Thermal Efficiency of Lava Tubes in the Pu‘u ‘Ō‘ō-Kūpaianaha Eruption

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Abstract

We have applied glass geothermometry to a suite of 37 pairs of glassy lava samples collected on Kīlauea's east rift zone, from the upper and lower ends of the episode 48 lava-tube system, when the Kūpaianaha pond was active (1986–90). We also present data on a small suite of skylight samples collected from 1986 through the end of episode 48 (in 1991), plus some data on skylight-coast sample pairs from episodes 51 and 53. The results for the pond-coast pairs are as follows:

(1) From November 1986 through January 1988 (15 months), the average change in inferred quenching temperature from pond to coast (for 12 sample pairs) is 12.4°C, and the average increase in crystal content (inferred from the observed enrichment of TiO_2 and K_2O in glasses in coastal samples) is 11 to 12 weight percent.
(2) From February 1988 through November 1989 (23 months), the average change in inferred quenching temperature from pond to coast (for 25 sample pairs) is 8.4°C, and the average increase in crystal content is 4 to 5 weight percent. Within this part of the data set, pond and coastal temperatures rise and fall together much of the time, even though these temporal fluctuations are at or below the limit of resolution of glass geothermometry $\Delta T \leq 3$°C).
(3) The minimum temperature difference for any pond-coast pair is 7°C. Of 37 sample pairs, 24 have ΔT=7–9°C, over the entire 3-year period.

In about half of the skylight samples, the enclosed glasses have MgO contents consistent with the location at which the samples were collected along the lava-tube system. Glasses in the rest of the skylight samples are displaced to lower MgO contents; such samples may not be consistently as well quenched as the pond and littoral spatter samples. The data from episodes 51 and 53 are from a new tube system that was somewhat shorter than the episode 48 (Kūpaianaha) lava tubes. The best-documented temperature difference observed for this 10-km long lava tube (6°C) gives exactly the same rate of temperature decrease with distance (0.6°/km) as the limiting ΔT value of 7°C observed for the 12-km-long Kūpaianaha lava-tube systems. This cooling rate may represent the limiting thermal efficiency of tubes of the current eruption.

Introduction

Observations of the ongoing long-term eruption of Kīlauea Volcano on the Island of Hawai‘i have enhanced our understanding of many eruptive processes at Kīlauea, in particular, how lava tubes form and function. Lava tubes are common in pāhoehoe flows at both Kīlauea and Mauna Loa (Greeley, 1987), where their formation is favored by low to moderate effusion rates (Rowland and Walker, 1990). In the current eruption, lava tubes have been the dominant means of lava transport from July 1986 to the present.

Glass geothermometry (Helz and Thornber, 1987), which derives quenching temperatures of samples from the MgO content of their glasses, has permitted tracking the thermal history of this long-term eruption in detail. Glass analyses and estimated quenching temperatures are well documented for the period 1983–94 (Helz and others, 1991; Helz and Hearn, 1998); the same technique has also been applied to episodes 49 (Mangan and others, 1995), 50 to 53 (Heliker and others, 1998) and 53 to 55 (Thornber, 2001). In this chapter, we focus on a subset of the data that documents the thermal efficiency of lava tubes from 1986 to 1990, the period when the active lava pond at Kūpaianaha was accessible for sampling.

Background

The early stages of the current Kīlauea eruption were characterized by episodic high lava fountains at the Pu‘u ‘Ō‘ō vent, which fed rapidly moving ‘a‘ā flows. After July 1986, this eruptive style gave way to more continuous effusive activity at a new vent 3 km downrift from Pu‘u ‘Ō‘ō. From July 1986 to mid-1990, lava surfaced at this site, where it formed a large open pond of circulating lava, named the Kūpaianaha vent. The active lava pond gradually crusted over, roofing over completely in late 1990. Eruptive activity continued, however, with lava moving through the site of the former Kūpaianaha pond to the coast by way of well-established lava-tube systems (Heliker and Wright, 1991; see Heliker and Mattox, this volume). The resulting lava aprons of tube-fed pāhoehoe and the mechanism of their emplacement were described in some detail by Mattox and others (1993) and Hon and others (1994).

