Iflation rates, rifts, and bands in a pāhoehoe sheet flow

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ABSTRACT

The margins of sheet flows—pāhoehoe lavas emplaced on surfaces sloping <2°—are typically delineated by structures that form to accommodate vertical flow inflation. We refer to these structures as inflation rifts. The surfaces of inflation rifts almost always exhibit bands of varying color and texture. Various explanations for the bands have been proposed, but active band formation has never been documented. In order to test our hypothesis that banding is caused by changes in the inflation rate, we collected time-lapse photographs of the margin of an actively inflating flow and simultaneously measured the height of the flow with an extensometer. Data collected over a period of ~1 d indicate that the height of the flow margin changed in a stepwise manner and that rate changes correlate with band formation. This confirms our hypothesis.

Inflation and rift-band formation is probably cyclic, because the pattern we observed suggests episodic or crude cyclic behavior. Furthermore, some inflation rifts contain numerous bands whose spacing and general appearances are remarkably similar.

We propose a conceptual model wherein the inferred cyclicity is due to the competition between the fluid pressure in the flow’s liquid core and the tensile strength of the viscoelastic layer where it is weakest—inflation rifts. The viscoelastic layer consists of lava that has cooled to temperatures between 800 and 1070 °C, it becomes viscoelastic (Hon et al., 1994). The 1070 °C isotherm will move into the flow according to:

\[ C_{1070} = 0.0779 \sqrt{t} \]

where \( C_{1070} \) is the distance from the flow top to the 1070 °C isotherm in meters, and \( t \) is time in hours (Hon et al., 1994). The 1070 °C isotherm marks the boundary between crust and liquid. Crust forms more slowly at the base of the flow; at a given time, the thickness of the lower crust is estimated to be 70% of the upper crust. The total crust thickness is given by:

\[ C_{\text{total}} = 0.132 \sqrt{t} \]

As determined from studies of Hawaiian lava lakes (Wright and Okamura, 1977), viscoelastic behavior gives way to brittle behavior when the temperature drops below 800 °C. The depth to the 800 °C isotherm (Hon et al., 1994), and thus the thickness of the upper brittle layer, may be calculated from:

\[ C_{800} = 0.0473 \sqrt{t} - 0.0233, \]

where \( C_{800} \) is the depth in meters, and \( t \) is \( \geq 0.243 \) h. Because the 800 °C isotherm follows the 1070 °C isotherm into the flow, the thickness of viscoelastic crust can be estimated by subtracting Equation 3 from Equation 1:

\[ C_{1070} - C_{800} = 0.0306 \sqrt{t} + 0.0233, \]

where \( t \geq 0.243 \) h.

The tensile strength of the viscoelastic skin and its ability to retard the flow of the liquid increase as it cools. In response to an increased resistance to flow, lava backs up in the conduit feeding the flow, and the fluid pressure increases at the flow front. Once formed, the skin undergoes little or no horizontal movement, but moves vertically as the fluid pressure increases and the flow inflates.

Liquid basalt commonly emerges at or near the surface undergoing inundation, where the skin is thinnest, and thus the resistance to flow is least. These processes are illustrated in Figure 1, a photograph showing pāhoehoe lava advancing over tephra. Liquid is emerging at the tephra surface, and scoria fragments have become embedded in the newly formed skin. The skin is deforming plastically and, along with the embedded fragments, is being translated vertically as the flow inflates.

Some sheet flows advance as an anastomosing mass of pāhoehoe toes. Most toes grow, collide with their neighbors, and coalesce while their skins are still plastic. Once coalesced, the mass behaves as a unified whole, with a continuous liquid interior contained between cooling upper and lower crusts (Hon et al., 1994).

Flow advance is controlled by the cooling rate, which determines the strength of the visco-

INTRODUCTION

On Hawai‘i’s Kilauea Volcano, pāhoehoe flows emplaced on surfaces sloping <2° have been observed to form sheet flows, i.e., smooth, flat-topped lava flows with margins that exhibit vertical scarps or rotated, outward-dipping lava slabs (Hon et al., 1994). These marginal features are evidence of flow inflation. The formation of pāhoehoe sheet flows was described in Hon et al. (1994). Features diagnostic of flow inflation were described by Nichols (1939), Walker (1991), Hon et al. (1994), Chitwood (1994), Self et al. (1998), and Walker (2008). In Hon et al. (1994), it was suggested that sheet flow inflation may be operative in the emplacement of flood basalts. This possibility was discussed by Self et al. (1998, 2000) and Anderson et al. (1999, 2000).

Sheet flows advance either as thin sheets of lava, initially only centimeters to decimeters thick, or as an anastomosing mass of decimeter-thick pāhoehoe toes. At its eruption temperature (1130–1170 °C; Thornber, 2003) and at low strain rates, the lava behaves as a Newtonian fluid (James et al., 2004). However, the exposed liquid at the front of an advancing flow chills rapidly, and its viscosity increases. When the temperature of the exposed outer layer drops to ~1070 °C, it becomes viscoelastic (Hon et al., 1994; James et al., 2004).

The 1070 °C isotherm will move into the flow according to:

\[ C_{1070} = 0.0779 \sqrt{t}, \]

where \( C_{1070} \) is the distance from the flow top to the 1070 °C isotherm in meters, and \( t \) is time in hours (Hon et al., 1994). The 1070 °C isotherm marks the boundary between crust and liquid. Crust forms more slowly at the base of the flow; at a given time, the thickness of the lower crust is estimated to be 70% of the upper crust. The total crust thickness is given by:

\[ C_{\text{total}} = 0.132 \sqrt{t}. \]

Geosphere; February 2012; v. 8; no. 1; p. 179–195; doi:10.1130/GES00656.1; 17 figures; 3 animations.

