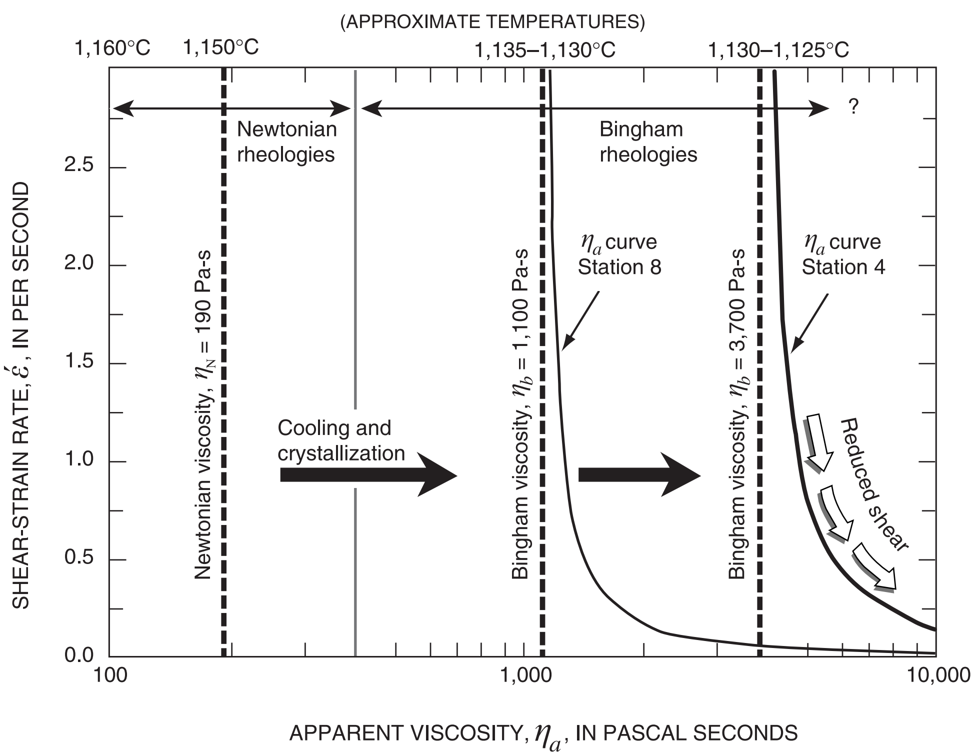

Figure 5. Diagram of shear-strain rate versus apparent Newtonian viscosity of lava, showing difference between a Hawaiian lava with near-Newtonian viscosity (data from Shaw, 1969) and Hawaiian lavas with Bingham viscosities and yield strengths (data in table 1, from Moore, 1987). The development of yield strength may be caused by crystallization of plagioclase microlites (Cashman and others, 1999). The apparent viscosity of the Bingham lava changes with the rate of shear strain (Kilburn, 1981). Both cooling and crystallization increase the Newtonian or Bingham viscosity of the fluid in the direction of the thick black arrows from one vertical Bingham viscosity (η_b) line to another. Decreasing the shear-strain rate causes the apparent Newtonian viscosity (η_a) of Bingham lavas to increase dramatically (white arrows).