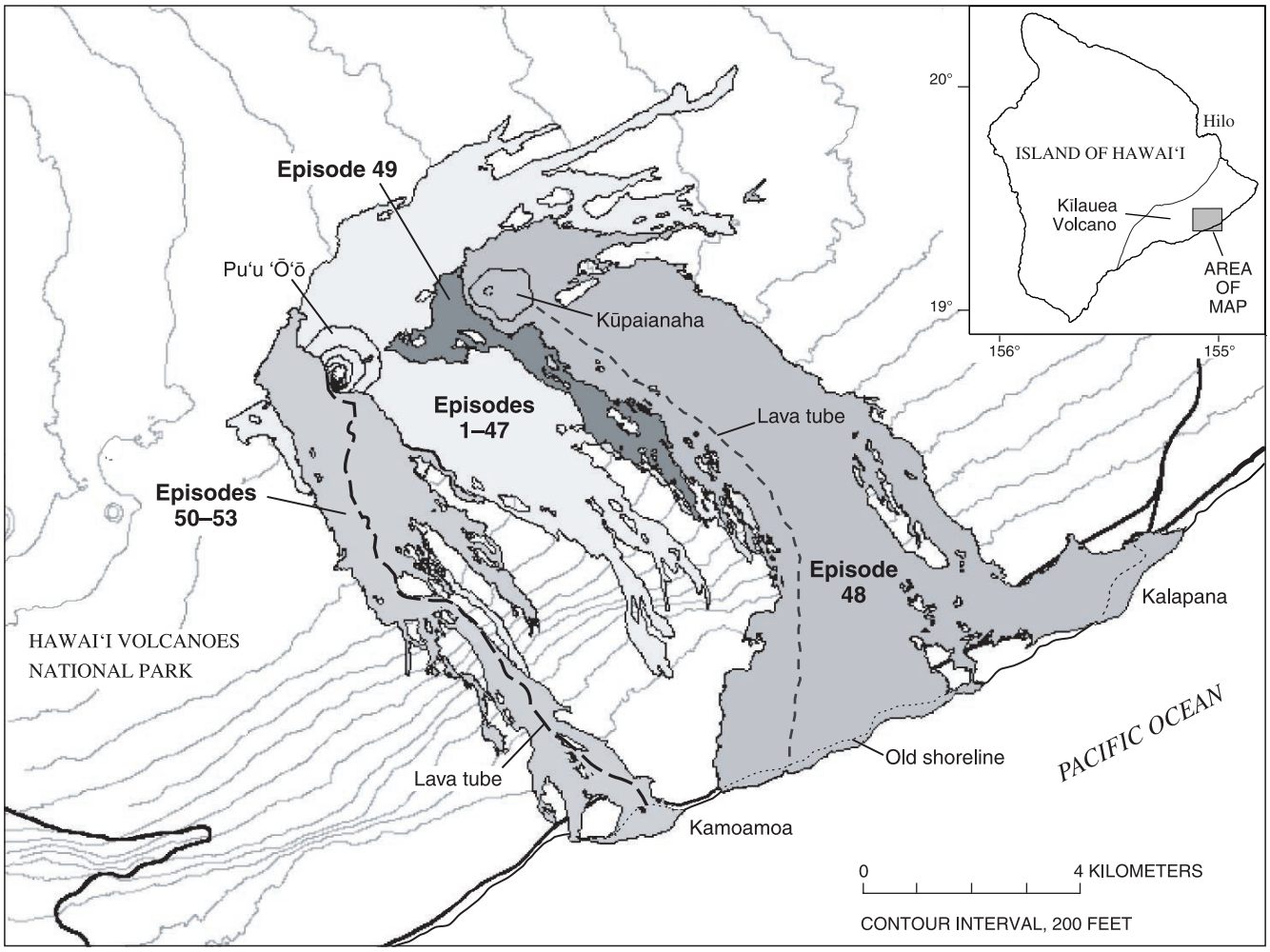

Figure 1. Kīlauea Volcano, Island of Hawai‘i, showing east rift zone through early stages of episode 53 (April 1993). Light dashed line, one of several lava-tube paths of Kūpaianaha period; heavy dashed line, episode 53 lava-tube system.

The Kūpaianaha vent, the extent of the episode 48 lava fields, and the episode 49–53 lava flows are shown in figure 1. Sample lava-tube paths are shown for episodes 48 and 53; these paths are not unique but indicate typical paths of the tube systems studied by Cashman and others (1994) and those discussed here. Scientists at the U.S. Geological Survey's Hawaiian Volcano Observatory (HVO) have measured these lava-tube systems to be 10 to 12 km long (Mattox and others, 1993; Cashman and others, 1994; Kauahikaua and others, 1998; also HVO, unpub. data).

Sample Characteristics

When the Kūpaianaha lava pond was open, and active tube systems delivered lava to the coast, HVO scientists made a special effort to collect paired samples of spatter from both the pond (at the upper end of the lava-tube system) and the coast, preferably of ocean-quenched littoral spatter. We report data on the existing set of 37 such sample pairs. The pairs were usually collected within a day of each other, although some pairs were collected 2 to 5 days apart (see table 1). We also report data for a few samples from the episode 48 suite that were collected through skylights in active lava tubes during the period when the Kūpaianaha lava pond was open (table 2) or after the pond closed up (table 3).

Pond Samples

The pond samples include spatter samples, one sample of pond overflow (KER48-668F) and samples of pond dip. The spatter and overflow samples date from November 1986 through March 1988, when the level of lava in Kūpaianaha was high, and circulation of the lava caused frequent episodes of spattering along the rim. When collected in real time, or when the spatter fell on a relatively cool surface, these samples underwent a rapid quench, making them suitable for geothermometry. All of the pond samples collected from April 1988 through May 1990 (after which the lava pond crusted over completely) are so-called dip samples; that is, they consist of glass that adhered to a rod or hammer thrown into the molten lava, or lowered from a helicopter into the lava. This

sampling method provides a very good quench, so long as the probe is retrieved from the molten lava quickly.