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elastic skin, and the rate at which liquid lava is supplied to the flow front. As long as the supply rate is sufficiently high, new low-strength skin is constantly generated at the flow front, and the flow advances. Fed by a roughly constant lava supply, the flow area and periphery increase with time. The rate of advance decreases, and more time is available for cooling to increase the thickness and strength of the skin. Locally the skin strengthens sufficiently to prevent liquid from flowing forward. Pauses in the advance are usually brief along the flow terminus but last longer along the flow margins behind the terminus, where they can become permanent.

Tensile stresses in the brittle crust—caused by cooling-induced shrinkage and continued influx of liquid into the core of the flow—produce fractures that propagate inward until they encounter the viscoelastic layer. Tensile stress that exceeds the layer’s yield strength results in ductile extension, often by tearing. If the tensile stress is great enough, and if it is applied rapidly enough, the layer will undergo ductile failure. Once a failure develops, it tends to persist, because it is the line of least resistance. The two sides of the failure surface separate as they are pushed apart by the pressurized liquid in the flow interior. As inflation proceeds, the failure line propagates inward, and the separation between former positions of the failure line increases. This process produces a curvilinear channel of increasing width and depth. In a cross section normal to the failure line, the channel is typically V shaped, though the sides of the V are commonly curved.

In Hon et al. (1994), such a channel was referred to as a “crack.” Walker (1991, 2008) called them “lava-inflation clefts.” We prefer the term inflation rift, or simply a rift. This usage is consistent with the following meaning of the term “rift” (Neuendorf et al., 2005): “A narrow cleft, fissure, or other opening in rock, made by cracking or splitting.” We refer to the failure line along which rift propagation occurs as the rift line. We refer to the process of rift formation as rifting. Rifting is bilateral; i.e., it creates a pair of surfaces as newly formed crust is pushed away from the rift line. We refer to these surfaces as rift surfaces. The term rift viscoelastic layer refers to the viscoelastic layer along the rift line, while non-rift viscoelastic layer refers to the viscoelastic layer that is outside the influence of the rift.

Inflation rifts are common in pāhoehoe lavas on Kīlauea’s coastal plain, particularly in tumuli and sheet flows. In tumuli, the extension is dominantly horizontal, the rift tends to propagate downward, and so the V is commonly pointed downward. In sheet flows, inflation rifts are typically best developed at flow margins. This is because the viscoelastic layer is thinnest at flow margins, and thus most susceptible to extension. When fluid pressure at the top of the flow’s liquid core exceeds the combined weight of the upper crust and the tensile strength of the rift viscoelastic layer, the upper crust is uplifted with little or no disruption. The vertical uplift is accommodated by rifting on the flow margins, where two rifting styles are evident. The dominant style is characterized by outward-dipping slabs of lava, while the subordinate style is characterized by vertical scars (Hon et al., 1994).

In the vertical-scarp style, the rift line propagates inward and upward, the V is pointed horizontally into the flow interior, and the rift surfaces are relatively accessible for examination. Cross sections in Figure 2A illustrate the formation of a vertical-scarp margin.

The dominant tilted slab marginal style is more complex (Fig. 2B). As the slab rotates, one rift line propagates inward and upward from the hinge line at the base of the slab while a second rift line simultaneously propagates downward and inward from the top of the slab. The rift at the base of the slab is rarely exposed in natural outcrops. With persistent inflation, the lower rift becomes inactive while the upper rift continues to inflate. In extreme examples, the persistent upper rift develops into a dominant vertical scarps with a subordinate tilted slab at its base. The tilted-slab rift style commonly transitions laterally into the vertical-scarp style within embayments in the flow margin.

An example of rifting in the vertical-scarp style on the margin of an actively inflating sheet flow is shown in Figure 3. The example illustrates a nearly ubiquitous feature on inflation-rift surfaces: bands. We refer to these as rift bands or, simply, bands.

Rift bands on sheet flows were described in Nichols (1939) and Hon et al. (1994). Walker (1991) described them on “lava-inflation clefts” in tumuli. Bands range in width from ~1 cm to a few tens of centimeters and to ~10 m or more in length. Rift bands are distinguishable by contrasts in color and/or surface texture. Unoxidized bands are typically gray or greenish-gray. Oxidized bands exhibit various shades of red, orange, purple, tan, or brown. Commonly, bands of relatively smooth, dull gray to reddish-gray crystalline basalt alternate with rough, glassy, sometimes spinoce gray to greenish-gray veneers a few millimeters thick. Irregular extrusions of smooth, glassy lava parallel the other bands on some rift surfaces; we refer to such rift bands as extrusion bands. Rift bands sometimes exhibit a crude bilateral symmetry across the rift.

Nichols (1939) suggested that the bands formed as hot gases escaping from the propagating crack variably melted the adjacent surface. Walker (1991) attributed the glassy veneers to lava that oozed up from the rift. In Hon et al. (1994) it was suggested that the crystalline and glassy bands correspond, respectively, to brittle
and ductile fracture propagation events, analogous to chisel marks on fracture surfaces of columnar joints (Ryan and Sammis, 1978). The glassy extrusions were attributed to fractures that propagated completely through the viscoelastic layer and into liquid lava. Anderson et al. (1999) suggested that banding is due to injections of small batches of liquid lava beneath the cooled flow crust through a network of preferred pathways.

We hypothesized that rift bands reflect changes in the rate of sheet flow inflation. Quantitative data to support this possibility did not exist. The only published measurements of pāhoehoe sheet flow inflation are those of Hon et al. (1994) and Kauahikaua et al. (1998), whose data document flow height versus time for a number of flows for periods of as long as 300 h. The inflation history for a given inflating flow is approximated by a power law:

\[ H = bt^a, \]  

where \( H \) is flow height in meters, \( b \) and \( a \) are constants, and \( t \) is time in hours.