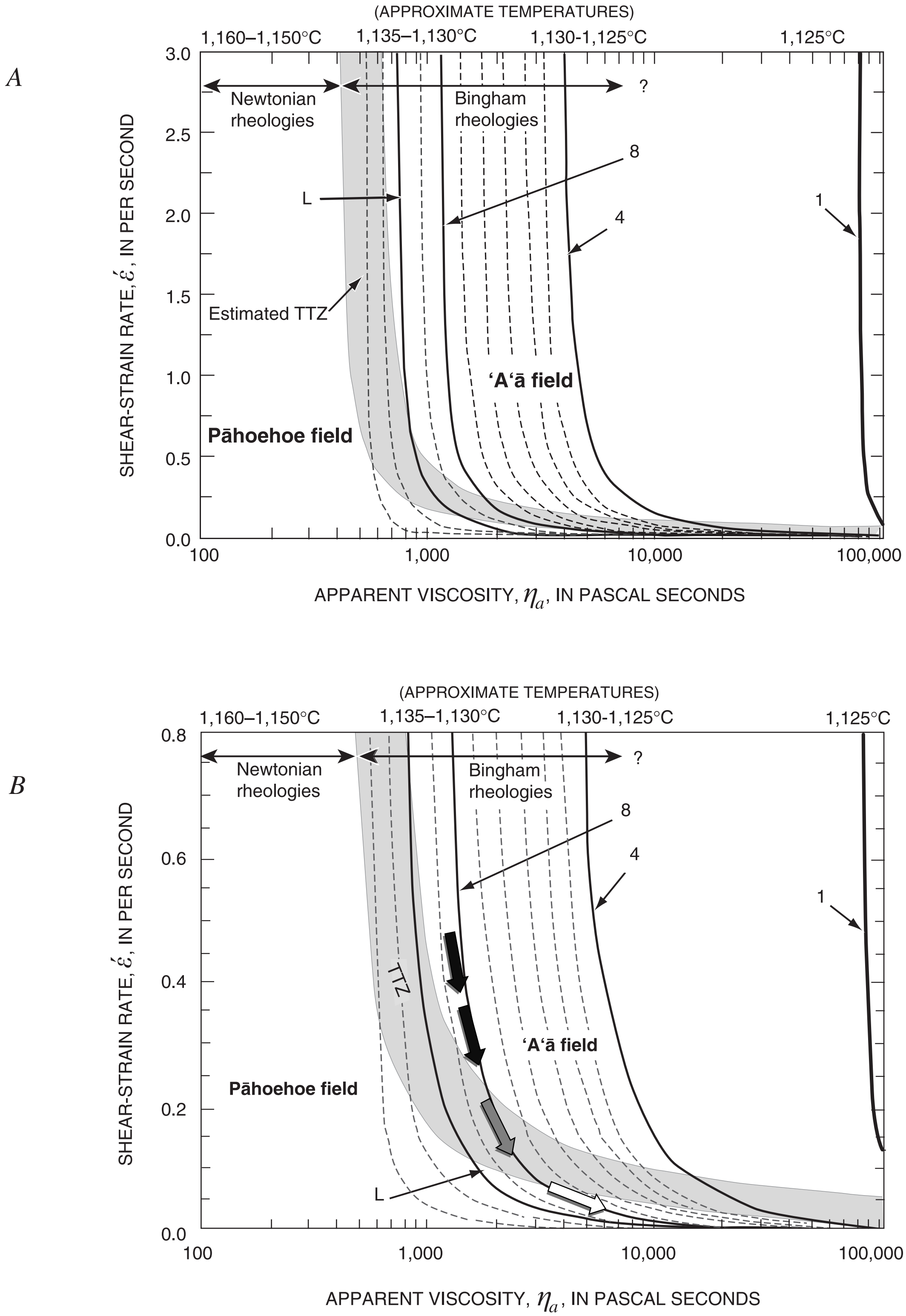

Figure 6. Diagrams of shear-strain rate versus apparent Newtonian viscosity, showing relation of pāhoehoe and 'a'ā lavas. *A*, Approximate stability fields for Hawaiian pāhoehoe and 'a'ā and approximate transition threshold zone (TTZ) within Bingham rheology field. Shaded area represents estimated TTZ. Solid curved lines are apparent Newtonian viscosities (η_a) for specific Bingham viscosities as calculated for Kīlauea (curve L) and for Mauna Loa (curves 8, 4, and 1) (see text and table 1 for explanation). Dashed lines are estimated and represent family of apparent viscosity curves for Hawaiian lavas with Bingham rheology. *B*, Region of figure 6*A* at lower range of shear-strain values (<0.8) showing lower section of TTZ in detail. Possible flow-front conditions during 'a'ā-to-pāhoehoe flow transition are shown by the arrows. The lava begins in the 'a'ā field (black arrows) and approximately follows an apparent viscosity (η_a) curve as shear strain is reduced. Then the lava enters the TTZ (gray arrow) and finally the pāhoehoe field (white arrow).