Coast (Littoral) Samples

The early (Nov. 1986–Jan. 1988) coastal samples include those collected where lava flows entered the ocean, plus some material from lava-tube breakouts near the coast. The quality of the quench for individual breakout samples depends on how quickly the sample was collected and whether the collector used water to quench it, or relied on air cooling. After January 1988, all coastal samples in the suite are of spatter from littoral cones or from lava tubes at the shoreline and were well quenched by interaction with seawater. Because the quenching history of the samples depends on where and how they were collected, such details are given in table 1 for all littoral samples.

Skylight Samples

Skylight samples were rarely taken when the Kūpaianaha lava pond was open; the two samples we have (KER48-852F and KER48-875F) were both collected on the same day as their associated pond samples (table 2). In addition, six skylight samples were collected in 1990 during the waning stages of the pond's life (table 2). Finally, we have analyzed several sets of skylight-littoral samples collected after the pond had completely crusted over (table 3). All the skylight samples were collected on a hammer lowered into the lava stream. The surfaces of the walls and roofs within lava tubes are known to be at basalt solidus temperatures (~985–1,000°C; Helz and Thornber, 1987) or higher. Kauahikaua and others (1998) reported a temperature of 1,080°C for the walls of a lava tube. Even the air temperatures over skylights are typically 800°C or higher (Peterson and others, 1994), so these samples may have undergone a different pattern of quenching from most other samples, with rapid cooling not occurring until after the hammer had cleared the skylight.

Petrography of Samples Used in This Study

All of our lava samples consist largely of clear brown glass, with minor amounts of crystals and vesicles. The photomicrographs in figure 2 show two pond-coast pairs that are representative of the whole population. In the first pair, pond sample 961P contains minor amounts of olivine (+ chromite, generally included in olivine), plus tiny crystals of augite. The coastal sample is somewhat more crystalline and contains trace amounts of plagioclase, as well as olivine + augite. In the second pair, both pond and coastal samples contain olivine + augite + plagioclase; again, the coastal sample is more crystalline than the pond sample. Sample 1076S (table 1) is

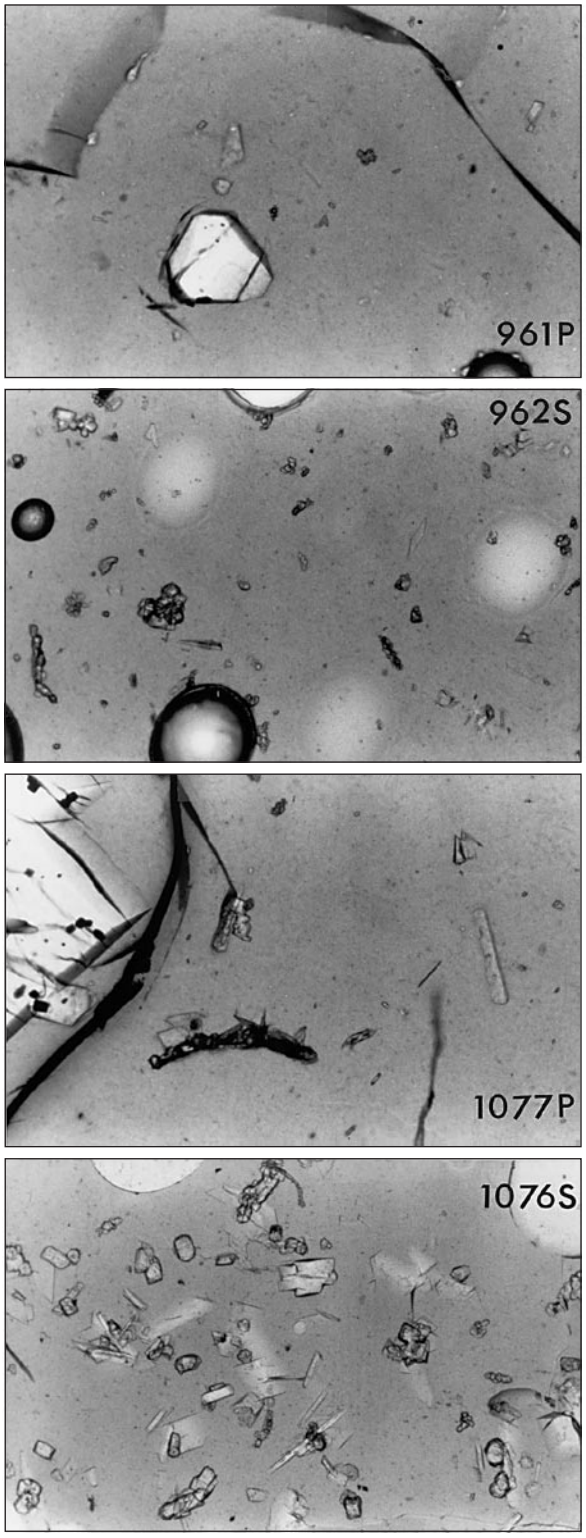

Figure 2. Photomicrographs (plane polarized light) of two pond-coast sample pairs of spatter from Kūpaianaha. Sample pair *A* and *B*; (samples 961P and 962S, table 1) were erupted in September 1988, and *C* and *D* (samples 1077P and 1076S, table 1) in May 1989. Field of view is 1.7 mm across for all images. Crystals present include olivine (+ chromite as inclusions), augite, and plagioclase; rest of area is glass and vesicles. Crystal content of coastal samples is higher and more variable than that of pond samples.