While the data of Hon et al. (1994) and Kauahikaua et al. (1998) document flow inflation over substantial time periods, the time intervals between observations are also substantial. This is a consequence of the measurement method. For each record, series of vertical angles to a point on a flow surface were measured manually with a theodolite. Inflation rate changes that might have accompanied rift-band formation are not apparent in the data.

Fox et al. (2001) and Chadwick (2003) documented inflation and drainback of a near-vent submarine sheet flow on the Juan de Fuca Ridge. They serendipitously measured vertical displacements when an instrument platform carrying a pressure sensor became embedded in a flow. Pressure data recorded at 15 s intervals indicate that the flow inflated 305 cm at variable rates for 72 min, and then deflated 281 cm at an approximately linear rate for 81 min. While the time resolution is rather high, it is unknown whether inflation rifting and rift banding occurred.

We will now describe an experiment designed to test the hypothesis that rift banding is caused by changes in the inflation rate.

**EXPERIMENTAL PROCEDURES**

In early April 2004, lava from breakouts near a kīpuka (land surrounded by one or more younger flows) on the Pu‘u ‘Ō‘ō upper flow field began to creep seaward toward Pīlālama pali, the large fault scarp that separates the upper flow field from the coastal plain (Fig. 4). Because the kīpuka contained a few banana trees, the flow was informally called the Banana flow. The Banana flow began to descend the steep slopes of Pīlālama pali on 13 April and reached the gently sloping coastal plain on 3 May. The flow terminus intermittently advanced toward the sea; a few days of rapid progress were followed by a few days of stagnation. While the terminus stagnated, the flow inflated, breakouts behind the terminus expanded laterally and then pushed seaward along the flow margins, eventually overtaking the stagnant terminus (Fig. 5; Animation 1). The stop-start cycle continued until the flow reached the sea late on 30 May or early on 31 May.

In the course of routine mapping of the Banana flow on the coastal plain, we noted that the flow was undergoing vertical inflation and decided to document it. Three previous attempts to document the inflation of other flows had failed. Our first attempt on the Banana flow, on 11 May, failed because the expansion stopped shortly after the experimental apparatus was deployed. Data collection was discontinued and by the following morning the measurement site had been inundated with lava.

The failures were due to the difficulty of selecting a site on a flow margin that would develop into a vertical-scarp rift. Vertical-scarp formation would permit us to document the vertical displacement of the flow top, as well as the development of bands on rift surfaces. Only the upwind flow margin was feasible, because air that is heated as it passes over the lava renders the downwind flow margin too hot for instruments or their operators. When we set up an experiment before the rift style was apparent, we inadvertently selected a tilted-slab margin. When we waited until the inflation was well under way to ensure that we had selected a vertical-scarp margin, inflation had already

\[ \text{Figure 2. Sequences of cross sections showing development of rifting styles on the margins of sheet flows. Gray indicates brittle lava; orange indicates viscoelastic layer and core liquid; black indicates preexisting surface. (A) Vertical-scarp style. (B) Tilted-slab style.} \]
ceased, or ceased shortly after setup. Flow margins were prone to lava inundation, so the experiments required constant attention.

On 12 May, we redeployed our experimental apparatus near the previous day’s attempt, but on a new flow margin that formed during the previous night (Figs. 6). To measure vertical flow inflation, we used a UniMeasure Model HX-P510 extensometer capable of measuring displacements of 0–200 cm over 0–10 V. The extensometer was housed within a military steel ammunition box (Fig. 6), along with a 12 V lead-acid battery for power, and an Omega Model OM-CP-Volt 101 data logger. A bracket on the front of the steel box (not visible in Fig. 6) was firmly secured to the old lava surface, adjacent to the inflating flow, with a hardened steel nail.

We strung fine-gauge stainless steel cable from the extensometer, through a pulley mounted on a ceramic anchor block immediately adjacent to the inflating flow margin, to a metal bar mounted on a concrete anchor block on top of the flow margin (Figs. 6 and 7). As the flow inflated or deflated, the wire advanced or retreated, changing the output voltage of the extensometer. Although the pulley slightly increased the complexity of the apparatus, it offers three important advantages. First, it increases the length of cable pulled from the extensometer as the flow inflates; this improves the accuracy of the measurements. Second, the attitude of the cable is constant as it is pulled from the extensometer; this keeps the frictional resistance constant throughout the experiment. Third, the cable is kept close to the ground between the extensometer and the pulley; this minimizes wind-induced noise.

The data logger, powered by its own internal battery, is capable of recording a maximum of $2^{16}$ 16-bit voltages at intervals ranging from 2 s to 12 h. We configured the logger to record the extensometer output voltage every 10 s and started and/or stopped data acquisition on site via the serial port on a laptop computer. The manufacturer specifies the time accuracy of the data logger as ±1 min/month. The accuracy of system time on the laptop computer determines the accuracy of data logger start time. Computer system time was calibrated with a global positioning system clock a few hours before starting the logger, so the start time is accurate to within ±1 s. The range of the data logger is 0–15 V, with a resolution of 0.5 mV. The raw output voltages were subsequently converted to cable-length changes using measured extensometer calibration values. The length changes, in turn, were converted to vertical flow displacements, using the geometry and equations shown in Figure 8.

Uncertainty analysis (Taylor, 1997) shows that the accuracy and precision of calculated vertical flow heights are 0.4 cm and 0.1 mm, respectively. A diurnal wind pattern common to the island of Hawai‘i—trade winds during daylight hours and low-to-nil winds at night—was active during the experiment. Wind-induced noise reached a maximum of ±0.5 mm during the day and diminished to nil at night. The length of the extensometer’s stainless steel cable did not change significantly in response to temperature changes.

