

Hawaiian Lava-Flow Dynamics During the Pu‘u ‘Ō‘ō-Kūpaianaha Eruption: A Tale of Two Decades

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Abstract

Two decades of the nearly continuous Pu‘u ‘Ō‘ō-Kūpaianaha eruption have provided many opportunities to study lava-flow dynamics. Many channelized ‘a‘ā flows evolve to form lava tubes that are covered by pāhoehoe lava. Their initial advance rate appears to be a crude function of effusion rate. Pāhoehoe flows have been more common than ‘a‘ā flows during this eruption, dominantly emplaced by inflation on low slopes. Flows with morphologies transitional between ‘a‘ā and pāhoehoe are interpreted as indicators of flow-field conditions. Observation and analysis of both ‘a‘ā and pāhoehoe flows reveal that a substantial increase in microcrystallinity generally results in the liquid solidifying as ‘a‘ā rather than as pāhoehoe. Lava tubes can form in both ‘a‘ā and pāhoehoe flows but are more common in pāhoehoe. A lava stream flowing within a tube has been documented to downcut through its base at a rate of 10 cm per day for a period of several months. Lava features, such as hornitos, rootless shields, and shatter rings, apparently form over tubes carrying an unsteady lava supply. Lava flows entering the ocean develop a unique set of features and behaviors. Many thermal-characterization studies have been done for active lava flows to calibrate satellite-borne sensors. Promising applications include thermal lava-flux monitoring and lava-flow and lava-tube mapping. The ultimate goal of much of this research is improvement of lava-flow hazard assessments and mitigation tools.

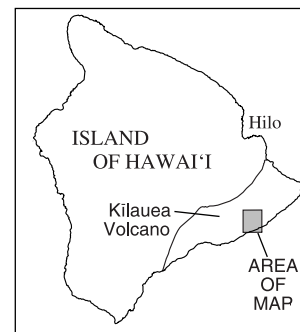
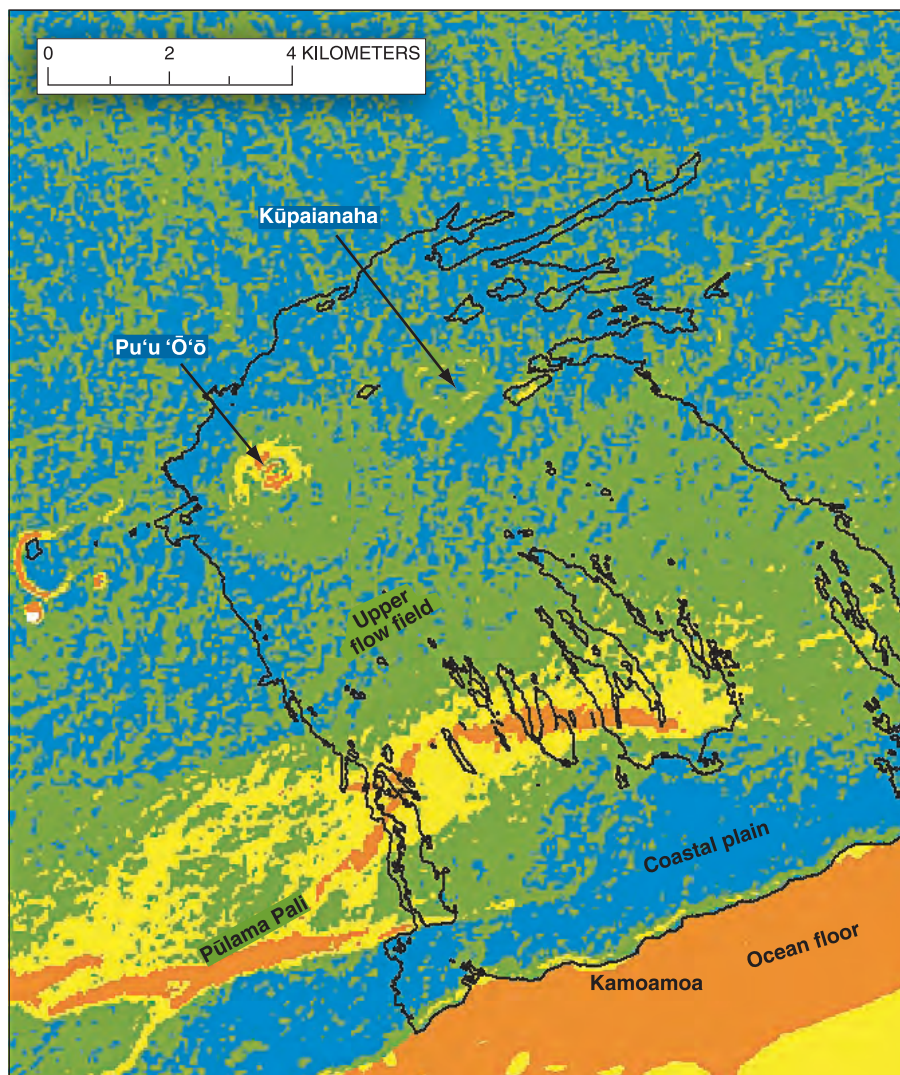
Introduction

In the past 2 decades, great advances have been made in our understanding of the physical processes that control basalt-flow emplacement, resulting in improvement of our tools for the mitigation of lava-flow hazards. Earlier studies of the 1969–74 Mauna Ulu eruption of Kīlauea Volcano and the 1984 Mauna Loa eruption provided a strong foundation for these advances. Although this chapter emphasizes the current eruption, we have incorporated data from these and other Hawaiian eruptions in our interpretations. Some of the lessons learned during previous eruptions have been relearned and expanded upon by a new generation of volcanologists.

The past 2 decades have also been unprecedented in modern times for the continuity of eruptive activity. This continuity has made possible both monitoring and experimenting on numerous aspects of lava-flow emplacement, instead of deducing processes from solidified products. Continuous eruptive activity has also allowed repeated experiments and observations on aspects of pāhoehoe-flow emplacement, such as lava flux, flow inflation, changes in bubble content, and the temperature of basaltic lava during transport through lava tubes. These studies, in turn, have provided new quantitative interpretations of the deposits from older eruptions.

The Pu‘u ‘Ō‘ō-Kūpaianaha eruption has produced lava-flow morphologies from ‘a‘ā to pāhoehoe and all the transitional forms between. The relative abundance of morphologic types, however, varied in both time and space. The first 3½ years of eruptive activity were dominated by fountain-fed ‘a‘ā, and the next 16½ years by tube-fed pāhoehoe. All eruptive activity played out on terrain that can be divided into five areas (fig. 1). The first area is the vicinity of the vent, which, for three of the first 3½ years and for the past 11 years, has been Pu‘u ‘Ō‘ō (see Heliker and Mattox, this volume). The second area, the upper flow field, encompasses terrain between the vent area and the top of Pūlama pali (“pali” is a Hawaiian word for escarpment or steep slope). Here, slopes are typically 1°–5°, and stable, long-lived lava tubes dominate the flow activity during periods of steady effusion; less commonly, lava shield and hornito formation dominates during varying or declining effusion. The third area is the face and base of the pali, where slopes are as steep as 20°. On these slopes, surface flows commonly change to ‘a‘ā, only to be resurfaced by pāhoehoe breakouts from established lava tubes. The fourth area is the coastal plain below the pali, with slopes less than 2°. This area is characterized by a prevalence of lava-flow-inflation structures and other features unique to “filled” lava tubes. The fifth area is the coast itself, a narrow (200–300 m) zone of >2° slopes bounded by low seacliffs and steep offshore bathymetry. This area is host to a range of activity related to the physical conditions of ocean entry as emplacement changes from subaerial to submarine.

Together, observations made over space and time have allowed us to address old questions, such as the process of lava-tube formation and the change from pāhoehoe to ‘a‘ā;



EXPLANATION

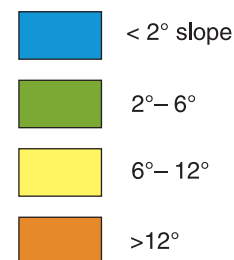


Figure 1. Slope map of Kilauea Volcano, Island of Hawai'i, showing lava-flow field of Pu'u 'Ō'ō-Kūpaianaha eruption, based on a digital elevation model (DEM) derived from February 2000 Shuttle Radar Topography Mission. Offshore bathymetry from Chadwick and others (1993).

develop new models for lava-flow-emplacment behavior, including flow inflation and the origin of shatter rings; apply new technologies, particularly in the realm of remote sensing; and improve the basis for hazard assessment and mitigation.

Channelized 'A'ā Flows

Channelized 'a'ā flows formed during each of the fountaining episodes (1–47) and thus constituted the primary emplacement style of the first 3½ years of the eruption. Channelized 'a'ā flows also formed during the subsequent 16½ years under the following conditions: (1) unusually high effusion rates (for example, the Feb. 1, 1996, surge event), (2) unsteady fluxes after an eruptive pause (for example, in May 1997), or (3) surface flows that reach the steep slopes of the Pūlama pali. Thus, conditions of 'a'ā-flow generation generally matched those summarized by Macdonald (1953), Peterson and Tilling (1980), and Rowland and Walker (1990), who noted that 'a'ā formation requires high strain rates, as well as increases in apparent viscosity.

Evolution of 'A'ā Channels

A comprehensive summary of the flow behavior during episodes 1 through 20 (Wolfe and others, 1988) provides a description of these fountaining episodes. Detailed observations of channelized 'a'ā flows during the 1984 Mauna Loa eruption—which was concurrent with episode 17—provide additional insight into the general characteristics of 'a'ā flow behavior (Lipman and Banks, 1987).

Lava flows from episodes 1 through 3, the last part of episode 35, and the beginning of episode 48 were fed from fissures that erupted extensive near-vent shelly pāhoehoe, some of which, in turn, fed 'a'ā flows. All the other eruptive episodes earlier than episode 48 involved channelized 'a'ā flows fed from a central vent. During episodes 2 through 19 and 42 through 47 (Wolfe and others, 1988; see Heliker and Mattox, this volume), pāhoehoe flows spilled from a central vent and then were rapidly directed into a narrow open channel, 2 to 5 m deep and 5 to 25 m wide, that typically was rectangular in cross

section. Channel levees were constructed by lateral displacement of ‘a‘ā rubble at the flow front and then strengthened over time with repeated coatings by pāhoehoe overflows. During episodes 20 through 41, substantially higher fountains apparently degassed and cooled the lava to the point where the flows started as ‘a‘ā (see Heliker and others, this volume).

During episodes 1 through 19, channel formation permitted transport of fluid lava to distances of 2 to 5 km from the Pu‘u ‘Ō‘ō vent. At this distance, the channel surface gradually became increasingly lumpy with incipient clinker. Once established, the position of this transition in surface morphology generally did not change over time. Farther downflow, the stable channel gradually transformed into a zone of dispersed flow at the front (Lipman and Banks, 1987). Fluid velocities in the channel were 10 to 15 m/s within a few tens of meters from the vent but decreased to 1 to 3 m/s 1 km from the vent. Velocities slowed through the transition zone, and the flow fronts advanced at velocities of <0.1 m/s (Lipman and Banks, 1987).

Superb observations of the channelized ‘a‘ā flow from the 1984 Mauna Loa eruption included detailed measurements of flow advance and changes in channel flow, lava density, and lava temperature over time and distance from the vent (Lockwood and others, 1985; Lipman and Banks, 1987). These studies, which provided evidence for extensive syneruptive crystallization in response to magma degassing, formed the basis of subsequent examination of the rates of syneruptive and posteruptive crystallization (Crisp and others, 1994) and cooling (Crisp and Baloga, 1994). Together, these studies provide the most complete data set and analysis of a Hawaiian ‘a‘ā flow.