Table 1. Details on paired samples of spatter collected from the Kūpaianaha lava pond and the coast.

Pond sample	Date collected	T_{MgO} °C	Littoral sample	Date collected	T_{MgO} °C	Time difference	ΔT_{MgO} °C	Location of littoral sample	Bulk analysis for samples?
663S	11/25/86	1,148	666F	11/29/86	1,139	4 days	9	Beach	yes for both
668F	12/05/86	1,156	674F	12/04/86	1,143	1 day	14	Under water	"
730S	04/17/87	1,156	739F	04/15/87	1,143	2 days,	13	Tidepool	"
747S	05/12/87	1,156	746F	05/12/87	1,147	same day	9	2-m elevation	"
749S	05/19/87	1,157	748F	05/19/87	1,148	same day	9	350 m from ocean	"
764S	06/17/87	1,155	759F	06/14/87	1,144	3 days	11	3-m elevation	not for 759F
774P	07/07/87	1,157	776F	07/02/87	1,143	5 days	14	9-m elevation	yes for both
790S	09/09/87	1,153	791F	09/10/87	1,140	1 day	13	150 m from ocean	"
822S	11/03/87	1,154	817F	11/04/87	1,139	1 day	15	50 m from ocean	"
853P	12/17/87	1,152	860F	12/17/87	1,135	same day	17	6-m elevation	"
868P	01/14/88	1,153	867F	01/13/88	1,137	1 day	16	20–30 m from ocean	"
874S	01/26/88	1,153	873F	01/25/88	1,146	1 day	7	9-m elevation	"
889S	02/08/88	1,151	891S	02/11/88	1,142	3 days	11	Littoral cone	"
894S	02/15/88	1,152	891S	02/11/88	1,142	5 days	10	Same sample	"
898S	02/25/88	1,157	897S	02/24/88	1,149	1 day	8	0-m elevation	"
902S	03/10/88	1,154	901S	03/11/88	1,147	1 day	7	0-m elevation	"
903S	03/15/88	1,154	904S	03/18/88	1,148	3 days	7	0-m elevation	"
905S	03/24/88	1,156	906S	03/26/88	1,147	2 days	8	0-m elevation	"
911P	04/21/88	1,157	912S	04/22/88	1,147	1 day	9	0-m elevation	"
937P	06/23/88	1,156	935S	06/22/88	1,147	1 day	9	0-m elevation	"
939P	07/07/88	1,156	940S	07/08/88	1,147	1 day	9	0-m elevation	not for 940S
948P	07/22/88	1,156	943S	07/18/88	1,148	4 days	8	0-m elevation	yes for both
956P	08/24/88	1,156	957S	08/28/88	1,149	4 days	7	0-m elevation	"
961P	09/20/88	1,157	962S	09/23/88	1,149	3 days	8	0-m elevation	"
983P	10/06/88	1,157	982S	10/03/88	1,149	3 days	8	0-m elevation	"
990P	11/17/88	1,155	991S	11/19/88	1,146	2 days	9	0-m elevation	"
1000P	01/04/89	1,156	1001S	01/05/89	1,149	1 day	7	0-m elevation	"
1003P	01/19/89	1,157	1004S	01/20/89	1,149	1 day	8	0-m elevation	not for 1004S
1016P	02/17/89	1,157	1014S	02/17/89	1,146	same day	9	0-m elevation	yes for both
1027P	03/09/89	1,158	1026S	03/07/89	1,148	2 days	10	0-m elevation	not for 1026S
1033P	04/06/89	1,156	1034S	04/07/89	1,149	1 day	7	0-m elevation	not for 1034S
1072P	05/05/89	1,157	1073S	05/06/89	1,149	1 day	8	0-m elevation	yes for both
1074P	05/11/89	1,156	1075S	05/11/89	1,148	same day	8	0-m elevation	for neither
1077P	05/19/89	1,155	1076S	05/19/89	1,144	same day	11	0-m elevation	not for 1076S
1078P	05/24/89	1,156	1079F	05/25/89	1,148	1 day	8	Floating pillow	not for 1079S
1087P	08/10/89	1,155	1090S	08/12/89	1,148	2 days	7	0-m elevation	not for 1090S
1107P	11/14/89	1,157	1106S	11/12/89	1,150	2 days	7	0-m elevation	yes for both

one of the coolest late littoral samples in this suite and so is more crystalline than most. The crystal contents of all samples are too low to be readily determined by point-counting the area of one thin section, and so we made no attempt to obtain modes by this method. The principal variations in the entire suite are as follows: (1) some pond samples have been quenched from above plagioclase-in, whereas all the coastal and skylight samples contain three silicate phases (documented in Helz and Hearn, 1998), and (2) all the coastal and skylight samples contain more crystals than their corresponding pond samples.