In addition to the extensometer, we employed an automatic camera system (Fig. 6; Orr and Hoblitt, 2008) to photograph the margin of the inflating flow at 0.5 min intervals. The photograph time uncertainty is 0.5 min. The resulting time-lapse sequence (Animations 2 and 3)
permitted us to recognize inflation events that were otherwise too slow for the unaided eye to detect and to compare these events to the extensometer record.

RESULTS AND DISCUSSION

The experiment lasted a little more than 24 h, from 11:19 a.m. Hawaii-Aleutian Standard Time on 12 May 2004 to 11:37 a.m. the next day. During this interval, the flow height increased ~38 cm (Fig. 9). From inspection of Figure 9, it is obvious that the vertical displacement did not smoothly follow a power law, as would be expected from the results of Hon et al. (1994). Rather, the displacement changed irregularly with time, in a stepwise manner.

To quantitatively compare the measured displacement to that expected from a power law, we must first estimate the time elapsed from the start of inflation to the start of our measurements. We can use Equation 1 to make this estimate if we know $C_{1020}$, the depth to the 1070 °C isotherm. At the start of the experiment, $C_{1020}$ was ~25 cm, the vertical distance from the flow top to the incandescent rift. Using this value, Equation 1 yields an elapsed time of ~10 h. Another quantity we require is the flow height at the start of the experiment. This was ~60 cm. The photograph in Figure 7 shows the incandescent rift and flow top ~19 min after the experiment began.

The displacement measurements were then adjusted by adding 10 h and 60 cm to their times and magnitudes, respectively. A least-squares fit of these adjusted values to the power law $H = b t^n$ is shown in Figure 10A. Although the data depart from the power law curve in detail, they are crudely consistent with it. A longer time series, particularly one that began near the start of inflation, would have permitted a more definitive comparison. We conclude that the flow inflation probably did broadly track a power law, but that some process or processes caused departures from it.

The record is dominated by inflation, but periods of stagnation and deflation are also present. As is apparent in Figure 9, a number of brief episodes of impulsive inflation were recorded, with inflation rates of more than 10 cm/hr. Most inflation, however, was nonimpulsive, with lower rates.

The inflation and deflation rates were generally too low to be detected by visual observation. Visual observations were, however, useful for noting times and locations of lava breakouts within ~50 m of the measurement site. The time-lapse photographs captured the displacements, as well as breakouts, that occurred within ~1 m of the measurement point. The inflation was also faintly audible. As the rift propagated through the rift viscoelastic layer, it released pressurized gas contained within vesicles. Each vesicle release produced a faint popping sound audible only within meters to decimeters of the rift, depending on wind conditions.

Proximal lava breakouts, i.e., those that occurred at or within a few meters of the measurement site, are reflected in the displacement record and are depicted with green vertical bars in Figure 9. The first proximal breakout is included within Animations 2 and 3. The onsets of both proximal breakouts coincide with impulsive inflations with rates of >35 cm/hr. The coincidence of proximal breakouts and impulsive inflations suggests that these events occurred when the flow’s pressurized core liquid caused the rift viscoelastic layer to rupture. Resistance to inflation then dropped, the flow’s upper crust was pushed rapidly upward, and some core liquid escaped.

Most impulsive inflations occurred without an accompanying breakout (Fig. 9). In such cases, the rift apparently propagated to the liquid core, but the uplift of the flow’s upper crust reduced the fluid pressure so efficiently that only a small quantity of lava was extruded along the rift line—too little to form an obvious breakout.

Distal lava breakouts, i.e., those that occurred 15–20 m from the measurement site, are reflected in the displacement record as diminutions in the inflation rate. These are indicated in Figure 9 with the two blue bars. The first was associated with a reduced inflation rate. The second (Fig. 11) was significantly more voluminous than all other breakouts, both proximal and distal. It was associated with a stagnation and deflation period. We attribute these two inflation rate diminutions to pressure reduction in the flow’s core as escaping lava fed the breakouts.

Most stagnation and deflation episodes did not coincide with breakouts within our field of view, so we are uncertain what caused them. Breakouts were certainly occurring on the flow margin outside our field of view, and some of these may have been sufficiently voluminous to diminish or stop inflation over areas well away from the point of breakout, including our experiment site. However, this explanation does not work well for the longest stagnation and deflation episode, which occurred at night, from ~11:20 p.m. to ~3:20 a.m. (hours 12–16 in Fig. 9), because we did not observe the incandescence that a large breakout would produce. It is also possible that the supply of lava to the lower Banana flow was temporarily diminished.

Figure 4. Index map showing the area inundated by lava during Kilauea’s Pu’u ‘Ō’ō eruption between January 1983 and June 2004, and the place on the margin of the Banana flow where flow-height data were collected from an inflating sheet flow.

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during this interval, perhaps by a blockage or substantial breakout above Pālāma pāli. However, we have no evidence to support this possibility. We prefer an alternative explanation: that the rift viscoelastic layer achieved sufficient tensile strength to restrain the fluid pressure of the core liquid for an extended period of time. The shorter stagnation and deflation episodes in Figure 9 were probably also caused by tensile strengthening of the rift viscoelastic layer, by breakouts beyond our field of view, or by a combination of these two processes.

The flow margin gradually deflated 8 mm during the longest stagnation and deflation episode (Fig. 9). One could argue that withdrawal of liquid lava from the flow interior could produce deflation of the central part of the flow, but this explanation is unconvincing at the flow margin, which is buttressed by solidified lava. The gradual deflation is more consistent with cooling-induced contraction. Contraction is continuous, but only becomes apparent as a decrease in flow height when the contraction rate exceeds the inflation rate.
Rift Banding

Comparison of time-lapse photographs to the extensometer record permits us to test the proposition that rift banding is caused by changes in the inflation rate. If the proposition is true, we would expect band formation to be synchronous with abrupt inflation rate changes. As can be seen in the time-lapse photographs (Animations 2 and 3), the first rapid inflation was indeed synchronous with the formation of a band.