directly measured (curve L) by Shaw and others (1968) with the empirically estimated viscosities (curves 1, 4, 8) of Moore (1987) allows generation of a set of curves with similar shapes (fig. 6). The Bingham viscosity data (table 1) are estimates of the apparent viscosity of the lava (a mixture of melt, crystals, and bubbles). These two data sets were determined by different methods, but the closeness in shape and position of curves L and 8 for lava with similar temperature and crystallinity (fig. 6*A*) shows that the Bingham viscosity estimates are in reasonable agreement (Moore, 1987, p. 1585).

Viscosity has been plotted on a log scale because it increases roughly exponentially with decreasing temperature (Shaw, 1969). Linear viscosity plots produce a pāhoehoe field that is extremely narrow and de-emphasizes the broad temperature range over which pāhoehoe occurs in Hawai‘i. Approximate temperatures have been placed at the top of figures 6*A* and 6*B* to provide a reference for several specific η_a curves. The dashed lines are interpolated to show the general shape of the η_a curves and possible paths resulting from reduction of shear strain; the spacing of these curves has no temperature significance.

The approximate ranges of Newtonian and Bingham rheologies are shown by arrows at the top of figure 6*A* and 6*B* (also shown on fig. 5). An apparent viscosity of 400 Pa·s was estimated as the boundary between the Newtonian and Bingham viscosity fields. Moore (1987, p. 1,584) estimated that Hawaiian lavas at 1,140°C have viscosities of 300–400 Pa·s. Lava with temperatures between 1,145 and 1,135°C may have crystallinities of 10–20 percent (Crisp and others, 1994; Cashman and others, 1999; Helz and others, this volume). Cashman and others (1999) suggested that lava with crystallinities of 25–30 percent, particularly when consisting of plagioclase microlites, begins to develop significant yield strength.

Cashman and others (1999) concluded that yield strength is necessary to develop ‘a‘ā or transitional crustal morphologies. This idea is supported by the data of Shaw and others (1968) and Moore (1987), who calculated significant yield strengths for fluids with temperatures below 1,135°C. Our field observations of pāhoehoe flows also show that pasty or spiny crusts begin to form at temperatures less than 1,140°C, suggesting the development of yield strength in these lavas. The estimated boundary provides a left-hand limit in figure 6 to the upper extent of the TTZ. The actual boundary may be lower or higher but is probably not less than 300 Pa·s or greater than 600 Pa·s. Changing the estimated position of the Newtonian-Bingham rheology boundary shifts the TTZ slightly but has no significant effect on our argument.

Curve L (Shaw and others, 1968) represents lava subjected to very low shear-strain rates in a lava lake with a pāhoehoe crust. Similar temperatures of 1,130–1,135°C were common for large ‘a‘ā flows during the early Pu‘u ‘Ō‘ō eruption (Neal and others, 1988). Thus, at higher shear-strain rates, Kīlauea lava with the rheology of curve L could produce ‘a‘ā crust. We suggest that curve L must cross the transition threshold zone (TTZ) somewhere in the region shown on figure 6*A*.

Curve 8 is for channelized Mauna Loa lava with ‘a‘ā crust subject to shear-strain rates of 1-3 s^{-1} (station 8 on fig. 58.17 in Moore, 1987). The TTZ must lie to the left of this curve at high shear-strain rates. Lipman and Banks (1987, p. 1,529) observed small pāhoehoe overflows from the channel near station 8 (2,500-m elevation on the 1852 vent) and stated that lava flowing slowly was capable of generating pāhoehoe down to the 2,400-m elevation. These data and observations suggest that curve 8 for Mauna Loa lava must also cross the TTZ. We arbitrarily chose a small value for the shear-strain rate crossover (about 0.1 to 0.2 s^{-1}) to minimize the pāhoehoe field and conditions favorable for the ‘a‘ā-to-pāhoehoe flow transition (fig. 6*B*).