Glass Analyses and Geothermometry

All the glass analytical data discussed here were obtained at the U.S. Geological Survey laboratory in Reston, Va., using the ARL-SEMQ and JEOL electron microprobes. The actual analyses, further details on analytical techniques, and estimates of glass quenching temperatures were reported by Helz and Hearn (1998). Critical information on pond-coast sample pairs and skylight samples, extracted

from their report and from HVO archives, is summarized in tables 1 through 3.

Glass geothermometry, as developed by Helz and Thornber (1987), depends on the observation that the MgO content of glass varies linearly with temperature for Kīlauea melts in equilibrium with olivine. Thus, the absence of any sign of reaction between olivine and glass in these samples (fig. 2) is important. The reproducibility of the MgO contents of naturally quenched glasses is also critical, because the precision of the MgO analysis determines the limit of resolution of temperature of glass (T_{MgO}) geothermometry. Helz and others (1995, fig. 3) show that the difference between initial and replicate MgO analyses in a large suite of glasses from Kīlauea never exceeds 0.15 weight percent absolute, which corresponds to a 3°C difference in apparent quenching temperature. Thus, within this data set, any temperature difference (ΔT) larger than 3°C is outside the range of analytical noise and may be interpretable.

Results for Pond-Coast Sample Pairs

The MgO contents and quenching temperatures of glasses from all pond-coast sample pairs are plotted against time in figure 3. The most conspicuous feature of this plot is that, without exception, the glasses in the coastal samples have lower MgO contents and, hence, were quenched from lower temperatures than the glasses in the pond samples. This signal is so consistent that it is reasonable to interpret the temperature decrease as due to cooling during transport in the 10 to 12-km-long lava-tube system.

Except for the single earliest pond sample (663S, table 1, containing glass with 6.76 weight percent MgO), all of the pond samples discussed here contain glasses having MgO contents between 6.86–7.19 weight percent. The range in MgO content for all pond glasses (when we include those

Table 2. Data on samples from the episode 48 suite that were collected through skylights in active lava tubes during the period when the Kūpaianaha lava pond was open.

Pond sample	Date collected	T_{MgO} °C	Skylight sample	Date collected	T_{MgO} °C	Elevation (ft)	Littoral sample	Date collected	T_{MgO} °C	Time difference	ΔT_{MgO} (°C)
822S	11/03/87	1,154	--	--	--	--	817F	01/14/87	1,139	1 day	15
845S	11/19/87	1,154	852F	11/19/87	1,147	1,900	--	--	--	same day	7
853P	12/17/87	1,152	--	--	--	--	860F	12/17/87	1,135	same day	17
868S	01/14/88	1,153	--	--	--	--	867F	01/13/88	1,137	1 day	16
874S	01/26/88	1,153	875F	01/26/88	1,145	1,980	--	--	--	same day	8
"	01/26/88	1,153	--	--	--	--	873F	01/25/88	1,146	1 day	7
--	--	--	875F	01/26/88	1,145	1,980	"	01/25/88	1,146	1 day	-1
1129P	02/07/90	1,155	1137F	02/10/90	1,152	1,830	--	--	--	3 days	3
1234P	02/15/90	1,158	1150F	02/15/90	1,147	1,420	--	--	--	same day	12
--	--	--	1202F	03/19/90	1,147	1,650	1220S	03/19/90	1,145	same day	2
--	--	--	1237F	04/08/90	1,154	1,650	--	--	--	--	--
1252P	05/10/90	1,158	--	--	--	--	--	--	--	--	--
1258P	05/31/90	1,156	1257F	05/31/90	1,150	1,850	--	--	--	same day	6
--	--	--	1260F	06/30/90	1,152	1,850	--	--	--	--	--
--	--	--	--	--	--	--	1261S	07/28/90	1,146	--	--

Table 3. Data on samples from the episode 48 suite that were collected through skylights in active lava tubes after the Kūpaianaha lava pond closed up.

Episode	Skylight sample	Date collected	T_{MgO} °C	Elevation (ft)	Littoral sample	Date collected	T_{MgO} °C	Time difference	ΔT_{MgO} °C
48	1305F	02/27/91	1,154	1,845	--	--	--	2 days (1311S)	2
48	1306F	02/27/91	1,155	1,840	--	--	--	2 days (1311S)	3
48	1308F	03/01/91	1,156	2,080	1311S	03/01/92	1,151–1,152	same day	4-5
51	1407F	11/12/92	1,153	2,350	1408S	11/19/92	1,149	5 days	4
53	1430F	04/16/93	1,151	2,010	--	--	--	13 days (1436F)	0
53	1431F	04/16/93	1,156–1,157	2,360	--	--	--	13 days (1436F)	5-6
53	1432F	04/16/93	1155	2,350	--	--	--	13 days (1436F)	4
53	1433F	04/16/93	1151	250	1436F	04/29/93	1,151	13 days	0