From the extensometer record, the first impulsive inflation began between 15:33:19 and 15:33:29, 4.24 h after the start of the experiment. This event is labeled inflation 1 in Figure 12, which is a subset of the complete extensometer record shown in Figure 9. The impulsive inflation begins between frames 15:33:06 and 15:33:36 in Animations 2 and 3. Inflation continues in the next frame, 30 s later, and an irregular band of lava begins to extrude along the rift line. The extrusion (extrusion band 1, Figs. 12 and 13A) appears as a dark band sandwiched between two incandescent bands. The extrusion is dark because it cooled more rapidly than the grooves above and below it. Inflation and band growth continue for the next 3 min. Pāhoehoe toes, extruding from the rift ~1 m from the band, accompany continuing inflation over the next 8 min (Animations 2 and 3, beginning at 15:37:07). Extrusion then stops; the extrusion band is bounded by incandescent grooves, both above and below. Inflation is discernible on the time-lapse photos until 20 min after the start of the event. Thereafter, until inflation 2 begins, inflation is evident on the extensometer record, but is indiscernible on the time-lapse record.

Correlating band formation with the first impulsive inflation is simple because the prop-
agating rift line is within the field of view of the time-lapse camera. However, the rift line propagated into the flow and upward, eventually passing out of the field of view of the time-lapse camera (Animations 2 and 3). Consequently, establishing synchronicity for subsequent bands is more complicated. As rifting progressed, the lower rift surface gradually came into the camera’s field of view. Thus, there was a time delay between formation of a band on the lower rift surface and its appearance in the time-lapse sequence.

Inflation 2 began ~1 h after inflation 1 (Fig. 12); ~4 min after inflation 2 began, a narrow, incandescent groove first became visible; ~7 min
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later, a dark band first became visible above the incandescent groove. The dark band and the groove beneath it are labeled extrusion band 2 and groove 2 in Figures 12 and 13B, respectively. Figure 13B is a photograph taken with a handheld camera 27 min after the start of inflation 2.

Inflation 3 began ~1 h and 11 min after inflation 2 (Fig. 12); ~20 min after inflation 3 began, another incandescent groove first came into view on the lower surface. This groove is labeled groove 3 in Figures 12 and 14A. Figure 14A is a time-lapse photograph taken ~70 min after the start of inflation 3. The quality of subsequent photographic data (collected after nightfall and the following morning) was inadequate to correlate succeeding inflation events with features on the rift surface.

The appearance of the lower rift surface at the end of the experiment, late in the morning of 13 May, is shown in Figure 14B. The upper

Figure 9. Vertical displacement (black) of the margin of a sheet-lava flow at 10 s intervals, and corresponding displacement rate (red), over a period of ~24 h, starting at 11:18:39 a.m. on 12 May 2004. The rate has been smoothed with a 5-point moving average filter. The noise in the rate record is due to wind. Vertical bars show times of minor extrusions: green indicates extrusion observed at the measurement site; blue indicates extrusion observed 15–20 m from measurement site. Events 1, 2, and 3 are impulsive inflation events discussed in the text (and documented in subsequent figures). Bracket shows data subset used in Figure 12.
The prominent extrusion band (extrusion band 1, Fig. 14B) unequivocally began to form at, or slightly after the start of, inflation 1 (Fig. 12). The band formed when the pressure of the core liquid exceeded the tensile strength of the rift viscoelastic layer, which then ruptured (Fig. 15A). The flow’s pressurized core fluid then pushed up the upper crust, increased the thickness of the liquid core, and forced some liquid lava through the rift (Fig. 15B). The extrusion and inflation rates declined as the fluid pressure declined. Extrusion continued until the declining fluid pressure became inadequate to overcome viscous resistance. Extrusion then stopped, the band quickly cooled, and a new rift viscoelastic layer bridged the gap formed during extrusion (Fig. 15C). The strength of the reestablished viscoelastic layer increased rapidly at first. Inflation continued at a reduced rate (Fig. 12) as the reestablished rift viscoelastic layer stretched in response to repressurization of the core liquid. Eventually the pressure of the core liquid exceeded the tensile strength of the rift viscoelastic layer, which again ruptured (Fig. 15D). The liquid core migrated upward and inward.

Some of extrusion band 1 adhered to the upper rift surface, but most of it clung to the lower surface. From examination of numerous rifts on the coastal plain, we find that this behavior is typical: most extrusion bands are found on only one side of the rift. Examples exhibiting extrusion bands split symmetrically down the middle or irregularly from side to side are present, but are less abundant.

The feature labeled extrusion band 2 in Figure 14B is the dark band (groove 2, Fig. 15B) that first became visible during inflation 2 (Fig. 12). Though its formation was not directly observed, we are confident that the dark band is an extrusion band. It was darker than the underlying incandescent groove 2 in Figure 14B because raised surfaces cool faster than indented surfaces. Extrusion band 2 formed in the same manner (Figs. 15D, 15E) as its predecessor; groove 2 is simply the recessed area at the lower front of extrusion band 2 (Fig. 15F).

The feature labeled groove 3 in Figure 14B is the incandescent band (groove 3, Fig. 14A) that first became visible after inflation 3 (Fig. 12). We interpret groove 3 as the recessed area at the lower front of extrusion band 3 (Fig. 14B).