Longer lived channelized ‘a‘ā flows emplaced during episodes 48 and later progressed to the final stage of the channelization process. With sustained flow, the upper parts of the channel formed stable pāhoehoe crusts that initiated lava-tube development. The insulation provided by the tube roof permitted hot lava from the vent to progress farther downstream and to subsequently bury ‘a‘ā levees with pāhoehoe. Given sufficient time, the entire flow would be overrun by pāhoehoe. This process illustrates the importance of thermal efficiency in the transport of pāhoehoe-producing lava flows, as discussed in detail below.

Constraints on Initial Flow-Advance Rates

Two lava-flow parameters of critical importance to hazard assessment are rate of advance and ultimate flow length. Episodic high fountaining from Pu‘u ‘Ō‘ō from 1983 to 1986 allowed us to acquire many data sets of advance rates for ‘a‘ā flows (Wolfe and others, 1988; Heliker and others, 2001). Flow length may increase either linearly over time or rapidly at first and then at a slower rate as the flow approaches a final length (fig. 2). Kilburn (1996) described these patterns of flow lengthening in other basaltic-lava flows; he found linear advance to be limited by lava supply, and slowing advance by cooling at the flow front.

To expand our analysis, we have compiled (fig. 2) lava-flow-advance rates for other recent eruptions of Kīlauea (Moore and others, 1980) and Mauna Loa (Finch and Macdonald, 1953; Lipman and Banks, 1987; Barnard, 1990–92). The rapid advance of most recent flows from Mauna Loa’s southwest rift zone precluded detailed estimates of flow advance; instead, advance rates had to be estimated from reports of flows crossing known roads or reaching the ocean. Together, these data show a crude correlation between initial rates of lava-flow advance and volumetric effusion rate (Q). We concentrate here on just the first several hours of each advance, not the entire advance history of each flow. Average initial flow-advance rates for $Q \approx 25$ m³/s are <0.02 m/s, and for $Q \approx 50$ – 70 m³/s ~ 0.08 m/s. In contrast, flows erupted at moderately high effusion rates ($Q \approx 300$ m³/s) can advance as fast as 0.3 m/s over the first several hours, whereas the extraordinarily high effusion rates ($Q \approx 1,000$ m³/s) of the 1950 Mauna Loa eruption produced flows that advanced 3 to 4 m/s during the first day of emplacement.

The observed correlation between flow-advance rate and volumetric effusion rate for flows that traverse distinct topographic settings suggests that slope is of secondary importance in controlling flow-advance rate. Underlying slope has previously been shown to have no statistically evident control on flow width (Peitersen and Crown, 1999, 2000). Theoretical and empirical expressions for flow length as a function of time (advance rate) show a dependence on the sine of the average ground slope (and, thus, of order <1). Instead, flow advance depends on the square root of the effusion rate (Kilburn and others, 1995). Additionally, the final length of an ‘a‘ā flow is best predicted by its effusion rate (Walker, 1973; Pieri, 1986; Pinkerton and Wilson, 1994). These observations point to the importance of accurately determining volumetric effusion rate during the early stages of an ‘a‘ā-producing eruption.

Rheology of ‘A‘ā Flows

‘A‘ā flows have long been recognized to have non-Newtonian rheologies that control the mechanics of flow emplacement (Robson, 1967; Walker, 1967; Hulme, 1974). ‘A‘ā is commonly modeled as a Bingham fluid, which has an apparent yield strength when stress/strain-rate curves measured at high strain rates are extrapolated back to zero shear strain. For a Bingham fluid, flow through a channel is plug-like, and levees form when marginal flow ceases because of yield-strength limitations. The final levee width and channel width have been used to estimate the yield strength (Hulme, 1974). A potential problem with this technique is the observed formation of levees by lateral displacement of solid material at the flow front, rather than by lateral flow limited by lava yield strength.

Data from selected episodes 1–20 ‘a‘ā flows, and from the “1 flow” of the 1984 Mauna Loa eruption, provide an opportunity to estimate rheologic changes in flowing lava as a function of emplacement style and transport distance

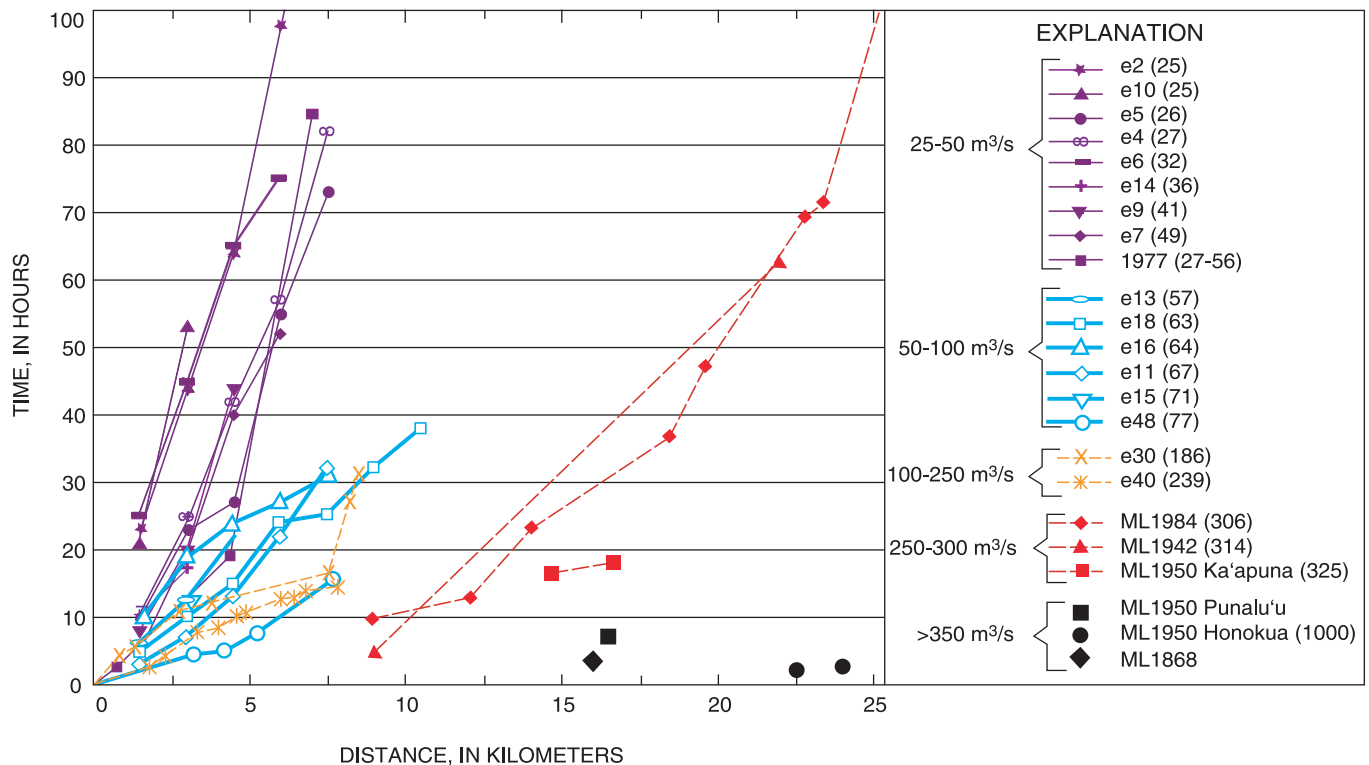


Figure 2. Advance of Hawaiian ‘a’ā flows grouped by range in effusion rate. Rate for each flow given in parentheses; e, Pu‘u ‘Ō‘ō eruptive episode; ML, Mauna Loa.

(Fink and Zimbelman, 1986; Lipman and Banks, 1987; Moore, 1987; Wolfe and others, 1988). Fink and Zimbelman (1986) used measured channel and levee geometries to estimate a 30-fold increase in yield strength and a 10-fold increase in viscosity near the distal end of the 7-km-long, episode 5 ‘a’ā flow accompanying a 30°C decrease in temperature along the lowermost 1.7 km. They concluded that the increase in viscosity is explainable by increasing lava crystallinity but that the increase in yield strength most likely reflects both greater crystallinity and more flow-surface brecciation. Fink and Zimbelman (1990) extended their analysis to episodes 2–5 flows, computing yield strengths of 300 to 40,000 Pa and viscosities of 60 to 0.3×10^6 Pa-s, increasing exponentially along the flow lengths. In a similar study of the 1984 Mauna Loa flow, Moore (1987) deduced a 10,000-fold increase in viscosity and 30-fold increase in yield strength along the 25-km length of the flow.

More recently, Baloga and others (1998; 2001) relaxed assumptions of constant flow density and thickness. They used flow width, thickness, along-flow distance, and elevation measured from flow maps for episodes 2 and 18 (Wolfe and others, 1988) as random variables, from which they estimated “statistics” of viscosity variation based on three different rheologic models. They preferred a model that accounts for the flow volume lost to stagnant, stationary components of the flow. The resulting derived viscosity for episode 2 increased by a factor of ~150 along the flow, similar to the viscosity increase calculated by Fink and Zimbelman (1990) for the same flow. Reanalysis of the data for the “1 flow” from Mauna Loa in 1984 (Baloga and others, 2001) shows the importance of

incorporating density into viscosity calculations and the spurious findings (Baloga and Pieri, 1986) that can result when flow thicknesses are not corrected for density.

Sakimoto and Gregg (2001) used numerical models to simulate observed channel flow during the 1984 Mauna Loa eruption and episode 51 of the Pu‘u ‘Ō‘ō–Kūpaianaha eruption. Using validating data from numerous polyethylene glycol (wax) analog simulations of channel flow, they found a Newtonian fluid-channel model to be more accurate than either Bingham or Newtonian infinite-sheet models for these flows (for example, Tallarico and Dragoni, 2000).

Pāhoehoe Flows

Pāhoehoe has been the dominant flow type throughout most of the Pu‘u ‘Ō‘ō–Kūpaianaha eruption since mid-1986. During the shorter (1969–74) Mauna Ulu eruption, Swanson and others (1979) and Tilling and others (1987) made excellent observations of pāhoehoe flow and lava-tube formation. These studies, along with others (Greeley, 1971, 1972, 1987; Swanson, 1973; Swanson and Fabbri, 1973; Peterson and Swanson, 1974; Peterson and others, 1994) provided a foundation for future work. Here, we focus on five aspects of pāhoehoe-flow-field development: emplacement mechanisms, the evolution of lava tubes, the origin of different morphologic flow types, and complex interactions of lava and seawater at ocean entries.

Inflation Is the Dominant Emplacement Mechanism of Pāhoehoe on Low Slopes

Close observations of newly emplaced pāhoehoe lava on the coastal plain during episode 48 (Mattox and others, 1993) led to an experimental study of inflating pāhoehoe sheetflows (Hon and others, 1993a, 1994) that complemented earlier deductive fieldwork (Holcomb, 1987; Walker, 1991; Chitwood, 1993, 1994) on flow inflation. The process is straightforward—continual growth of upper and lower crust that surrounds a molten interior continually replenished with fresh lava. The interior is a conduit for molten lava and so acts as a lava tube feeding the advancing flow front. An inflating pāhoehoe flow can thicken from tens of centimeters to meters within a few days as lava influx under hydraulic pressure forces the crust upward and outward. Inflation creates hummocky flows or sheetflows, differing only in scale. In hummocky flows, inflation is localized to form tumuli, from meters to tens of meters wide (Swanson, 1973; Hon and others, 1994). Wholesale inflation of sheetflows occurs over hundreds of meters, commonly leaving broad, flat areas (plateaus) elevated above the surrounding terrain.