without corresponding coastal pairs) is larger (6.76–7.36 weight percent MgO; Helz and Hearn, 1998). The narrow range in MgO content for pond glasses in the suite of paired samples considered here is important because it means that the variable time gaps between collection of samples considered to be pairs, whether a few hours or several days (see table 1), do not contribute significantly to the variation in ΔT values. Figure 3 shows clearly that the bulk of the variation in ΔT results from the variation in MgO content of the glass in the coastal samples. This variation in turn may be due either to (1) variations in the quality of quench undergone by the coastal samples or (2) variation in the efficiency of the lava-tube systems over time.

Can we distinguish between these two possible causes of variation in MgO content of the coastal glasses? A fairly obvious shift in the size of ΔT occurs over time. The ΔT values for the 12 sample pairs collected from November 1986 through January 1988 average 12.3°C, whereas the ΔT values for all subsequent sample pairs average 8.4°C. In many of the first 12 pairs, the littoral sample was collected not at the beach but slightly inland and so may not have been as well quenched as the later samples, all of which were collected at the shore. However, the three earliest littoral samples, which were water-quenched, still have ΔT=9, 13 and 14°C. Many of the other samples were dipped into a can of water after collection on the point of a hammer, and so the larger ΔT in these early sample pairs cannot be dismissed solely as the result of inadequately rapid quenching. Finally, if we consider relative collection times, we see that some of the largest ΔT values are for sample pairs collected on the same day or adjacent days (table 1), and so these samples are not simply mismatched.

We cannot discard any of the early pairs as having unequivocal quenching problems, nor can we define a "best" subset of sample pairs with only small ΔT values. Therefore we regard the variation in ΔT values over time as meaningful, with the data showing that the lava-tube system became more consistent in its behavior and more thermally efficient after January 1988. This conclusion is consistent with field observations that the lava apron fed from Kūpaianaha was thin when it first reached the ocean, but thickened steadily thereafter (see Hon and others, 1994).

Variations in the Later Sample Pairs

The 25 sample pairs collected from February 1988 through November 1989 vary much less in ΔT than the earlier sample pairs. During this period, all littoral samples were retrieved from the surf zone and so were uniformly well quenched. Furthermore, although the pond surface became increasingly hard to reach, cable samples recovered from the rim continued to provide well-quenched material from the upper end of what was a mature tube system in a large lava field.

Within this remarkably uniform part of the data set, two features stand out: the first striking feature is that MgO contents of glass (and quenching temperatures) of the sample