Curve 4 describes Mauna Loa lava subjected to a strain rate of about 1 s^{-1} and producing ‘a‘ā (station 4 on fig. 58.17 in Moore, 1987). Most of the channel overflows in this region were ‘a‘ā, though a slabby pāhoehoe overflow was also reported (Lipman and Banks, 1987). This suggests that curve 4 might enter the TTZ at low strain rates but may not pass through it.

‘A‘ā flows at station 1 (curve 1 in fig. 6*A*; table 1), about 5 km farther from the vent than station 4, have order-of-magnitude higher apparent viscosities but similar temperatures (Moore, 1987; Lipman and Banks, 1987). These flows were not easily modeled as Bingham rheologies (Moore, 1987) and appear to behave pseudoplastically or thixotropically (Shaw, 1969; Moore, 1987; Pinkerton and Norton, 1995). Hawaiian lava with this rheology probably cannot form normal smooth-surfaced pāhoehoe, though more viscous transitional morphologies, such as spiny pāhoehoe (Peterson and Tilling, 1980) and toothpaste lava (Rowland and Walker, 1987), may form at low shear-strain rates.

The semiquantitative graphs (fig. 6*A, B*) of shear-strain rate versus apparent viscosity for Hawaiian lava differ significantly from the earlier dimensionless graphs (Peterson and Tilling, 1980; Kilburn, 1981) by having values on the axes. However, the graphs should not be used to determine absolute numbers, owing to the uncertainties associated with the original viscosity determinations. Nonetheless, estimating these values as a starting point for further work and discussion is a valuable exercise.

The TTZ (fig. 6*A, B*) asymptotically approaches the apparent viscosity axis at low shear-strain rates, rather than being truncated by this axis, as shown by Peterson and Tilling (1980). This allows a larger range of rheologies to produce transitional crustal morphologies (for example, much of the pāhoehoe derived from differentiated Etna lavas would be considered transitional in Hawai‘i). Peterson and Tilling (1980) did not show apparent viscosity curves in relation to their TTZ, and Kilburn (1981) suggested that the TTZ was represented by a family of apparent viscosity curves. Kilburn (1990) modified the TTZ to cut across and truncate apparent viscosity curves on a dimensionless graph. We show the apparent viscosity curves cutting across the TTZ.

Apparent viscosity η_a is plotted on a log scale (fig. 6*A, B*) to expand the vertical part of the TTZ and the relatively narrow viscosity field for Newtonian rheologies. The Newtonian rheologies represent a broad temperature range of 20–30°C that is highly compressed without the log scale. This small viscosity interval, however, represents a relatively large amount of cooling time compared to that represented by the much larger interval of apparent viscosity between curves 1 and 4 (fig. 6*A*), because cooling and crystallization rates for lava in open channels are roughly uniform for Hawaiian lava (Cashman and others, 1999). As lava cools below 1,140°C, the increasing crystallinity causes a nearly exponential increase in viscosity (Shaw and others, 1968) over a relatively short time period. The temperature data in the region below 1,140°C also become somewhat erratic (table 1; fig. 6), because the heat of crystallization offsets cooling as increasing amounts of plagioclase microlites form (Crisp and others, 1994; Helz and others, this volume).

The position of the transition threshold zone (TTZ) of Peterson and Tilling (1980) has been approximated in this study, largely on the basis of empirical observations discussed earlier. The exact position and shape of the TTZ are tentative. We here purposely minimized the conditions under which lava that produces ‘a‘ā crusts at higher shear rates might produce pāhoehoe at lower shear rates. It is here suggested that the TTZ cuts across the family of apparent viscosity curves for Hawaiian lava in the approximate interval of 1,125–1,140°C (fig. 6*A*), which is typical for pāhoehoe on the coastal flat 10–12 km from the vent (Hon and others, 1994; Keszthelyi, 1995; Keszthelyi and Denlinger, 1996). If this is correct, lava can pass back from the ‘a‘ā field into the pāhoehoe field under this restricted set of rheological conditions.