The upper crust of inflated flows thickens during flow emplacement and, in solidified inflated flows, is recognizable by its vesicularity, which contrasts with a dense flow interior. Where crustal growth rates are known, the thickness of the upper vesicular crust provides an estimate of the duration of flow emplacement, as demonstrated by a comparison of known and calculated flow durations for specific lobes of the 1990–91 Kalapana flow (Cashman and Kauahikaua, 1997). Additionally, vesicle distribution within the upper crust confirms that the flow interiors are overpressured (Cashman and Kauahikaua, 1997). The source is hydrostatic and is transmitted through the fluid-filled parts of lava tubes or flow interiors. Overpressuring explains several features of lava breakouts from tumuli, including squeezeups through cracks, breakouts that push apart tumuli from within, and the “blue glassy” breakouts that issue from the base of tumuli as dense, vesicle-poor lava which had some of its original gas resorbed in response to the overpressure (Hon and others, 1994).

Extension of the upper crust over the axis of inflation creates a downward-propagating axial crack. Banding caused by alternating brittle- and ductile-failure patterns on the interior crack walls has been interpreted either as pulsed crack formation (Hon and others, 1994), similar to that proposed for columnar-joint formation (DeGraff and Aydin, 1993), or pulsed inflation (Anderson and others, 1999, 2000). The pulsed-inflation interpretation is inconsistent with observations of actively inflating sheetflows (Self and others, 2000).

Once identified in Hawai‘i, inflated flows soon were recognized in several different submarine and terrestrial environments. Sea-floor analogs to inflated sheetflows have been documented on submarine extensions of Hawaiian rift zones (Umino and others, 2000; Smith and others, 2002) and in midoceanic environments (Appelgate and Embley, 1992; Gregg and Chadwick, 1996; Chadwick and others, 1999, 2001). Inflated features have recently been recognized in Australian

and South African komatiites (Hill and Perring, 1996; Hol-lamby, 1996; Dann, 2001) and more recent (<200 ka) flows in Queensland, Australia (Atkinson and Atkinson, 1995; Stephenson and others, 1998; Whitehead and Stephenson, 1998).

Identification of Hawaiian-like inflation structures within the Columbia River Basalt Group has led to the suggestion that inflation may be important in the emplacement of large igneous provinces (Self and others, 1996, 1997, 1998, 2000). Tumuli and inflated pāhoehoe flows have been recognized in the older part of the Deccan Volcanic Province, India (Duraishwami and others, 2001), an observation consistent with the low regional slopes of preeruption flood-basalt provinces but raising questions about the limiting conditions of volumetric effusion rate or flow-advance rate in which inflation can occur.

Limited drilling in large igneous provinces in the North Atlantic (Eldholm and others, 1989; Larsen and others, 1994; Duncan and others, 1996) and on the Kerguelen Plateau (Coffin and others, 2000) confirms the presence of inflated basalt flows in both regions, although core samples of ‘a‘ā and transitional flows are also common. Large basaltic provinces, like those of Hawai‘i, probably vary in eruption and emplacement conditions.

Lava Tubes Are Common Within Pāhoehoe Flows and Can Erode Their Bases

As the crust around a pāhoehoe flow thickens, the interior remains molten and continues to flow. Differential movement of solid crust and fluid core constitutes lava-tube flow. Initially, tubes conform to the shape of the flow—wide and not especially tall (low aspect ratio; Cooper and Kauahikaua, 1992). As the crust at the flow margins grows, the flow focuses, and the aspect ratio increases. Lava flowing in a tube may downcut its base in the center, resulting in keyhole-shaped cross sections (Kauahikaua and others, 1998a). Circumstantial evidence indicates that lava either remelts or abrades at the base of a stream (Greeley and others, 1998) and that the surfaces of lava streams within tubes slowly recede from the ground surface without a reduction in stream volume (Peterson and others, 1994). A single field measurement of the rate at which this downcutting can occur—10 cm per day over several months—was obtained during episode 53 (Kauahikaua and others, 1998a). This rate can be modeled by assuming steady forced convective heat transfer by a laminar channel flow at large Péclet numbers (Kerr, 2001).

The process of building a tube from a new vent to the ocean, a distance of 10 to 12 km, has occurred four times during this eruption (episodes 48, 51, and twice during 55), each time taking 2 to 8 months (Heliker and Decker, in press). The complexity of each tube system increased over time as sustained breakouts from the tube formed branches.

Tubes that form on slopes of more than a few degrees generally have headspace (air- and gas-filled space in the

tube above the lava stream), form skylights, and downcut their bases. On flat terrain, such as the coastal plain, tubes generally form as elongate inflated tumuli that remain full and are actually overpressured; downcutting generally does not take place, and skylights generally do not form (Kauahikaua and others, 1998a). Notable exceptions include the Highcastle tube in 1995 and the Lae‘apuki tube in 1996–97, both of which initially formed elongate tumuli. Inflation stopped as soon as downcutting began. Hydrostatic pressure can be transmitted only through the parts of the tube that are full—those that form on flat terrain. Skylights and downcutting are also visible at the coastline in the steeper terrain atop the coastal cliffs.

Flow-field development is controlled by lava-effusion rate. Steady effusion promotes the formation of stable, long-lived tube systems, whereas varying effusion rates result in temporary deflation of tumuli over filled tubes, formation of shatter rings (described below) above wide sections of tubes, or partial collapse of the tube system. Partial collapse commonly results in the formation of a new tube system that widens the flow field.

Lava tubes form readily within pāhoehoe flows. When blockages occur, the lava stream is forced out through the roof onto the ground surface in what is called a breakout. If the blockage is removed, the stream resumes flowing through the tube, cutting off lava supply to the surface breakout. If the blockage persists, the breakout commonly builds another tube that joins the system.

Occasionally a new tube system develops a site of persistent breakouts at the end of a tube segment, at an unusually sharp bend, break in slope, or anywhere else. If breakouts occur on a pali, they form a fan with a surface slope less than that of the original ground slope. If they form on flatter ground, they may transform into rootless shields, as described in detail in the next section.

Several lava falls that were observed in long-lived lava tubes, commonly below skylights, are not necessarily the result of one lava tube flowing into a lower tube, for we observed lava falls forming within a single lava tube (Kauahikaua and others, 1998a). The mechanism seems to be enhanced downcutting and widening of a plunge pool at the base of the falls, rather than headward erosion typical of waterfalls, as had been deduced from observations in ancient tubes (Kempe, 1997). Evidence against headward plucking is provided by the stationary position of lava falls beneath skylights for as long as 20 months, even as the falls’ height increased to approximately 4 m.

The formation of thermally insulated tubes permits lava to be transported great distances without significant cooling. For the Mauna Ulu lava tubes, temperature loss over the entire 12-km-long tube system was less than 15°C, the estimated precision of the optical pyrometer then in use (Swanson, 1973; Swanson and Fabbi, 1973; Peterson and Swanson, 1974). Advances in obtaining in-place liquid lava temperatures by using glass compositions (Helz and Thornber, 1987; Helz and others, 1995) yield a tool for slightly more precise estimation ($\pm 3^\circ\text{C}$) of tube insulation. Helz and others (1993) reported

a consistent temperature drop of 9–10°C over the 12-km distance between the Kūpaianaha lava pond and the coast, decreasing to 7°C as the tube matured (see Helz and others, this volume). Using the same method, a maximum temperature drop of 6°C was estimated for lava in both the episode 48 Waha‘ula lava tube and the episode 53 Kamoamoā lava tube (Cashman and others, 1994). Thornber (2001) estimated a maximum temperature drop of 6–15°C during five sampling runs down episodes 53 and 55 lava tubes between May 1996 and September 1998. These estimates were made on tubes less than 14 km long, in contrast to a similar estimate for a 40-km-long tube system within the A.D. 1445 ‘Ailā‘au lava flows northeast of Kīlauea caldera (Clague and others, 1999). We conclude that lava tubes are remarkable insulators that allow temperature drops of only 6–16°C ($\pm 3^\circ\text{C}$) over 14 km or greater distances ($\leq 1^\circ\text{C}/\text{km}$).

Lava flowing through tubes continues to degas passively, and large bubbles of gas preferentially rise and escape. Gas escape from surface lava flows and tubes accounts for nearly 1 percent of the overall SO_2 output of the eruption (Cashman and others, 1994). Bubbles continue to nucleate at a rate of about 25 percent of the nucleation rate in the vent conduit. Surface breakouts from the tube, extremely depleted in small vesicles, suggest coalescence of bubbles.

When lava tubes drained during episode 53, especially the Highcastle and Lae‘apuki tubes, we were able to observe and analyze encrustations that formed while the tube was active or while it cooled. Analysis of hollow lava (“soda straw”) stalactites revealed that the outside surfaces are enriched in Fe and depleted in Si, and oxide phases and the vesicle walls are enriched in Ti, as discovered previously by Baird and others (1985). This unique enrichment led to the conclusion that these stalactites and stalagmites formed by remelting of the tube walls (Baird and others, 1985; Thornber and others, 1999).

After the tubes cooled sufficiently for rainwater to enter, various sulfates, primarily hydrous, precipitated when water remobilized sulfur deposited within the tube roof while the tube was active (Finch and Emerson, 1925; Thornber and others, 1999). A visit to the Lae‘apuki lava tube 2 years after it drained found lava falls, soda-straw stalactites, and abundant hydrous sulfate stalactites composed of bloedite and thenardite (Porter, 2000).

Morphologic Subtypes of Lava Flows

Before the current eruption, morphologic types of subaerial basaltic-lava flows were assigned to three main categories: pāhoehoe, ‘a‘ā, and blocky (Jones, 1937, 1943; Wentworth and Macdonald, 1953; Macdonald, 1967). Several morphologic subtypes were also defined: entrail pāhoehoe results from lava dribbled down steep slopes, slabby pāhoehoe has an upper surface festooned with broken slabs of cooled crust, and shelly pāhoehoe has thin crusts around a completely empty interior and occurs mainly near vents. Swanson (1973) described lobelike and sheetlike varieties of shelly pāhoehoe

for the Mauna Ulu eruption. Wentworth and Macdonald (1953) also described entrail, shark-skin, filamented, corded, festooned, elephant-hide, and slabby pāhoehoe.