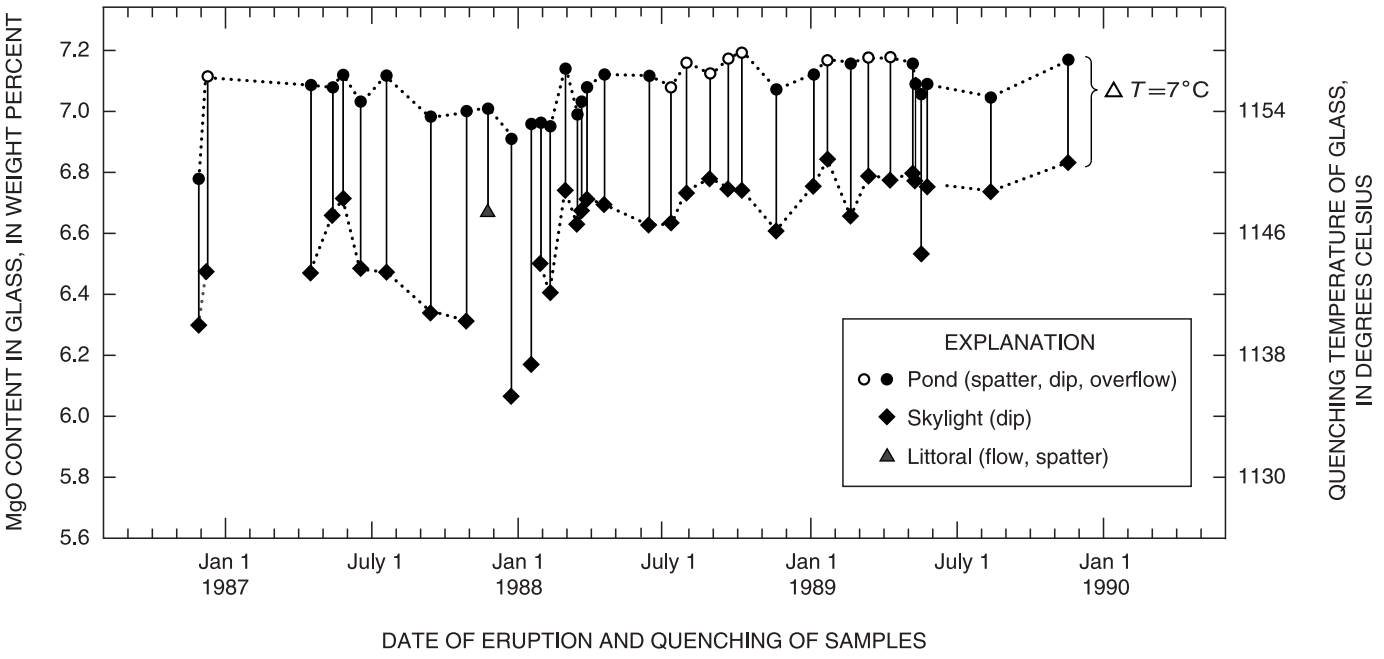

Figure 3. MgO content of glass and inferred quenching temperature versus time for pond-coast sample pairs from Pu‘u ‘Ō‘ō-Kūpaianaha eruption. Small divisions on time scale designate months. Open circles, samples with crystalline assemblage olivine + augite. All filled symbols, samples with three-phase assemblage olivine + augite + plagioclase.

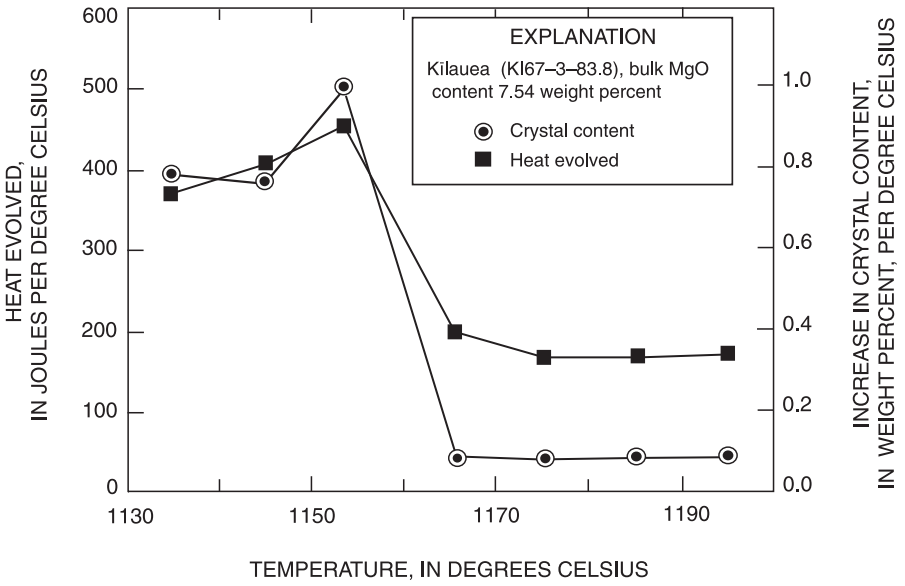

Figure 4. Heat evolved (= decrease in enthalpy) and increase in crystal content that occurs during equilibrium crystallization of a Kīlauea basalt as temperature decreases from 1,195 to 1,125°C, calculated using thermodynamic modeling procedure of Ghiorso (1985).

pairs seem to rise and fall together, with a near-constant ΔT= 7–9°C. This tight coupling of pond and coastal temperatures implies that the time needed for lava to flow from the pond to the coast is short, relative to (1) the gap in collection time between the samples in each pair and (2) the time needed to recharge the lava pond and tube system, either during continuous activity or after a pause in effusive activity, with lava of a slightly different temperature.

The net length of the lava-tube systems fed from Kūpaianaha was 12 km (not considering smaller-scale sinuosities), according to many observers (see Mattox and others, 1993), and the flow rate of lava in the tubes was determined (Kauahikaua and others, 1998) to be 1 to 3 m/s, where lava was visible through skylights. Thus, a particular parcel of lava could have taken as little as a few hours, mostly less than a day, to traverse the lava-tube system and emerge at the shore during periods of steady-state flow.

In contrast, the time scale for obvious fluctuations in magma supply ranged from many hours to a week, as did the collection time between samples (see table 1). Specifically, the April 1988 pause, the only pause within the time frame of figure 3, took place over a week (Heliker and Wright, 1991; see Heliker and Mattox, this volume). The tight correlation in MgO content between glasses in pond and coastal samples is consistent with rapid lava transport, in which the lava-tube system was flushed quickly and was always occupied by only one "batch" of lava, however defined.

The second striking feature of the 1988–89 data set is the well-defined minimum ΔT value of 7°C, observed in 7 (of 25) sample pairs, and corresponding to a temperature decrease of 0.6°C/km over the 12-km tube system. No sample pairs have a smaller ΔT value. We suggest that this limiting ΔT value represents the maximum thermal efficiency for lava transport from pond to coast, given the constancy of all other boundary conditions (slope, magma-supply rate, and physical properties of the lava).