Tracking Changes in Shear-Strain Rate and Viscosity

Figure 6*B* shows a possible path for a Bingham lava to change from producing ‘a‘ā crust to producing pāhoehoe crust by simply reducing the shear-strain rate. The black arrows represent lava producing ‘a‘ā crust at higher strain rates, the gray arrow represents lava within the transition zone producing slabby pāhoehoe crust, and the white arrow represents lava producing pāhoehoe crust. Again, the actual shear-strain rates are speculative. The path assumes that the Bingham viscosity of the fluid producing the ‘a‘ā and the pāhoehoe is nearly constant. This case is relatively easy to demonstrate for pāhoehoe-toe breakouts from stagnated ‘a‘ā fronts. Since the lava that produced the ‘a‘ā and the pāhoehoe must have traveled down the same channel system to arrive at the front, the molten lava that arrives last to produce the pāhoehoe will have cooled more, as a result of waning supply rates and increased storage time in the flow front. This clearly demonstrates that lava feeding these flows was capable of producing either pāhoehoe or ‘a‘ā, depending on the strain rate.

The relationship is more complex for the partially tubed ‘a‘ā and pāhoehoe flows. For these flows, the length of open channel from the tube mouth to the flow front remains nearly the same as the flow advances. Cooling during open-channel flow is faster (0.0005°C/s) and has a much greater effect on crystallization and lava viscosity than slower cooling (0.0001–0.0002°C/s) during flow in a lava tube (Cashman and others, 1999; Helz and others, this volume). Assuming that the time of transport in the open channel remains nearly constant, the only increase in cooling rate would be caused by lengthening of the lava tube as the flow moves downslope. This condition requires that the flow front advances at about the same rate that the lava tube crusts over, in agreement with our observations. Cooling is time-dependent, so the length of open channel can actually be shorter on flat ground where the velocity of the lava drops and the lava cools more per unit distance traveled. Thus, the viscosity of the flows will increase fairly slowly, as shown on figure 6*B*. In contrast, the shear-strain rate will be controlled mainly by the ground slope, which changes rather rapidly from >5° on the slope to <2° on the coastal flat. The shear-strain rate may also be affected by surges in flux rate, caused by constriction of channels or tubes and fluctuations in vent output.

A hidden difficulty exists in attempting to track paths of discrete elements of molten lava described by Peterson and Tilling (1980, p. 286) on graphs such as that in figure 6*B*. Implicit in any discussion of lava travel paths is transport time, which affects cooling rate and crystallization. Since time is not on one of the axes, plotting any type of path on this graph poses a challenge. Changes in temperature along a flow could be transformed into length of time, using the approximate cooling rates given by Cashman and others (1999), and further transformed into distances if transport velocities along the flow are known. The release of heat during crystallization also adds to the complexity of estimating rates of cooling, and some flows actually appear nearly isothermal while increasing greatly in apparent viscosity (Crisp and others, 1994). As can be seen on figure 6*A* and 6*B*, any attempt to transfer cooling rates into time will produce a highly nonlinear scale.

Another problem with the idea of a path is that the conditions to which lava is subjected when flowing in a channel are not analogous to conditions the same lava will undergo upon reaching the flow front, where most crust forms. Two different, but related, processes are represented on the graphs (fig. 6): one is the change in rheology of the fluid, and the second is the formation of crust. Crust clearly plays an integral part in the emplacement of any lava flow and has a significant effect on the bulk viscosity of the flow (Kilburn, 1993, 2000; Hon and others, 1994; Keszthelyi and Denlinger, 1996; Cashman and others, 1999). Most crust on lava flows begins forming at the flow front, not during transport in a channel or tube system, because lava at the front spreads over a wide area and is exposed to higher shear-strain rates and faster cooling. In contrast, most lava moving from the vent to the flow front generally does not undergo conditions conducive to crust production; otherwise lava in the channel would solidify.