A distinctive subtype of pāhoehoe that attracted attention during the current eruption was variously described as P-type pāhoehoe (Wilmoth and Walker, 1993), dense-glass pāhoehoe (Hon and others, 1994), or blue-glassy pāhoehoe (Oze, 1997; Self and others, 1998; Sage and Mattox, 2000). The various descriptors each emphasize a distinct characteristic of this subtype: “P-type” indicates pipe-vesicle-bearing, “dense-glass” describes a rind of glass as thick as 1 cm with few vesicles, and “blue-glassy” describes the distinctive gun-metal-blue hue of this type of pāhoehoe, which typically emerges from the base of tumuli. Hon and others (1994) proposed that its unique characteristics resulted from being subjected to 2 to 4 bars of excess pressure within a tumulus, forcing gas back into solution before the lava was extruded and cooled. Cashman and Kauahikaua (1997) documented vesiculation in the crust of an inflating sheetflow as subdued with depth relative to normal vesiculation, an observation supporting the existence of fluid pressures higher than lithostatic within the flow. Alternately, Wilmoth and Walker (1993), Friedman and others (1996), and Self and others (1998) hypothesized that after spending a week or two inside inflating flows, the lava cools, and the bubbles coalesce before a pressure surge from the vent forces them out onto the surface. In frequent observations of blue-glassy flows issuing from 1- to 2-day-old tumuli, we noted that breakouts require a minimum tumulus height and are triggered by a threshold pressure (Hon and others, 1994; Kauahikaua and others, 1998a),

rather than a threshold residence time of a week or two. Time-lapse videos of inflating flows confirms the absence of a surge pulse before breakouts (Ka ‘Io Productions, 2000).

The glassy surface of blue-glassy pāhoehoe flows commonly has the shark-skin texture (fig. 3) described by Wentworth and Macdonald (1953), who concluded that the texture results from “the escape of gas from the lava surface, each bubble dragging with it a filament of the enclosing liquid,” and attributed the glass surface to “quick chilling . . . as by heavy rain.” Observations during this eruption reveal that the dense-glass rind can form without rain and that the rough, shark-skin texture is the result of “olivine phenocrysts that were draped by the highly fluid lava during emplacement” (Hon and others, 1994).

Blue-glassy pāhoehoe is one of three subtypes of pāhoehoe that were used as indicators of robust sheetflow and tumulus inflation during the closely watched lava inundation of Kalapana (Mattox and others, 1993; Heliker and Decker, in press). The other subtypes are a shiny, silvery pāhoehoe, probably equivalent to the S-type pāhoehoe of Wilmoth and Walker (1993) and the “silvery” lobes of Self and others (1998), and a pastier, bulbous, duller, rough-surfaced, gun-metal-gray pāhoehoe. The shiny, silvery pāhoehoe is abundant in an advancing pāhoehoe flow. In contrast, pasty, dull pāhoehoe indicates either a reduced lava supply (Kauahikaua and others, 1996) from a draining tube system or residual lava pushed out of a reoccupied tube. Both surface characteristics (pasty and dull) reflect higher groundmass crystallinities afforded by increased transport time.



Figure 3. Blue-glassy flow surface, showing embedded crystals draped to form “shark-skin” texture.

Hornitos, Rootless Shields, and Shatter Form over Pāhoehoe Lava Tubes

Hornitos (fig. 4) formed at least once during episodes 48 and 53 and three times during episode 55 in 2001–2. Each hornito formed directly over a lava tube or intrusion within 4 km of the vent.

The episode 48 hornitos were observed near Kūpaianaha. One stood 1-m high over a lava tube emerging from the south-southeast side of the shield in August 1986. A cluster of three hornitos formed around the edges of a bulge, believed to be the result of intrusion, on the north side of Kūpaianaha in July 1987. A pair of hornitos developed side by side during episode 53, each no more than 1 m high and 0.5 m in diameter at the base. The episode 55 hornitos were 2 to 12 m high and approximately 2 to 3 m in basal diameter.

Each hornito commonly had spatter and Pele’s hair at its base and on the downwind side. Escaping gas could be heard from incandescent holes in the top or sides. The formation of hornitos was commonly preceded or accompanied by a small breakout.



Figure 4. Two hornitos formed in February 2002, about 4 km from Pu'u 'Ō'ō. Hornito above blue case is 15 cm high; distant hornito is 4 m high. Photograph taken February 21, 2002.

The hornitos in 2001–2 began forming when the lava tube under them apparently filled, as inferred from very low frequency (VLF) monitoring measurements. In the upper part of the flow field above Pūlama pali, the lava stream normally occupies only the lower meter or two of the tube beneath a large gas-filled cylindrical cavity (Kauahikaua and others, 1998a), as observed through skylights and by VLF measurements. In several places, we observed hornitos actively forming, spitting lava clots on or immediately adjacent to a VLF tube-monitoring location. Coincident with the hornito activity, the VLF-measured cross-sectional area of lava in the tube showed an abrupt increase that indicated a full tube. When next observed, the VLF-derived cross-sectional area had abruptly decreased, and the hornitos were no longer actively spattering. The local filling of a lava tube could have resulted from a partial obstruction farther downstream that caused lava to back up.

The rootless shields (fig. 5) that form over lava tubes are not fed directly from a deep-seated (>1 km depth) source; instead, the outflowing lava accumulates around the breakout point to create a shieldlike structure. Four rootless shields formed within a period of 2 weeks directly over an active lava tube in September and October 1999, during reoccupation of the tube after an especially long eruptive pause of nearly 12 days. Each breakout site formed a perched pond that overflowed in all directions to eventually form the shields built of predominantly shelly pāhoehoe. The shields can build to 5 to 20 m in height relative to the preshield surface; the largest shield is more than 500 m in diameter and is topped by a flat, ponded-lava surface 175 m across. The four 1999 shields subsequently merged together to form a broad, elongate ridge approximately 2 km long, directly over the lava tube.

Another set of rootless shields that formed in late 2001 to early 2002 at approximately the same distance from Pu'u 'Ō'ō as the 1999 shields were built over continuously active lava tubes, and their growth was not related to a pause. When this episode of shield building started, the shields formed over a tube system that fed two ocean entries. Within about 8 weeks



Figure 5. Rootless lava shield (skyline) at 2,080-ft elevation. Shield is about 500 m wide at its base and nearly 20 m high. View southward; photograph taken January 28, 2002.

after shield building began, both ocean entries stopped, and lava was seen only 1 km beyond the lowest shield. Skylights, never abundant, ceased forming as hornitos and shields were built over the lava tubes.

Shatter rings—broken rubble arranged in approximately concentric circles or ellipses (figs. 6–8)—also form directly over lava tubes on gently sloping ground (Kauahikaua and others, 1998a). Observations during episode 53 and measurements during episode 55 demonstrated that shatter-ring formation is more complex than simple collapse.

As verified by electromagnetic profiles, the shatter rings (two in the upper flow field in episode 53 and four on the coastal plain in episode 55) all formed over active lava tubes. When first noticed, the shatter rings resembled broad, collapsed tumuli. Growth of each shatter ring continued by episodic, nearly constant flexing of the lava-tube roof, as determined by repeat leveling across one of the early episode 55 features and indicated by a persistent grinding sound. The constant flexing breaks the lava-tube roof around the edges of the circular uplifted area. Fractured rock then collects around the rim in circular or ellipsoidal rings, while the remaining tube roof or crust within the center, generally consisting of the original pāhoehoe surface, continues its up-and-down motion. The range of vertical movement in the center can exceed 2 m up or down, with total relief of as much as 4 m relative to the surrounding flow surface (fig. 8). Subsidence into the drained lava tube increases the relief inside some rings (fig. 9). Survey results confirm that only vertical movement takes place, with no net compression or extension.

The two shatter rings formed during episode 53 started building after the eruptive pauses in March and November 1995, respectively, above a lava tube that had made its first ocean entry in November 1992. The first episode 55 shatter ring started to form more than 6 months after the tube was established at that site in July 1997, after the first of two

eruptive pauses in January 1998. Coincidentally, this shatter ring was centered on a geophysical profile established during October 1997 that provided pre-shatter-ring baseline data (fig. 9). The most recent shatter ring was first noticed in July 1999 above a lava tube that reached the ocean nearly 12 months earlier. Lava had resumed flowing through the tube in June 1999 after a 4-day pause.

All of these shatter rings formed over unusually wide sections of the active lava tube on slopes less than 5°. Electrical-conductivity signatures indicate that the lava tubes in this low-slope environment are commonly 3 to 5 m, rarely as much as 10 m, wide (Kauahikaua and others, 1998a). In contrast, electrical-conductivity signatures over active shatter rings indicate a lava-tube width of 50 to 100 m. The 1998 shatter ring formed on the coastal plain, where the underlying tube was 80 m wide when first surveyed in October 1997. Similar observations have been made on Hualālai Volcano for a few shatter rings under which the related tube can be explored. In



▲ **Figure 6.** The 1999 shatter ring on coastal plain when first noticed. Reflective lava inside ring is original ground surface that has not been completely disrupted. Ring is approximately 50 m in diameter. Aerial oblique view north-eastward; photograph taken July 22, 1999.

◀ **Figure 7.** The 1999 shatter ring on coastal plain. View northward; photograph taken August 10, 1999.

the clearest example, the lava tube is 50 m wide beneath a 60-m-diameter shatter ring. Tubes are less than 7 m in diameter upslope and downslope from all the observed shatter rings.

Shatter rings form during several months of continuous lava-tube activity, when lava issues from breakouts around the outside of the shatter ring. The two episode 53 shatter rings were centered on low rootless shields that formed from breakouts around the ring. The three shatter rings above the 1997–98 Waha‘ula lava tube were actively deforming for about 6 months until the tube was abandoned in early July 1998. The most recent shatter ring (figs. 6, 7) grew for about 3 months until a pause drained the underlying lava tube in September 1999.

The fluctuations in elevation inside the shatter ring directly correlate with variations in the volume of lava moving through the tube. The 1998 shatter ring was resurveyed several times in the first half of the year, when seven pauses in lava supply occurred. During the first several pauses, the interior of the shatter ring subsided; when flow resumed through the tube, the interior of the shatter ring rose significantly. The repeated leveling (fig. 10) captured only the range in elevation changes and one rapid rise after the last pause before the tube was abandoned.

Shatter rings have been recognized and puzzled over worldwide. We have mapped dozens on Hualālai, two in the Koa‘e fault system, and several on the lower end of the main ‘Ailā‘au lava tube (Clague and others, 1999) of Kīlauea. A few shatter rings have been observed on Mauna Loa and Haleakalā. Shatter rings are known outside Hawai‘i as “collapsed tumuli” on the 1614–24 pāhoehoe flows on Mount Etna (Guest and others, 1984), “unusual craters” on the Aden basalts of New Mexico (Summerour, 1989), “craters with raised rims” on the Cave Basalt of Washington (Greeley and Hyde, 1972), and “lava ponds” on the Undara lavas of Queensland, Australia

(Atkinson and Atkinson, 1995). They are also reported from Etna (Sonia Calvari, written commun., 1998) and the Hallmundarhraun area of Iceland (Wood, 2001).

Both shatter rings and rootless shields apparently result from unsteady flow of lava through tubes. During the current eruption, shatter rings were initiated by pauses, and rootless shields were built by flows with insufficient lava supply to form tubes. As such, shatter rings and rootless shields provide an estimate of both the longevity and steadiness of tube flow and can be used to assess emplacement conditions of older, solidified lava flows.

Lava Entering the Sea Forms Deltas and Benches, Continues Flowing Underwater, and Alters Nearby Ocean Water

During the 1969–74 Mauna Ulu eruption, lava entering the ocean was described on the basis of depositional environment—subaerial (Peterson, 1976) or submarine (Moore and others, 1973). Lava entered the ocean 5 times between 1969 and 1971, for less than 2 months on each occasion. The pāhoehoe flows built lava deltas, as wide as 400 m and extending nearly 2 km along shore. Small littoral tephra cones grew on the deltas. Lava tubes delivered fluid lava to the ocean floor through the surf zone. Pillow lavas and lava tongues were observed forming and moving on the ocean floor off these entries above 70-m depth, the limit of the diving undertaken.