Extrusion band 1 is vitreous, relatively smooth, and is olive-gray (Fig. 14B). Though less obvious, the same is true of extrusion bands 2 and 3. The color of the extrusion bands reflects the fact that they cooled rapidly and so did not have time to oxidize. Extrusion bands 1, 2, and 3 were produced during inflations 1, 2, and 3, respectively, as liquid lava was extruded into the rift each time it underwent impulsive inflation.
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The area between extrusion band 1 and groove 2 has a rough, spinose texture; the same is true of the area between extrusion band 2 and groove 3 (Fig. 14B). This texture was produced when the vesicular rift viscoelastic layer underwent ductile tearing during periods of sustained non-impulsive inflation. Its color is dominantly olive-gray but locally grades into a red-pink color. The red-pink areas evidently cooled slowly enough to undergo high-temperature oxidation.

Cyclicity?

Some inflation-rift surfaces on the coastal plain contain numerous rift bands whose spacing and general appearances are remarkably similar. This almost certainly is due to a cyclic process. The stepwise pattern evident in our inflation data (Fig. 9) also suggests episodic or crude cyclic behavior. It is difficult to conceive, however, of a recharge mechanism capable of monotonously producing the virtually identical bands seen on some rift surfaces.

Inflating sheet flows have continuous liquid cores that act as conduits for lava transport (Hon et al., 1994; Kauahikaua et al., 1998). Crust growth continuously reduces the cross-sectional area of a conduit’s liquid core, continuously reducing its volume and output. Assuming that the lava supply rate is approximately constant, lava input will exceed output, and so lava backs up within the conduit upslope; thus, the pressure in the fluid core increases. The increased pressure pushes the flow’s upper crust upward, widening the liquid-filled gap between the upper and lower crusts. As the gap widens, the output increases and pressure decreases. If the system were always in perfect equilibrium, crust growth and the consequent reduction in the liquid cross section would be instantaneously balanced to keep lava input equal to lava output. Perfect equilibrium is implicit in Equation 5, which models flow inflation as a function of time.

Figure 10A shows a least-squares fit of Equation 5 (blue line) to our experimental data (black line). The stepwise deviations from ideality documented in Figure 10A show that the inflating flow we studied did not maintain perfect equilibrium. The deviations were recorded as bands on the surface of the inflation rift. Given that banding is ubiquitous in sheet flows, we conclude that deviations from an ideal steady state are an intrinsic feature of flow inflation.

The stepwise deviations are probably caused by the competition between the fluid pressure in the flow’s liquid core and the tensile strength of the viscoelastic layer where it is weakest, along the rift line. The pressure of the core liquid exerts an outward-directed force, while the tensile strength of the rift viscoelastic layer resists that force. The brittle crust does not contribute significantly to the tensile strength of the crust because it contains a network of shrinkage cracks that develop normal to the cooling surface.

Strength of Viscoelastic Crust

The strength of the viscoelastic crust is related to its thickness, $C_{\text{vis}}$. When liquid lava is exposed to air, $C_{\text{vis}}$ will initially grow accord-
The rate of growth of $C_{10^7}$ is the derivative of Equation 1:
\[ \frac{dC_{10^7}}{dt} = 0.0390 \, \text{m/hr}, \] (6)
where $dC_{10^7}/dt$ is in m/hr, and $t$ is in hours.

The rate quickly decelerates with time and abruptly declines at $t = 0.243$ h when brittle crust begins to form (Fig. 16). After brittle crust formation begins, the rate of change of the viscoelastic crust is given by the derivative of Equation 3:
\[ \frac{d(C_{10^7} - C_{400})}{dt} = 0.0153 \, \text{m/hr}, \] (7)

Figure 16 applies to viscoelastic crust formation on the flow top and, with a 30% reduction, to the flow bottom. Because of uninterrupted growth, the non-rift viscoelastic layer in the upper and lower crusts is thicker, and thus stronger, than in the rift viscoelastic layer.

Crust formation is more complex within a rift because the viscoelastic layer is repeatedly disrupted as it stretches and ruptures in response to pressurization of the liquid core. This exposes hotter lava, which increases the rate of heat loss and the rate of viscoelastic layer formation. Consequently the 1070 °C isotherm moves inward discontinuously and more rapidly in the flow margins than in the upper and lower crusts.

For viscoelastic materials, the relationship between stress and strain depends on time (Lakes, 2009). When subjected to a constant stress, viscoelastic materials undergo temperature-dependent and time-dependent deformation known as creep. Creep is most pronounced at temperatures near a material’s melting point. Plots of strain as a function of time at constant stress may exhibit three creep stages (Fig. 16B): primary, in which the curve is concave downward; secondary, in which strain increases linearly with time; and tertiary, in which strain accelerates to rupture (Lakes, 2009).

Parts of the measured inflation record mimic this pattern (e.g., inflation 1, Fig. 12). At least part of the resemblance may be fortuitous. The initial concave-downward part of inflation 1 corresponds to rapid uplift consequent to rupture of the viscoelastic layer. If primary creep occurred within the rift viscoelastic layer, it is not distinguishable from the rupture-triggered uplift. Secondary creep of the rift viscoelastic layer may have been responsible for the approximately linear part of inflation 1, and tertiary creep may have been responsible for the strain acceleration that preceded rupture (inflation 2, Fig. 12).

**Excess Pressure**

Excess pressure $P$ is the pressure available to lift the flow’s upper crust. It is given approximately by:
\[ P = \rho gh, \] (8)
where $\rho$ = density, $g$ = gravitational acceleration (9.8 m/s$^2$), and $h$ = elevation difference between the top of the flow and the top of the lava column upslope. Frictional losses, proportional to flow velocity, will reduce $P$ slightly. For the Banana flow experiment, the lava column top is probably at the base of Pālāma pali (Fig. 4). We estimate: $\rho = 2 \times 10^3$ kg/m$^3$, and $h = 20$ m. Therefore the excess pressure in the flow’s liquid core at the experiment site probably varied about a value of ~0.4 mPa.