Instead, graphs like those in figure 6*A* and 6*B* are, perhaps, better used to visualize shear-strain rate and viscosity conditions to which lava is exposed at a certain point in time. We suggest that tracking the conditions at the flow front as it moves downslope is as important as tracking the rheological changes in the lava as it is transported to the flow front. The changes in lava rheology would be tracked on the same graph, but without the TTZ. Once lava reaches the area where crust is being produced, the TTZ is useful in predicting which type of crust will form.

Migration of the Pāhoehoe-to-ʻAʻā Transition

Both Lipman and Banks (1987) and Wolfe and others (1988) found that the position of the pāhoehoe-to-ʻaʻā flow transition on large, open-channel flows remained very close to the initial transition (marked by the change from pāhoehoe to ʻaʻā crust on the levees) throughout the life of the flow. In contrast, the position of the pāhoehoe-to-ʻaʻā flow transition on smaller, partially tubed flows can migrate downstream with respect to where the change occurred on the solidified levees. The migrating transition may be marked along the channel by coatings of smooth pāhoehoe crust on the surrounding ʻaʻā levees.

The migration of the transition along the channel is commonly due to the formation of lava tubes (Peterson and Tilling, 1980, p. 276), which can form in a few hours or days on small channels (1–5 m) but can take weeks or months to form on larger channels. In Hawaiʻi, large eruptions with high effusion rates last only a few days or weeks and generally end before lava tubes can form on their wide channels (10–20 m). An exception is the Kaʻūpūlehu flow on Hualālai Volcano, where the channels were eroded by a bedload of dunite xenoliths. Those channels have a narrow and deep cross section that permitted rapid tube formation (Kauahikaua and others, 2002) and allowed the pāhoehoe-to-ʻaʻā transition to migrate downslope, as indicated by pāhoehoe coatings lining the channel walls. The old, high-effusion-rate flows on Mauna Loa studied by Jurado-Chichay and Rowland (1995) also partially tubed over and allowed the transition zone to migrate to the coastal region.

If the boundary between the crusted channel and the open channel begins to migrate faster downslope than the flow front, the entire channel will eventually be tubed over to the flow front and the flow will then become pāhoehoe. This is the ultimate fate of all the relatively small channelized ʻaʻā and pāhoehoe lava flows that reach the coastal flats. The critical observation is whether the channel crusts over before or after a channelized ʻaʻā flow changes to pāhoehoe. Some flows reach the flat with channel systems that stay open until after the flow has undergone a change from ʻaʻā to pāhoehoe. Other ʻaʻā flows may actually become completely tubed over as they slow down and stop on the flatter ground. This insulates lava delivered to the flow front and produces outbreaks of pāhoehoe from the ʻaʻā front. The rheology of the lava would appear to shift to the left if conditions at the flow front were tracked on a graph like that of fig. 6*A*. Clearly no individual parcel of molten lava can get hotter upon arrival at the flow front.

Do Pāhoehoe and ʻAʻā Liquids Exist?

The terms pāhoehoe and ʻaʻā were expanded by Peterson and Tilling (1980) to include both fluid and crust, in a way similar to the use of the term lava for both molten and solid rock. We suggest, however, that the use of pāhoehoe and ʻaʻā be restricted to the solidified lava surfaces or lava flows dominantly covered by crust, as Macdonald (1953) suggested. Solidified pāhoehoe crust does not change to ʻaʻā crust, nor vice versa. The nature of the molten lava can only be determined by the crust it becomes and not reliably by any morphology of the molten lava itself. Furthermore, the nature of the molten lava beneath the crust cannot always be inferred from the morphology of the crust, particularly in channels.

A number of studies (Kilburn and Guest, 1993; Jurado-Chichay and Rowland, 1995; Cashman and others, 1999) demonstrated that the fluid cores of ʻaʻā flow channels and lobes have a complex thermal stratification. In contrast, the interiors of pāhoehoe flow channels and flow lobes appear to be relatively well mixed (Hon and others, 1994; Cashman and others, 1999). A flow that is undergoing the pāhoehoe-to-ʻaʻā transition defined by Macdonald (1953) would have a fluid core whose outer surfaces are cooling and producing solidified products that change progressively from pāhoehoe to ʻaʻā crusts along the length of the flow. The molten interior of a lava channel can easily be masked by a cooler outer layer that is producing crust (Cashman and others, 1999). Therefore, estimating the rheological properties of lava within a given channel or lobe is not as simple as it appears. If our analysis of the TTZ is correct, some molten lava may also have the ability to form either pāhoehoe or ʻaʻā crusts, depending solely on the strain rate.