The current Pu‘u ‘Ō‘ō-Kūpaianaha eruption has produced much longer lived ocean entries than any other eruption of Kīlauea since A.D. 1500 (Don Swanson, oral commun., 2002). Of the nearly 80 ocean entries recorded to date, 30 lasted for



Figure 8. Vertical view and topographic map of 1999 shatter ring. Photograph taken December 23, 1999.

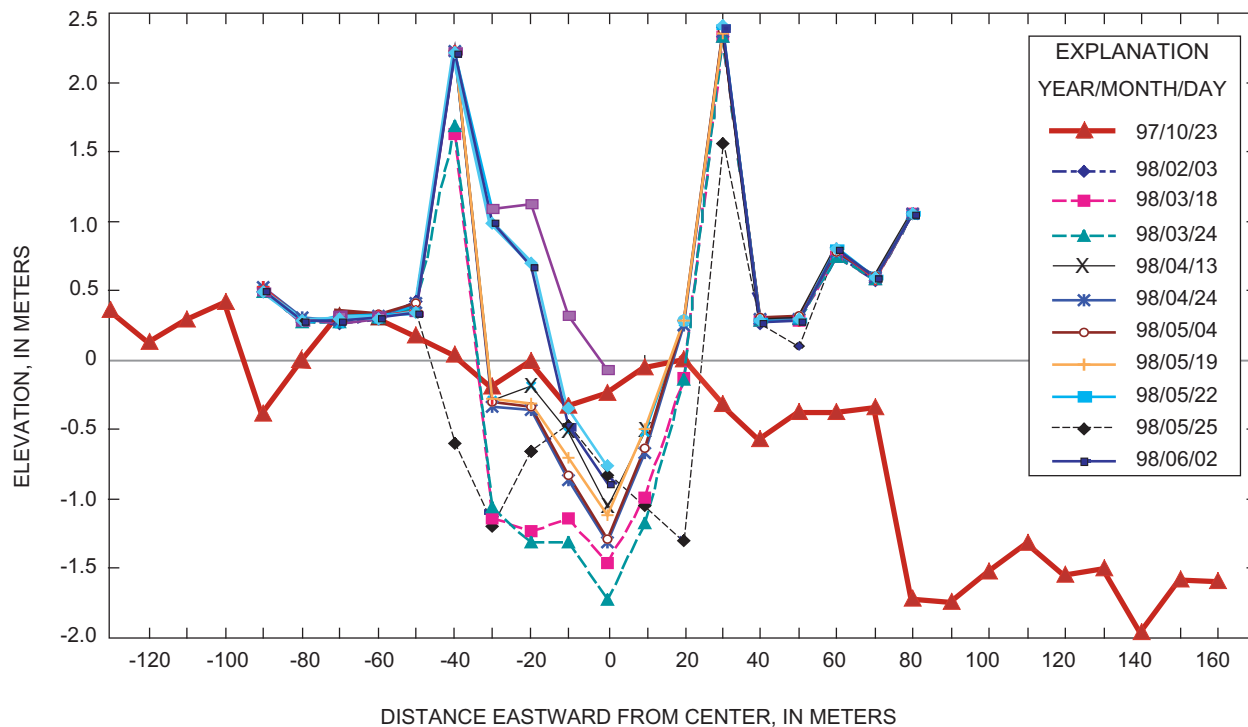


Figure 9. Results of repeat elevation surveys across a shatter ring. First data set, from October 23, 1997 (97/10/23), was measured before initiation of deformation.

more than 2 months, and the longest lasted 15½ months. The first ocean entry occurred in November 1986, and by 1988 the lava deltas were exhibiting a previously unobserved behavior: they were catastrophically collapsing into the ocean (Kelly and others, 1989; Hon and others, 1993b). Collapse resulted from the accumulation of subaerial flows and lava on the delta, which overburdened the hyaloclastite fans on which they were built and caused parts of the delta to calve into the ocean with little warning. The new coastline, which continued building beyond the old one, could be overrun by a new surface flow that would add to the lava delta. This process prompted the fol-

lowing distinction in terminology: A “lava delta” refers to all lava built beyond the preeruption coastline, whereas a “lava bench” refers to a part of the lava delta that has built outward within a previous collapse scar and that could collapse because of its unstable structure. The elevation of a lava bench is therefore abruptly lower than that of a lava delta.

Full or partial bench collapses commonly initiate explosive interactions between lava and ocean water. Mattox and Mangan (1997) describe four different types of explosive events: tephra jets, lithic blasts, bubble bursts, and littoral lava fountains. The explosions result either from open mixing of

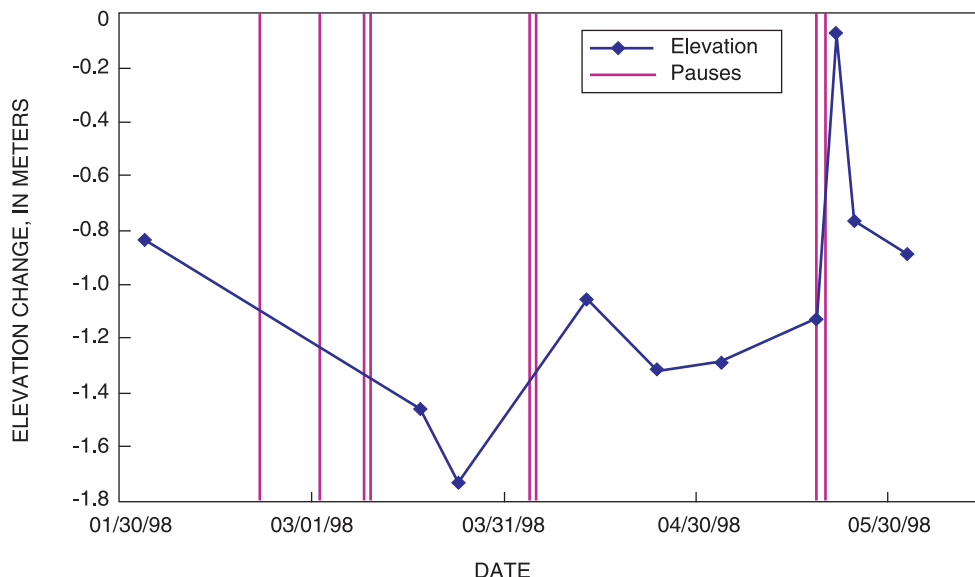


Figure 10. Elevation changes in interior of 1998 shatter ring. Vertical lines denote eruption pauses.

molten lava and seawater induced by wave action or from confined mixing within lava tubes in the bench. The explosions can produce abundant glass lapilli, ash, spatter bombs, Pele's hair, and limu o Pele (Hon and others, 1988). When interactions are persistent, littoral cones can develop.

Lava deltas subside while they are forming (Kauahikaua and others, 1993). Repeated surveying of two lines across the lava delta built off Kamoamoia in 1993 showed that it was subsiding faster at its seaward edge than at the old coastline. The maximum subsidence rate at the new coastline was just less than 1.5 mm per day over the 3 weeks between surveys.

The flexing and subsidence of a lava delta diminish over time after flow activity has ceased, presumably owing to increasing cementation of the submarine hyaloclastite fan. Three survey lines established in 2000 on lava deltas last active in 1996 and 1998 show a vertical change of less than 3 mm in 2 years—the limit of the survey's precision. Notable subsidence characterizes only the outermost survey pin along each line, where the rocks are persistently buffeted by surf. These rocks were visibly downdropped from the position they held when the pin was initially set. On two lines, blocks containing the outermost pins disappeared.

We conclude from these survey data that lava deltas build outward farthest over shallow marine embayments and relatively stable submarine slopes. The slopes then become oversteepened, and later lava flows of the same eruption expand the delta seaward, while supplying the slopes with new hyaloclastic debris. These slopes grow more stable during the decades between major eruptions, and the ground is thus prepared for new seaward growth in the next major eruption.

Elevation profiles of the Kamoamoia lava delta show that the subaerial slope within 250 m inland of the coastline steepens seaward from about 0.4° to 2.3° (Kauahikaua and

others, 1993). Examination of the slope map (fig. 1) confirms that this steepening is a common feature of the new coastline. If lava tubes were at an approximately constant depth within the delta, the steepening would result in a marked increase in flow velocity. This fringe of steepened topography immediately behind the new coast is the most common place on the coastal plain for skylights to form, presumably because the lava stream cuts downward, leaving its roof unsupported (Kauahikaua and others, 1998a).

Underwater observers report both pillow lava and highly channelized lava streams flowing down a steep and unconsolidated submarine slope (Tribble, 1991). Partially crusted lava in the 0.75- to 1.5-m-wide channels advances at a velocity of 1 to 3 m/s, similar to the flow velocities through subaerial lava tubes and small channels. Bubbles of water vapor form and collapse constantly on the hot surface and are commonly audible to divers as explosions. A sonobuoy survey on Lō'īhi could detect both these explosions and more constant sounds indicative of mass wasting relating to bench collapse (Caplan-Auerbach and Duennebie, 2001; Caplan-Auerbach and others, 2001).

We conducted two brief, near-shore bathymetric surveys to determine whether our coastal-hazard analysis could be broadened by monitoring changes on the submarine slopes. Previously published maps show bathymetric contours that trend parallel to the coastline and define a slope dipping 21° in the first 1,000 m of water depth (Chase and others, 1981; Chadwick and others, 1993). After an extensive bench collapse in March 1998, we mapped the seafloor at 100- to 1,000-m depth using a small fishing vessel equipped with an ocean-bottom finder (fig. 11). This collapse destroyed not only the bench outboard of the existing seacliff but also a large tract of land behind the seacliff, 10 ha in all. The most conspicuous features of the survey are a set of shallow

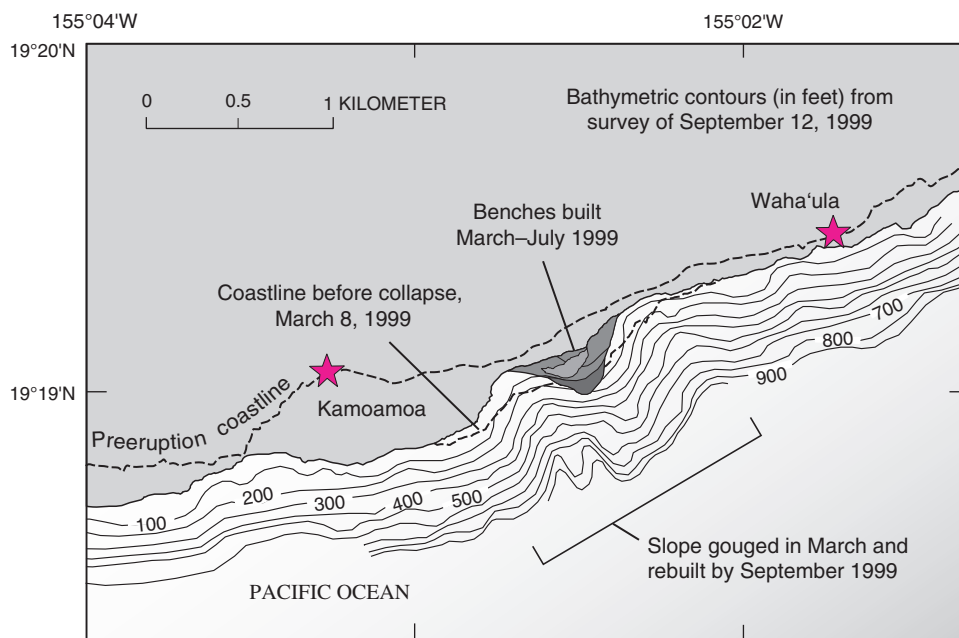


Figure 11. Bathymetry of coastal area near 1999 Kamokuna ocean entry.

canyons in the submarine fan, possibly gouged by collapsing debris. Subsequent regrowth of a lava bench during the 6 months between this collapse and our survey allows an alternative interpretation that the canyons are merely artifacts resulting from bench growth in the intervening sector. We prefer the erosional interpretation because such large canyons are found only along this stretch of our 6-km-long survey, and long-lived lava benches elsewhere along the survey lack conspicuous submarine constructional forms.