Figure 13. (A) Photograph showing the rift 27 min after the start of inflation 1 (Fig. 12). Extrusion band 1 (the dark band sandwiched between two incandescent bands) was extruded during a 3 min period beginning ~0.5 min after inflation 1 began. (B) Photograph showing the lower rift surface ~27 min after the start of inflation 2 (Fig. 12). Extrusion band 2 (the dark red band) first became visible ~12 min after the start of inflation 2. Groove 2 (the incandescent band beneath extrusion band 2) first became visible ~5 min after the start of inflation 2. The rift line is partly visible as the bright yellow areas above extrusion band 2.

**Figure 16** applies to viscoelastic crust formation on the flow top and, with a 30% reduction, to the flow bottom. Because of uninterrupted growth, the non-rift viscoelastic layer in the upper and lower crusts is thicker, and thus stronger, than in the rift viscoelastic layer.
Thickness of the Liquid Core

The two red lines in Figure 10A show the evolution of the upper and lower crusts, calculated from Equation 1. The equilibrium thickness of the liquid core at a specified time is the difference between the red lines. This difference, calculated by subtracting the sum of the upper and lower crust thicknesses (Equation 2) from the measured flow height. The maximum nonequilibrium deviation from the equilibrium liquid core thickness in Figure 10B is ±4 cm, or ±0.2 of the equilibrium value.

Pressurization Rate

On the basis of mass conservation, we assume that \( \frac{dP}{dt} \), the pressurization rate of the flow's liquid core, is directly proportional to the difference between the volumetric lava supply rate and the volumetric rate at which lava is expended in inflation and breakouts:

\[
\frac{dP}{dt} = K(Q_0 - Q),
\]

where \( K \) is the bulk modulus, \( Q_0 \) is the volumetric lava supply rate, and \( Q \) is the volumetric rate at which lava is expended in flow inflation and breakouts.

The volumetric flux difference decreases if \( Q_0 < Q \). This probably occurs most commonly when upward displacement of the upper crust increases the core thickness (core volume gain). Breakouts may accompany upward crust displacement. Lava expended in inflation and breakouts causes the elevation of accumulated lava in the conduit upstream to drop, thereby decreasing the excess pressure. During the experiment, the flow was in this state through the upward-sloping intervals on the plot of nonequilibrium core thickness (black line, Fig. 10B).

The volumetric flux difference increases if \( Q_0 > Q \), a circumstance that probably occurs most commonly when there are no lava breakouts or when crust growth reduces the thickness and volume of the core liquid. Consequently, liquid lava backs up in the conduit upstream, increasing the excess pressure. During the experiment, the flow was in this state through the downward-sloping intervals on the plot of nonequilibrium core thickness (black line, Fig. 10B).

Hypothetical Cycle

We propose that cycles are a consequence of two opposing forces: (1) the excess pressure of the core liquid, which pushes on the upper crust, and (2) the tensile strength of the rift viscoelastic layer, which resists the excess pressure.

The excess pressure reaches its zenith at the point at which it exceeds the rupture strength of the rift viscoelastic layer, causing the tensile strength to drop to nil (Fig. 17A). At this point, the thickness of the core liquid rapidly increases (Fig. 17B) as the flow top is displaced upward (Fig. 17C) and some lava is extruded along the rift. The increase in the core liquid thickness increases \( Q \), which decreases \( Q_0 - Q \) and \( \frac{dP}{dt} \).
The elevation of lava that had backed up within the conduit upslope begins to drop, as does the excess pressure. The tensile strength of the rift viscoelastic layer remains at nil until the inflation rate slows sufficiently for the 1070 °C isotherm to begin to diffuse into the core liquid exposed along the rift line. This occurs (Fig. 17A) when the core thickness approaches the equilibrium value (Fig. 17B). The growth rate of the tensile strength of the rift viscoelastic layer is greatest when the excess pressure reaches its nadir (Fig. 17A), the core thickness reaches the equilibrium value (Fig. 17B), and the inflation and rift-line propagation rates become linear as secondary creep begins (Fig. 17C). At this point, \( Q_o \) equals \( Q_e \), so \( dP/dt \) is nil.

The thickness and tensile strength of the rift viscoelastic layer now grow at an ever-diminishing rate (Fig. 17A) as the 1070 °C isotherm diffuses inward. Growth of the upper and lower crusts reduces the thickness of the liquid core. The decreasing core thickness decreases \( Q_e \), which increases \( Q_o - Q_e \), \( dP/dt \), and excess pressure as lava backs within the conduit upslope. At first, the tensile strength grows faster than the excess pressure, so the difference between them increases. However, the growth rate of the rift viscoelastic layer decelerates while the pressurization rate accelerates, so eventually the difference between the tensile strength and excess pressure begins to diminish. Subsequently, necking of the rift viscoelastic layer (tertiary creep) causes the tensile strength to decline (Fig. 17A) and the core thickness to increase (Fig. 17B) as inflation accelerates (Fig. 17C). Ultimately, the excess pressure exceeds the tensile strength of the rift viscoelastic layer, which undergoes ductile rupture. The cycle is then repeated.

The key parameter in this model is the tensile strength of the rift viscoelastic layer because, in its absence, deviations from equilibrium, and consequent rift banding, would not occur.

Inflation History Uniqueness

If we had collected data simultaneously at other points along the flow margin, would we have observed the same inflation changes that we recorded at a single site? Although we cannot answer this question definitively, we offer the following.