The argument in this paper differs little from the original premise of Peterson and Tilling (1980), who demonstrated that it is the shear-strain rate acting on molten lava with a given rheology that determines the type of crust produced. Kilburn (1990) amplified this concept by suggesting that the TTZ is a failure envelope controlled by shear-strain rates that cuts across rheological boundaries. The clarification being made here is that the transition is defined by the type of crust formed and not by the rheology of the molten lava producing the crust.

Discussion

Are the examples given above truly examples of an ʻaʻā-to-pāhoehoe flow transition? Using the original definition given by Macdonald (1953), we believe the answer is "yes" for our first type, in which ʻaʻā flows change sequentially into pāhoehoe flows. This change occurs at the flow front, where the flow displays a morphological sequence of ʻaʻā to slabby pāhoehoe to pāhoehoe on the levees as the flow advances, the exact reverse of the widely accepted pāhoehoe-to-ʻaʻā transition.

The ‘a‘ā-to-pāhoehoe flow transition is most commonly associated with a drop in shear stress while the molten lava delivered to the flow front remains at a relatively constant temperature. These conditions can occur for channelized flows advancing from steep to flat ground and for overflows from ‘a‘ā flow channels. In such cases, the conditions of apparent viscosity and shear-strain rate stay fairly close to the region between curves L and 8 (fig. 6*A*), if our approximation of the transition is correct.

Kilburn (2000) discounted pāhoehoe outbreaks from ‘a‘ā fronts because they incorporate “hotter” lava from the core of the flow. However, the molten lava core that produces the pāhoehoe in this low-shear-strain environment has traveled past the point of pāhoehoe-‘a‘ā transition, as marked by solidified ‘a‘ā on the levees of the stagnated flow. Clearly lava that produced an ‘a‘ā fan and levees on the steeper slope also produced pāhoehoe upon reaching the flatter ground. The lava that produces pāhoehoe-crusted flows cannot be the same element of lava that earlier produced ‘a‘ā crust. Both elements of lava arrived at the front through the same transport system and should be rheologically identical. The only difference is that the later lava experienced lower shear-strain rates on the flatter ground.

Cashman and others (1999) suggested that the molten lava producing transitional pāhoehoe breakouts comes from the hotter inner component (about 1,135°C and 15–20 percent crystals) of the ‘a‘ā channel, whereas the cool outer component (~1,100°C and 40 percent crystals) appears to have undergone an irreversible transition and is capable of producing only ‘a‘ā crust. They demonstrated that the increasing number of plagioclase microlites below 1,140°C raises the yield strength and is largely responsible for the change in rheology of Hawaiian lava.

Similar data from the 1984 Mauna Loa eruption suggest that molten lava capable of producing either ‘a‘ā flow fronts or pāhoehoe overflows has crystallinities of 20–30 percent at temperatures of 1,125–1,140°C (Lipman and Banks, 1987; Crisp and others, 1994). Temperatures of large ‘a‘ā flow fronts during the early Pu‘u ‘Ō‘ō stages were commonly about 1,100°C, but large surges from the flow interior had temperatures of about 1,135–1,138°C (Neal and others, 1988). The surges were fluid, spiny ‘a‘ā, but small leaks of pāhoehoe from stagnant flows indicate that the same fluid core could produce pāhoehoe at low strain rates (Wolfe and others, 1988).

All of these data suggest that Hawaiian lava in the approximate temperature range of 1,125–1,140°C can produce either ‘a‘ā or pāhoehoe crust, depending on the rate of shear strain (fig. 6*A*). Along the course of a channel, the fluid can become pāhoehoe crust in one place, continue over a steep slope and become ‘a‘ā clinker, then proceed out onto a flat plain and again form pāhoehoe crust.