The long-lived lava entries have enabled studies of the chemical alteration of the seawater and the composition of the steam plume. Sansone and others (1991), Sansone and Resing (1995), and Resing and Sansone (2002) reported that the seawater near the ocean entry is highly enriched in H₂, Mn, and Si and has high particle concentrations. Temperatures are elevated to at least 69°C in a surface layer 1 to 2 m thick.

A roil (fig. 12) is frequently observed a few tens of meters offshore from an active ocean entry. These roils are crudely circular areas of calmer water that are generally darker than the light-colored plumes of heated water emanating radially away from the molten-lava-contact point (Realmutto and others, 1992). Commonly, two such light-colored plumes occur with a roil between them (fig. 12). Temperature measurements indicate that the roil is an area where heated water is upwelling over a submarine lava extrusion (Sansone and Resing, 1995). The water temperature and chemistry within a roil are nearly identical to those outside the general ocean-entry area. When a roil is viewed with an infrared video camera, however, small amounts of warm water are seen emerging at the roil's center and spreading out radially (fig. 13). Although the general area of the roil is cooler than the warm plumes to either side, the fine structure confirms that the roils are driven by convection of warm water from the sea floor.

Reversible (?) Transition Between Pāhoehoe and 'A'ā

The discussion about the transition between pāhoehoe and 'a'ā became semiquantitative in the reports by Peterson and Tilling (1980) and Kilburn (1981), who described the transition as a relation between viscosity and the rate of internal shear strain. Kilburn (2000) recast the transition in terms of a relation between applied stress and deformation rate, with a primary emphasis on crust formation, and reiterated the long-held belief that the pāhoehoe-'a'ā transition is irreversible. The long Pu'u 'Ō'ō-Kūpaianaha eruption presented numerous opportunities to observe this transition.

Observations

For tholeiitic magma, low-effusion-rate eruptions produce pāhoehoe flows, and high effusion-rate eruptions produce 'a'ā flows, with the threshold at about 5 to 10 m³/s in Hawai'i

(Rowland and Walker, 1990). This data set includes 'a'ā flows from Pu'u 'Ō'ō and episode 48 pāhoehoe from Kūpaianaha. Cashman and others (1999) and Polacci and others (1999) analyzed lava sampled from channels and reported that the lava that solidified to 'a'ā contains at least 30 to 50 volume percent plagioclase microlites, whereas lava that solidified to pāhoehoe has a much lower microlite crystallinity. These observations suggest that plagioclase microlite abundance might be an index for the transition between these morphologic and behavioral characteristics. Hoover and others (2001) and Saar and others (2001) detailed the establishment and increase of yield strength in a Newtonian fluid with the addition of solid particles, such as crystals. Their results reveal



Figure 12. Roil (darker area just outboard of steam plume) between two lighter-colored warm-water plumes. Roil is 23–25°C, and plumes are at least 70°C on either side. Photograph and radiometric temperatures taken May 7, 1998.

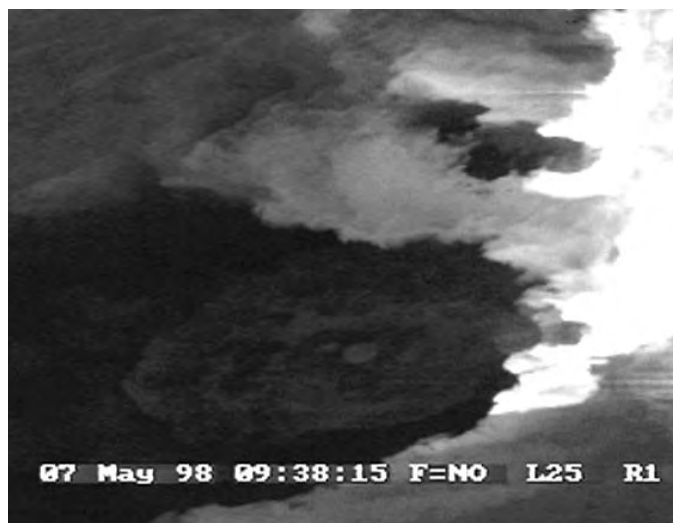


Figure 13. Infrared video image of roil between two warm-water plumes, showing slightly warmer water in center beginning to spread out radially. White area is hot; ocean entry to right. Video taken May 7, 1998.

how microcrystallinity increases the yield strength to the point that a fluid tears under shear stress instead of flowing. The factors that can induce an increase in microlite crystallinity are (1) degassing, either in high lava fountains (see Heliker and others, this volume) or shallow subsurface transport (Lipman and others, 1985); (2) cooling (Cashman, 1993; Crisp and others, 1994); and (3) high internal shear stress, such as that generated gravitationally by flowing over steep slopes.

During the current eruption, lava flows commonly emerged from vents as pāhoehoe over ground that sloped as much as 3°, then changed to ‘a‘ā going over Pūlama pali, which slopes as much as 20°, on the way to the coastal plain slope (0°–1°) and the coast. We observed two distinct behaviors of ‘a‘ā flows that raise questions about the irreversibility of the pāhoehoe-‘a‘ā transition. In the first case, lava flows initially solidifying as pāhoehoe began to solidify as ‘a‘ā as they advanced down steeper slopes, then continued to solidify as a crust transitional to pāhoehoe (see Hon and others, this volume).

In the second case, ‘a‘ā fronts stagnated upon reaching the flat terrain and then leaked fluid lava. For example, the initial episode 55 lava flow (July 1997) traveled over the pali as a channelized ‘a‘ā flow, moved obliquely down the pali, and turned toward the ocean before stalling. Lava solidifying as pāhoehoe apparently broke out of its core and continued 1 km to the ocean. This flow continued to supply lava to the ocean entry for several months as the ‘a‘ā channel fed lava through its front into the tube system of the pāhoehoe flow. The time between the stalling of the ‘a‘ā front and the initiation of the pāhoehoe was, at most, hours; in other words, the pāhoehoe was not a later flow reoccupying the previous ‘a‘ā channel (Kilburn, 2000). Similar behavior was observed for a 1-km-long tube breakout on November 10, 1998. The initial flow was channelized ‘a‘ā that leaked a lava apron which solidified as pāhoehoe. The Mother’s Day, 2002, flow advanced rapidly from the flank of Pu‘u ‘Ō‘ō as an ‘a‘ā flow for the first 3 km, then stalled and continued as a more fluid lava solidifying as pāhoehoe. Thus, the core of these two ‘a‘ā flows was sufficiently fluid to form a pāhoehoe crust. Similar observations were made during the 1991–1993 Mount Etna eruption (Calvari and Pinkerton, 1998).

An Alternative Model

A solidified piece of pāhoehoe cannot convert to a solidified piece of ‘a‘ā, and conversely. Thus, any transition must refer to an identified *pāhoehoe liquid* that can solidify as either a pāhoehoe or an ‘a‘ā crust. If the liquid begins to host slightly solidified chunks on its surface, we would probably call it an *‘a‘ā liquid* and expect only ‘a‘ā crusts to solidify from this liquid. ‘A‘ā liquid might be an accurate description of the liquid’s surface morphology, but the liquid deeper in the channel could still solidify as either an ‘a‘ā or pāhoehoe crust upon leaking to the surface. The transition from low to high microlite crystallinity in the liquid must either precede or accompany the solidification of ‘a‘ā crust.

This discussion shows that the designation of a liquid as pāhoehoe or ‘a‘ā from its morphologic characteristics can be a poor indicator of its future behavior. We therefore suggest that liquid should be identified as neither pāhoehoe nor ‘a‘ā. Those morphologic terms should be used only as the Hawaiians originally used them, for solidified products. The change from one morphology to another along channels can still be defined from solidified products deposited on levees. We can also picture the transition zone as marked by a contour of microlite crystallinity in the liquid. That transition zone might be three dimensional in a channel where the surface liquid solidifies as an ‘a‘ā crust, but the deeper liquid is sufficiently insulated to continue down channel and either bleed out the front of the ‘a‘ā flow or move farther down channel before coming to the surface and solidifying as a pāhoehoe crust.

Implications for Modeling and Interpretation of Remote-Sensing Data

Thermal Models and Applications

Thermal models of lava flows have applications ranging from remote volcano monitoring to volcanic geothermal energy development. The single study of such development during this eruption sought to measure the convective heat flux available in lava during episode 2 (1.8–8.1 kW/m²; Hardee, 1983). The application of thermal models to volcano-hazard and lava-flow studies is more numerous.

Cooling of pāhoehoe flows has been studied by using measured internal and surface temperatures, and modeled simply as conduction with no radiation (Hon and others, 1994). The resulting model works well when using crustal thickness for estimating the age of crust on pāhoehoe lobes within a timespan of hours to days. The model extends the conductive crustal-growth model for months to years on the basis of data obtained from the cooling of Makaopuhi lava lake (Wright and Okamura, 1977). Significantly higher temperatures than predicted by conductive cooling alone were measured, however, beneath an advancing pāhoehoe flow (Keszthelyi, 1995); the high temperatures were interpreted as the sudden release of latent heat from crystallization. These data were used to construct a model that included both conductive and radiative cooling (Keszthelyi and Denlinger, 1996). Their intent was to use this model to predict cooling rates not only of Kīlauea lava but also of lava with different rheology in different environments and even on different planets. This model currently works well for only the first 5 minutes of the cooling process.

Much work has been done to characterize lava-flow cooling by using surface temperatures alone to calibrate satellite-based sensors. Cooling models with two thermal components (Crisp and Baloga, 1990) have been used since 1990 for most work on thermal radiation from both lava flows

and lakes. Flynn and Mougini-Mark (1992) made night-time measurements on an active episode 50 lava flow. The data were modeled as having a crustal temperature of 768°C and a hot core of 1,150°C that composed 3.6 percent of the lava flow area (hot radiating area) upon emplacement. In the next 52 minutes, the crust cooled to 420°C at rates as high as 15°C per minute. Flynn and Mougini-Mark (1994) also made spectroradiometric measurements on an active episode 50 channelized flow. The two-component thermal model was more nearly steady state in this experiment, with a crust at 940°C, a hot core at 1,120°C, and a hot radiating area of 60 percent in the channel center and a crust at 586°C, a hot core at 1,130°C, and a hot radiating area of 1.2 percent at the channel margin. The thermal-radiance model parameters obtained for Hawai'i flows and lava lakes are summarized in table 1.

Surface-temperature studies were summarized by Pinkerton and others (2002), who concluded from their extensive ground measurements that a minimum of four thermal components are required to fully characterize Hawaiian lava flows: core (>1,050°C), viscoelastic skin (750–900°C), rigid solid crust (<750°C), and flow margins (<175°C).