The core liquid of the entire flow was probably interconnected. Evidence for this is the relatively uniform uplift of the upper crust that characterizes inflating sheet flows. Because of fluid interconnection, pressure perturbations at one point on the flow margin would eventually propagate throughout the core liquid. The core liquid was continuously pressurized by lava influx. The influx rate during the experiment is unknown, but Kīlauea’s typical lava tube flux, sans vesicles, is ~3.5 m³/s (Kauahikaua et al., 1996). The continuous pressurization was interrupted by depressurization events, i.e., breakouts along the flow margin and uplift of flow’s upper crust.

The depressurizations caused by the two breakouts that occurred within 15–20 m of the experiment site clearly caused a local inflation diminution. Despite the fluid interconnection, an extensometer located far from the breakout points probably would have recorded little or no inflation diminution. The lava volume lost in the largest of the two (Fig. 11) was probably less than a few cubic meters, a tiny fraction of the total flow volume. The volume lost on the flow margin probably would have been replaced at the head of the flow in 1 s or less. Depressurizations from the small breakouts probably were damped quickly with distance from the breakout.

The elevation of lava that had backed up within the conduit upslope begins to drop, as does the excess pressure. The tensile strength of the rift viscoelastic layer remains at nil until the inflation rate slows sufficiently for the 1070 °C isotherm to begin to diffuse into the core liquid exposed along the rift line. This occurs (Fig. 17A) when the core thickness approaches the equilibrium value (Fig. 17B). The growth rate of the tensile strength of the rift viscoelastic layer is greatest when the excess pressure reaches its nadir (Fig. 17A), the core thickness reaches the equilibrium value (Fig. 17B), and the inflation and rift-line propagation rates become linear as secondary creep begins (Fig. 17C). At this point, \( Q_o \) equals \( Q_e \), so \( dP/dt \) is nil.

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Inflation rates of a pāhoehoe sheet flow

SUMMARY AND CONCLUSIONS

1. The change in the height of the flow margin observed during the experiment was crudely consistent with the power law that Hon et al. (1994) fit to inflating flow data sets. The power-law model implies perfect equilibrium, with lava influx instantaneously balanced by flow inflation.

2. Our data exhibit deviations from the power law that indicate that some shorter time-scale process or processes are superimposed on the power-law model.

3. The record is dominated by inflation but periods of stagnation and deflation are also present.

4. Proximal lava extrusions, i.e., those that emanated from the inflation rift at or within a few meters of the measurement site, coincided with impulsive inflations with rates of >35 cm/hr.

5. Distal lava extrusions, i.e., those that occurred 15–20 m from the measurement site, were associated with diminution in the measured inflation rate.

6. Inflation occurred when pressure in the fluid core of the flow was sufficient to lift the weight of the upper flow crust and to exceed the tensile strength of the rift viscoelastic layer.

7. Periods of stagnation were probably caused either by core pressure loss due to lava breakouts or tensile strengthening of the rift viscoelastic layer.

8. Deflation was probably caused by cooling. Cooling-induced deflation was continuous, but was apparent only when its rate exceeded the rate of inflation.

9. Comparison of extensometer data and time-lapse photographic data indicates that rift banding is correlated with changes in the inflation rate. This confirms the hypothesis that motivated us to perform the experiment.

10. Rift banding is ubiquitous on the margins of sheet flows, so inflation rate changes are probably an intrinsic feature of sheet flow inflation.

11. The mechanism that produces rift banding is probably cyclic.

12. We propose a conceptual model wherein the inferred cyclicity is due to the competition between the excess pressure in the flow’s liquid points, so inflation diminutions probably did not extend far from the experiment site.

Depressurizations caused by larger breakouts probably affected inflation over proportionately larger areas of the flow. Large sustained breakouts or uplift of large areas of the flow’s upper crust probably affected inflation over the entire flow. Depending on the propagation rate, a perturbation might arrive at different parts of the flow at substantially different times.

Consequently, we speculate that the inflation history at a given point is the product of pressure perturbations whose sizes and areas of influence varied widely. If this is true, the correlation between simultaneously measured inflation histories along the flow margin should generally decrease with the distance between the measurement sites.

Figure 16. (A) Graph showing the growth in the thickness of the viscoelastic layer (red) for the first hour after liquid lava is exposed to air, calculated from Equation 1 (below \( t = 0.243 \) h) and Equation 3 (above \( t = 0.243 \) h). The corresponding growth rate (black) was calculated from Equation 6 (below \( t = 0.243 \) h) and Equation 7 (above \( t = 0.243 \) h). (B) Graph of deformation as a function of time for a viscoelastic material placed under a constant load at \( \varepsilon_0 \); \( \varepsilon_0 \) is elastic strain. Strain above \( \varepsilon_0 \) is plastic strain, or creep, which is divided into three stages: primary, secondary, and tertiary. Tertiary creep culminates in ductile rupture.
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ACKNOWLEDGMENTS

Olivier Lengliné and Kevin Stuart volunteered to help carry equipment to the site of the experiment. We—especially our backs—are grateful. We wish to thank Taeko Jane Takahashi for carefully scrutinizing the manuscript for style, consistency, and grammatical errors. We are grateful to Matthew R. Patrick, Scott K. Rowland, William W. Chadwick, Jr., and Samuel A. Soule for thorough and insightful reviews.

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Figure 17. Graphs showing variation of parameters with time in a hypothetical flow-inflation cycle. Functional forms are conceptual. (A) Excess pressure (red) and tensile strength of the rift viscoelastic layer (blue). (B) Thickness of the infl ating flow’s core liquid (red); $Q_0 = $ volumetric lava supply rate; $Q = $ volumetric rate at which lava is expended in flow inflation and breakouts; see text. (C) Flow height (black).
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MANUSCRIPT RECEIVED 23 NOVEMBER 2010
REVISED MANUSCRIPT RECEIVED 18 OCTOBER 2011
MANUSCRIPT ACCEPTED 20 OCTOBER 2011

Geosphere, February 2012