When flowing in channels, Hawaiian lava with temperatures between 1,125 and 1,140°C can also have a smooth surface appearance; thus, determining in the field whether pāhoehoe or ‘a‘ā crust will form may be difficult. Retaining the original use of the terms pāhoehoe and ‘a‘ā for the solidified crust and not for the molten lava is suggested. Where sufficient patches of crust have formed on the channel surface, the lava should be called “lava with pāhoehoe crust” or “lava with ‘a‘ā crust,” rather than pāhoehoe lava or ‘a‘ā lava. Difficulties can arise for lava flows that contain both fragments of pāhoehoe crust and ‘a‘ā clinker. In some of these flows, pāhoehoe crust may have formed in the channel and rafted some distance to the front. In transitional slabby pāhoehoe flows, both spiny ‘a‘ā crust and pāhoehoe slabs may form simultaneously in response to variable rates of shear strain in the moving front.

Graphs, such as those in figure 6, can be used to track changes in the rheology of lava transported to the flow front. The transition threshold zone (TTZ) can be used separately to evaluate crustal formation at the flow front on the same graph. The type of crust being formed at the flow front can be identified unambiguously in the field, as cooled crustal fragments of either pāhoehoe or ‘a‘ā will generally be present. The graph can be used to predict the type of lava flow that might form when a flow front reaches a specific position, given estimations of lava rheology and slope.

Conclusions

Using available data, we constructed a semiquantitative graph (fig. 6) of shear strain versus apparent viscosity for Hawaiian lava. The approximate shape and position of the transition threshold zone (TTZ), estimated from field data and observations, differs from those of previously published estimates (Peterson and Tilling, 1980, Kilburn, 1981). This new interpretation allows an ‘a‘ā-to-pāhoehoe flow transition to exist in a limited rheological interval (about 1,125-1,140°C) for Hawaiian tholeiitic lava.

Lava streams in ‘a‘ā and transitional flows probably consist of two components (Cashman and others; 1999; Kilburn and Guest, 1993). This stratification allows molten lava still capable of solidifying to pāhoehoe crust to travel in the core of a lava stream, while the surface and outer edges (the cooler and higher shear-strain regions) are producing ‘a‘ā clinker. The rapid formation of lava tubes also plays a crucial role in the ‘a‘ā-to-pāhoehoe flow transition by keeping the Bingham viscosity and yield strength nearly constant as the shear strain is reduced. This allows the transition zone to migrate downslope, rather than remain in a relatively fixed position as it does in open-channel flows (Lipman and Banks, 1987; Wolfe and others, 1988).

Most authors have stressed the role that high effusion rate plays in creating ‘a‘ā fields, while low effusion rates promote the growth of pāhoehoe fields (Macdonald, 1953; Rowland and Walker, 1990; Cashman and others, 1999). ‘A‘ā can, however, be a significant component of low-effusion-rate pāhoehoe flow fields. Unlike the simple pāhoehoe-to-‘a‘ā transitions observed on flows with high effusion rates, the nature of the transition on flows with low effusion rates is nearly impossible to decipher after the eruption has ceased, owing to the progressive overplating by pāhoehoe.

The reversible transition between pāhoehoe and ʻaʻā flow types is strongly controlled by slope and local lava supply, rather than by effusion rate.

Additional studies are needed to understand the flow behavior described here. A simple cataloging of temperatures and crystallinities of pāhoehoe found on the coastal region far from the vent, to compare with ʻaʻā crystallinities and temperatures, would be a significant start. A determined attempt to study and sample these ephemeral lava flows in the process of changing from ʻaʻā to pāhoehoe, much like the studies done on the pāhoehoe-to-ʻaʻā transition (Cashman and others, 1999; Polacci and others, 1999), is also needed. Perhaps this current eruption will provide another 20 years of opportunity to answer these questions.

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