Cooling of Hawaiian pāhoehoe flows is thus well understood. The crust and the much hotter core both cool substantially by radiation in the first few minutes after emplacement, then predominantly by conduction to the air. The flow surface cools rapidly while the crust cracks. The flow interior cools by conduction. It is difficult to generalize these results to flows in substantially different environments.

Estimating Lava Volume Rate from Thermal Satellite Data

Harris and others (1998) proposed a method of estimating instantaneous lava-effusion rate from Thematic Mapper measurements of the total thermal flux of active surface flows. This method produced reasonable matches to VLF-based estimates during 1991 (Kauahikaua and others, 1996). Cloud cover limits this method's application. Wright and others (2001) pointed out that use of Harris and others' method with advanced very high resolution radar (AVHRR) data, which are commonly saturated at active-lava-flow temperatures, yields average effusion rates based on flow area rather than instantaneous effusion rates.

Advances in Lava-Flow-Hazard Assessment and Mitigation

A long-lived and much-studied eruption promotes improved monitoring techniques and methods. Increased understanding of the mechanics of eruptions and lava flows can help mitigate future volcanic hazards in Hawai'i through more precise estimates of both hazard and risk. Low-risk alternatives can then be explored for future land-use planning.

Use of Infrared Imagery in Routine Flow-Field Monitoring

Landsat and other thermal imagery has shown with rare clarity the system of tubes and active flows (Realmuto and others, 1992); however, Landsat images take several weeks to months to acquire, and other specialized imagery may take even longer. Prevalent cloud cover may make it difficult to get a clear view. From July 1999 to July 2002, Landsat 7, which takes an image of the flow field every 16 days, has recorded only three clear views of the entire flow field, 29 views of at least 50 percent of the flow field, and 35 views of mostly clouds. The infrequent acquisition and low yield of cloud-free images make use of this imagery impractical for hazard-mitigation work. Monitoring requires frequent and immediately available imagery.

Beginning with episode 54 in January 1997, the active volcanoes of Hawai'i have been imaged every 15 minutes, using the AVHRR sensor on the geosynchronous-orbiting environmental satellite (GOES) and processed specifically to detect thermal changes in the current eruption (Harris and others, 1997a, b, 2000; Harris and Thornber, 1999). A Web site, designed and maintained by the Hawai'i Institute of Geophysics and Planetology, makes these images available to the public at URL <http://goes.higp.hawaii.edu/bigisland/latest.shtml>. Unless clouds obscure the vent and lava flows, timing of eruption events (lava flows of >10,000 m² area) can be documented within the 15-minute sampling rate of the GOES. The tradeoff for frequent images is the crude spatial locations afforded by the 4- by 4-km pixel size of the thermal band.

Handheld infrared video cameras offer a timelier, more spatially precise alternative. Keszthelyi (1993) demonstrated that lava tubes and flows are easily visible with these tools. We have been using uncalibrated forward-looking infrared radiometer (FLIR) video several times a year for the past 10 years. A single pass over the flow field can catalog flows that have been active in the past 4 to 6 weeks, as well as those active currently. In addition, the infrared images can show the location and configuration of lava tubes beneath the surface in areas without significant surface-flow activity.

Use of an uncalibrated infrared video camera is sufficient to map lava flows and tubes (fig. 14). Relative temperatures are all that we need if the studies are aimed only at cataloging the heat sources and not at understanding the process of cooling.

Telemetered Video Images of Pu'u 'Ō'ō

Since 1997, live video images of Pu'u 'Ō'ō's crater have been available at the Hawaiian Volcano Observatory (Thornber, 1997). These images have proved useful for monitoring eruptive activity within the crater, such as the initiation and growth of vents and the filling and draining of the crater. The images are of relatively low resolution but are sufficient to track changes in the crater between observational visits. The real value of continuous imagery is its potential for correlating visible activity with satellite observations and seismic

Table 1. Summary of radiometrically measured or modeled lava temperatures. Episode numbers refer to Pu‘u ‘Ō‘ō-Kūpaianaha eruption.

Crust temperature (°C)	Core temperature (°C)	Hot fraction (per cent)	Average total radiant flux density (kW/m ²)	Comment
--	1,130–1,145	0.5	--	Episode 4 and 18, Mauna Loa 1984, 1 and 1A (Crisp and Baloga, 1990).
768	1,150	3.6	72.2	Emplacement, episode 50 flow (Flynn and Mougini-Mark, 1992).
410	900	0.5	12.7	52 minutes after emplacement.
940	1,120	60	130	Center of channel, episode-50 (Flynn and Mougini-Mark, 1994).
586	1,130	1.2	29	Margin of channel.
180–572	900–1,250	<2.1	22	1987–1988 Kūpaianaha lava lake, stage 1, fountaining and overturning crusts (Flynn and others, 1993).
100–340	900–1,250	<0.39	5.3	1987–1988 Kūpaianaha lava lake, stage 2, rifting.
80–345	900–1,250	<0.05	4.9	1987–1988 Kūpaianaha lava lake, stage 3, quiescent.
--	--	--	128–170	7/91 Kūpaianaha lava lake (Harris and others, 1999).
400–505	--	--	--	2/2795–3/10/95 Pu‘u ‘Ō‘ō lava lake (HVO unpub. radiometric data).
220–240	--	--	--	3/11/95 Pu‘u ‘Ō‘ō lava lake (HVO unpub. radiometric data).
--	980–1,110	--	--	1995–2001 various skylights (HVO unpub. radiometric data).

and deformation monitors (Harris and Thornber, 1999). Barker and others (this volume) report on an application of this technique.

Monitoring of Lava-Flow-Field Formation and the Volumetric Rate of Lava Produced

During the past 5 years, lava tubes and flows have been mapped by using hand-held Global Positioning System (GPS) receivers. Both commercial- and military-grade receivers have been used, resulting in an accuracy of 5 to 20 m. Many lava-flow contacts are mapped by walking along with GPS in hand, recording waypoints or trackpoints at intervals of 5 to 20 m. Some of the more difficult parts of a flow field are mapped by GPS in a helicopter, with the pilot flying as close to the contact as possible. At times, partial flow contacts located by GPS are completed freehand by referencing oblique aerial photographs. The GPS data are combined into flow polygons for use with GIS software; the eruption-update maps are GIS products. All the episode 55 flow contacts illustrated by Heliker and Mattox (this volume) and Heliker and others (this volume) were determined in this way.

Lava tubes are mapped on the basis of surface features, such as fume, elongate tumuli, skylights, hornitos, and breakouts, and can be more precisely located by using surface geophysics, such as Geonics EM-16 (VLF) or EM-31 conductivity tools (Zablocki, 1978). Molten lava is electrically conductive and so easily detected by using shallow electromagnetic tools (Kauahikaua and others, 1996). We use these techniques weekly, together with visual observations and active flow mapping, to track active lava tubes and flows.

Lava flows commonly originate where a tube ruptures, and so a current map of a lava tube is a basic predictor of where surface flows will originate (Mattox and others, 1993).

In addition to field mapping, numerous digital photographs and some video tapes record flow features from the air and ground. The digital images can be distributed quickly, and the public is then informed, commonly within 24 hours of any new developments, through the HVO website at URL <http://hvo.wr.usgs.gov/>.

Repeated VLF profiles at fixed spots along a lava tube can be used to estimate its cross-sectional conductance. Lava is the only conductor, and we can independently estimate its conductivity. The estimated conductance allows an easy way to monitor the cross-sectional area of fluid lava in the tube. In combination with velocity measurements made at a nearby skylight with a radar gun, a volumetric flux can be estimated. These measurements can be made nearly anywhere on the lava-tube system, but those nearest a vent on the master tube are better approximations of the total output of the vent. Linearly declining flux estimates accurately predicted the demise of the Kūpaianaha vent in early 1992 (Kauahikaua and others, 1996). Similarly declining flux estimates on just a branch of a lava tube without corresponding decline in either the total flux or the other branches has signaled the end of activity of that branch. These geophysical results have compared favorably with a flux estimated from differences in the volume of the flow field (Rowland and others, 1997), satellite thermal-flux data (Harris and others, 1998), and gas-emission data (Sutton and others, 2001; this volume). Currently, these estimates are made once per week, a frequency that is easily capable of demonstrating long-term behavior but incapable of showing response to short-term events either at Pu‘u ‘Ō‘ō or at Kīlauea’s summit.

Prediction of Lava-Flow Behavior

The ultimate goal of lava-flow-hazard mitigation is prediction of the direction and advance rate of lava flows. That challenge has inspired the development of computer code to simulate lava flows (Ishihara and others, 1989; Young and Wadge, 1990; Miyamoto and Sasaki, 1997, 1998; Harris and Rowland, 2001). Much of the physical parameterization of lava flows in terms of viscosity, yield strength, and density has improved computer simulations. More robust approaches to develop limited forecasting tools include the delineation of lava sheds and preferred pathways for possible lava flows.

Lava Sheds and Preferred-Gravitational-Flow-Path Maps

Rather than simulating lava flows, complete with complex rheology, we may be able to make significant contributions by splitting the prediction question into several parts. A primary question is, where will a lava flow go? Answering that question is a simple terrain analysis problem solvable with standard GIS tools.

A catchment (Guest and Murray, 1979) or lava shed (Kauahikaua and others, 1998b) is an area within which any lava flow will be confined if it is erupted from a vent within that lava shed. Lava sheds are computed as watersheds that drain into the ocean. When an eruption begins and the position and configuration of the vents are known, the vents can be plotted on a lava-shed map. The initial lava flow is predicted to advance within any lava shed that contains a part of the eruptive vent. A lava-shed map can also determine which areas are topographically shielded from lava flows (Guest and Murray, 1979).

A refinement on this idea is to include the major pathways, or paths of steepest descent, predicted for gravitationaly driven fluids on the island's surface. These pathways also are easily computed with standard GIS tools. Hanley and Zimelman (1998) evaluated such computed paths as predictors for lava flows from the first 18 episodes of activity at Pu'u Ō'ō and calculated that they account for 60 percent of the flow's location and orientation. A "steepest path" map for part of the Island of Hawai'i, showing those pathways that drain at least 1 km² of surface area for the region including the lower flows erupted from Mauna Loa in 1984, is shown in figure 15. Both the lava sheds and the preferred pathways

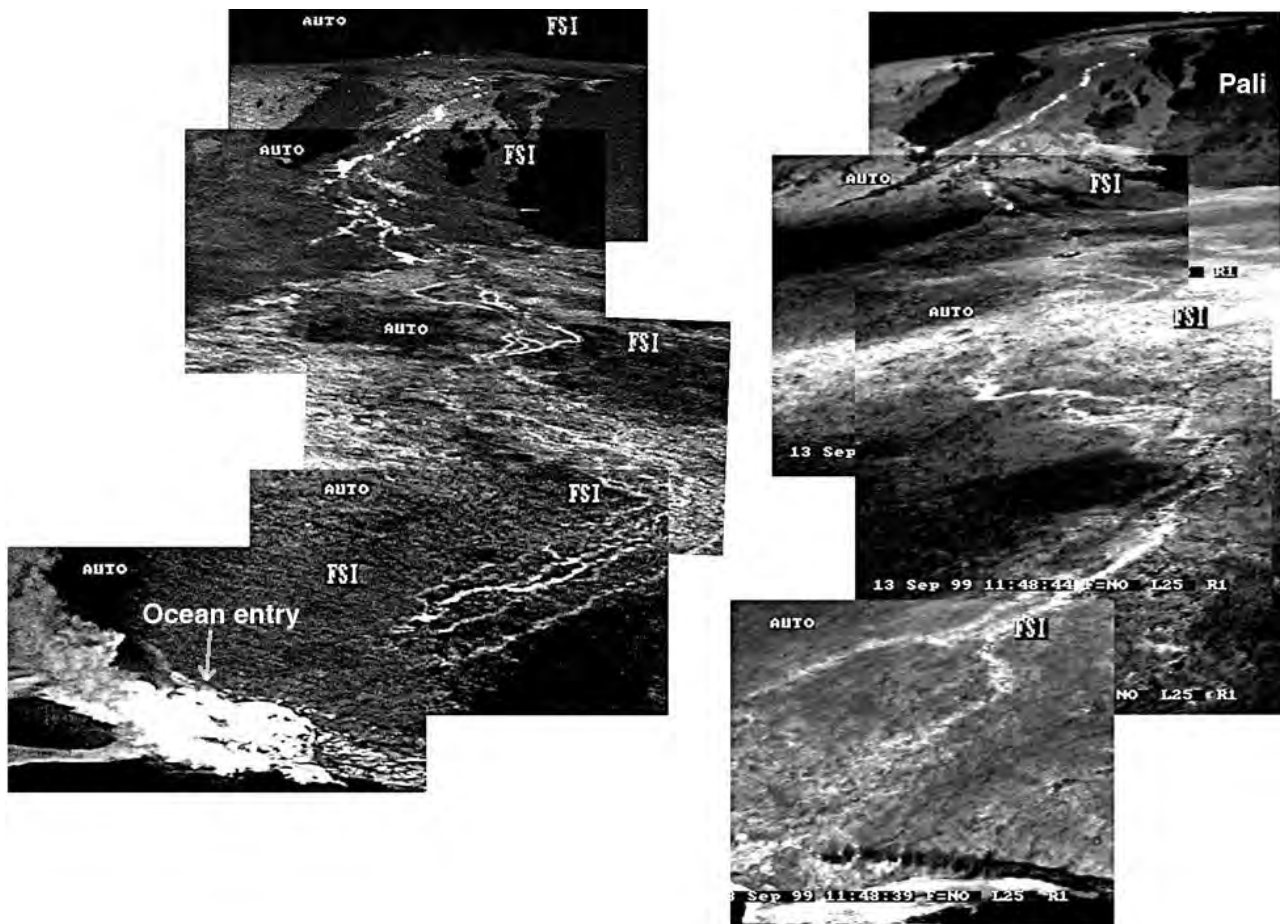


Figure 14. Composite infrared video images of lava-tube system, showing young system with braided streams. In right image, braids have consolidated into a few main braids. Note the redirection of the lower extent of the two braids near coastline. Videos taken December 29, 1998 (left), and September 13, 1999 (right).

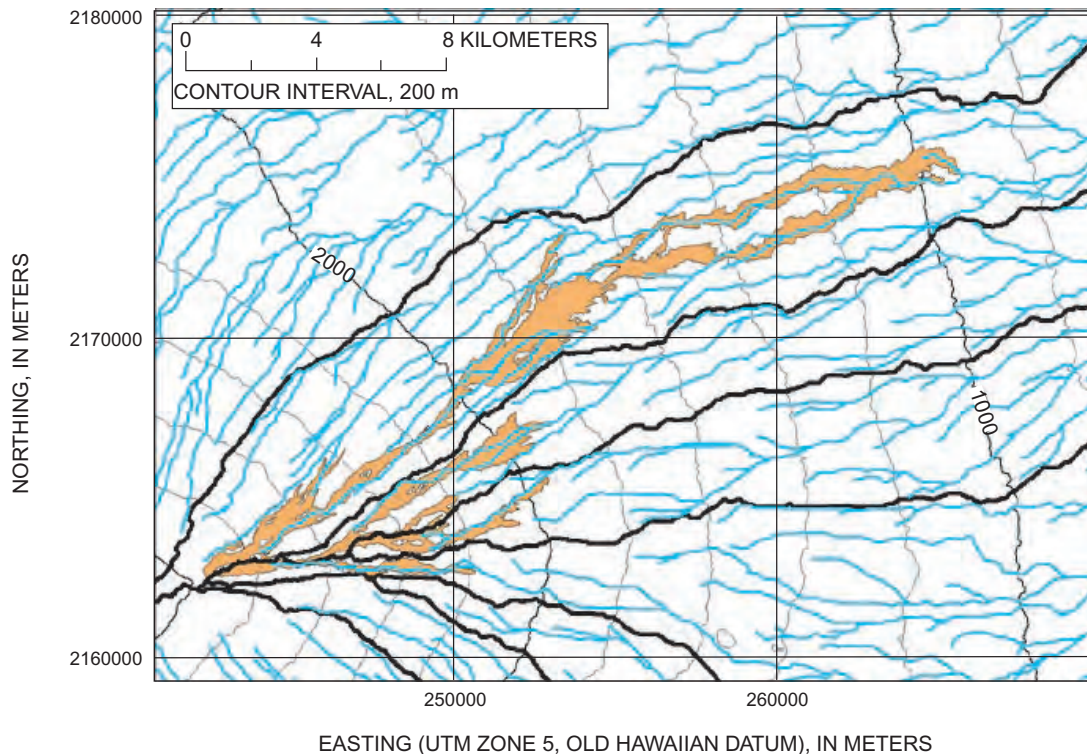


Figure 15. Lower 1984 lava flows from Mauna Loa (orange), Island of Hawai'i. Heavy black lines outline lava sheds; blue lines denote estimated preferred lava pathways.

were estimated from 1978 topography. Note how well both the lava sheds and the preferred pathways would have predicted the ultimate path of the 1984 lava flows. Several such pathways may exist within a single lava shed, and so spatial predictive capability can be further refined.

Estimation of Probabilistic Lava-Flow Hazards

Probabilistic-hazard maps address the question “How often will lava flows inundate a given area?” If we can assume that future inundation will be statistically identical to past inundation, then the question becomes “How often in the past has lava inundated a given area?” Three steps are needed to answer this question. The first step is to obtain a digital geologic map in which each lava flow is mapped and dated. The second step is to determine the distribution of lava-flow-recurrence intervals within subregions. The third step is to choose an appropriate statistical distribution (for example, Poisson or Weibull) to estimate the average recurrence interval from the frequency of lava-flow occurrence. The average recurrence interval can then be used to estimate the probability of future inundation (Kauahikaua and others, 1995a). This methodology was applied to estimating the lava-flow hazard for the east rift zone geothermal subzones (Kauahikaua and others, 1995b) and for a proposed prison site on the northeast rift zone of Mauna Loa (Kauahikaua and others, 1998b). The results give the probability of lava-

flow inundation within some nominal period, chosen to be 50 years for those studies.

Two problems arise when estimating inundation probabilities on the basis of past lava-flow frequencies. The first problem is the general condition that younger flows progressively bury older flows, and so an uncorrected frequency-versus-age histogram for any volcano shows exponentially decreasing frequency with age. The second problem is the apparent disparity in eruption frequency between the past 150 years of close observation and previous periods represented only by mappable and datable lava flows. These problems have been overcome; correction methods were compared by Kauahikaua and others (1998b).

An alternative approach is to estimate the probabilities of lava-flow inundation from the combined estimate of the probability of an eruption occurring anywhere on the volcano, the probability of the eruption location, the probability of a lava flow being generated, and the probability of specific lava-flow parameters (such as composition, effusion rate, or temperature; Newhall and Hoblitt, 2002). This approach has been used for Mount Etna (Wadge and others, 1994) where lava flows were simulated by using both a stochastically chosen vent site and a set of parameters from a library of such parameters for lava flows erupted between 1763 and 1989. The resulting map displays the frequency of inundation for areas around the volcano. A total of 380 flows were simulated, representing a 2,400-year period. The zones of greatest hazard were areas inundated by more than 10 flows at an average recurrence interval of less than 240 years. Changes in topography were not

explicitly incorporated into their computer simulations, and so Wadge and others emphasized that the resulting map is not “a map of the likely coverage of lava flows in the next 2,400 years but a map of the potential for inundation by lava flows now”—that is, areas inundated by the most simulated flows are those of highest hazard for the next flow but not necessarily those that will be inundated by the most flows in the next 2,400 years. The next flows will change the topography and alter the paths of all future flows. For comparison, Wadge and others (1994) presented a map of historical frequency of lava-flow inundation for the period 1763–1989, within which the maximum frequency is only six flows. Although this estimate of the lava-flow hazard has been labeled “probabilistic,” the result is not easily converted to probability of inundation.

Future Directions in Monitoring

Significant progress in volcano monitoring has been made over the last two decades, but we should not conclude that little is left to do to advance our understanding of lava-flow dynamics. Lava erupting from a vent drives all the flow-field processes, and so we could and should better understand the conduit system within Pu‘u ‘Ō‘ō and exactly how it connects to the lava-tube system. The cause of the numerous collapse structures both inside and outside Pu‘u ‘Ō‘ō crater is unclear. Do they overlie the upper reaches of a deep tube exiting flank vents on the west side of the cone, or are they part of a structure, such as concentric fracturing around the base (see Heliker and others, this volume)?

Further progress could be made by improving our current ability to monitor the lava output from Pu‘u ‘Ō‘ō. Elevation data over the entire flow field (possibly from satellite- or aircraft-borne radar) for time intervals during which no lava entered the ocean could help calibrate or check volume-rate estimates obtained by the currently used techniques of VLF electromagnetic profiles or gas monitoring. New monitoring technologies are being developed to study lava-tube flow by passive seismic listening or active ultrasound probing (Rick Hoblitt, oral commun., 2002). Development and increasing reliability of such tools, in addition to an increase in the frequency of data acquisition by way of telemetered sites, should improve our understanding of the short-term relation between deformation, seismic events, and lava output. We will be better able to correlate short-term flow-field events, such as tube ruptures, breakouts, and shatter-ring and rootless-shield development, with short-term and, possibly, subtle variations in lava supply. Our current monitoring capabilities and weekly data allow resolution of only long-term changes and relations.

More can be learned by combining remote sensing technologies with traditional ground-based observations. Multispectral satellite images are invaluable but are too infrequently obtained to be routinely useful. It would be ideal to be able to acquire the same sort of data on demand rather than waiting for a favorable satellite pass and clear sky, such as by using

hand-held multispectral imagers from the ground or a helicopter. Imagers could also be borne on small remote-controlled blimps, balloons, or aircraft. Interferometric radar for measuring topographic changes may be usable from aircraft or from the ground.

We would also like to progress with real-time telemetered information, such as tube-flux monitors based on the VLF electromagnetic technique, microphones for monitoring gas jetting, and stationary video monitors, possibly including infrared sensors. Multiple telemetered sites and continuous data sets could allow synchronous observation of lava vents and tubes, or of several vent sites—data crucial to deciphering vent structure and mechanics.

More quantitative measurements and change detection could be done with oblique digital photographs. We currently take many photographs from the ground and helicopters and can keep track of most changes through familiarity with features by personnel. Subtle changes may be missed, however, until they increase and become more significant. Time-lapse videography has shown great promise in revealing lava flow processes that happen slowly, such as inflation of both pāhoehoe and ‘a‘ā channels (for example, Ka ‘Io Productions, 2000).

Last, but not least, much remains to be understood about lava flow emplacement. Although we have empirical interpretations for the emplacement modes of pāhoehoe and ‘a‘ā, and for the various transitional types of lava morphologies observed in the last two decades, quantifying their emplacement conditions would be a significant advance.

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