Crystallization Behavior of Kīlauea Basalts

Model Crystallization Behavior

To evaluate more fully the significance of these temperature decreases, in terms of the processes that occur during lava transport in tubes, we need to consider how basaltic lavas crystallize. Specifically, we need to know what phases crystallize, at what temperature, and how much heat must be released as temperature decreases. The amount of heat evolved (the decrease in enthalpy) and the increase in crystal content that will occur, during equilibrium cooling and crystallization of a sample of Kīlauea basalt, are plotted in figure 4. The calculation procedure used is the thermodynamic model of Ghiorso (1985). The conditions of this calculation were equilibrium crystallization, at fo_2 equal to the quartz-fayalite-magnetite (QFM) buffer. The temperature decrement used was 10°, from 1,200 to 1,080°C. This temperature interval was chosen to facilitate comparison with the results of melting experiments, where the interval between runs is typically 10°C. The model calculates olivine and plagioclase compositions fairly closely but does not allow for any components in pyroxene other than the four end members (Di, En, Hd, Fs); thus, the calculated augite contains no Al_2O_3, Cr_2O_3, TiO_2 or Na_2O. Therefore, the calculated results only approximate the observed behavior of natural samples.

The results presented here are for a sample from Kīlauea Iki lava lake, with a bulk MgO content of 7.54 weight percent; calculations using various Kīlauea and Mauna Loa basalts show patterns very similar to those shown in figure 4. The phase relations of this basalt composition were determined experimentally (Helz and Thornber, 1987) and are typical of those of many other Kīlauea basalts that have

been studied experimentally (see Thompson and Tilley, 1969). Olivine ± chromite crystallizes first, at 1,169°C in this bulk composition, followed by augite at 1,165±5°C, and plagioclase at 1,155±5°C (roughly 10°C lower than augite). The calculations, which show olivine crystallizing first, with augite and plagioclase coming in between 1,170° and 1,160°C, do not resolve the difference between augite-in (the temperature at which augite first crystallizes) and plagioclase-in, but otherwise adequately reproduce the actual phase relations. This crystallization sequence is virtually identical to that observed in lava samples from the Pu‘u ‘Ō‘ō-Kūpaianaha eruption (Helz and others, 1991; Helz and Hearn, 1998), where augite-in is at 7.34 weight percent MgO (T=1,162°C) and plagioclase-in is at 7.00 to 7.12 weight percent MgO (T=1,155–1,157°C). Therefore, the data in figure 4 should be applicable to the sample suite considered here.

Figure 4 shows that, in Kīlauea lava compositions, the increase in crystal content and amount of heat released (the decrease in enthalpy, or -ΔH) have two distinct stages. There is a higher-temperature stage where olivine alone is crystallizing; here the enthalpy change is -176 J/°C per 100 g of basalt, and 0.1 weight percent of the melt crystallizes per degree. In the second stage, olivine + augite + plagioclase crystallize together. Along this three-phase cotectic, the change in enthalpy is -418 J/°C per 100 g of basalt, and 0.8 weight percent of the melt crystallizes per degree, or eight times as much as in the first stage.

The narrow (5–10°C) temperature interval within which the crystallizing assemblage is olivine + augite is not captured in the model calculations. In our discussion below, we assume that stage 2 is reached only when plagioclase begins to crystallize. Our lumping of olivine + augite with stage 1 is based on the observation that the amount of augite present in Pu‘u ‘Ō‘ō-Kūpaianaha samples at plagioclase-in is very low (see fig. 2*A*), and hence that the amount of heat released by its crystallization is small.

Virtually all samples in our suite have the assemblage olivine + augite + plagioclase and hence must crystallize and evolve heat at the rates specified for stage two above. Thus, for the range of temperature decreases commonly observed, the model predicts the following changes in enthalpy and crystal content:

for ΔT = -7°C	ΔH= -2930 J	5.6% more crystals in coastal samples
for ΔT = -9°C	ΔH= -3760 J	7.2% more crystals in coastal samples
for ΔT = -11°C	ΔH= -4600 J	8.8% more crystals in coastal samples

Translating these values to changes per kilometer, for 100 grams of basalt in a 12-km tube system, we obtain the following:

for ΔT = -7°C	0.6°/km cooling	ΔH= -240 J/km	0.47% crystals/km
for ΔT = -9°C	0.75°/km cooling	ΔH= -310 J/km	0.60% crystals/km
for ΔT = -11°C	0.9°/km cooling	ΔH= -380 J/km	0.73% crystals/km

These values show the range of increases in crystal content and decrease in enthalpy of the basalt that the thermodynamic equilibrium-crystallization model would predict for most pairs in this sample suite.

Observations on Crystallization and Differentiation of Samples

We cannot directly observe the heat lost during transport of lava in the 1986–89 tube systems, years after the fact. We can, however, evaluate whether the crystal content of the samples varies as expected. Two processes are potentially involved: (1) within-tube crystallization of the glass and (2) bulk differentiation of the lava. In the first process, some of the melt will have crystallized to produce additional olivine + augite + plagioclase. The new crystals will be tiny, having grown during the few hours of transport time in the lava tube, so they are unlikely to separate during flow. The second process would occur if some of the larger crystals present in the pond samples were lost during flow, or picked up by erosion of the walls and floor of the lava tube. The crystals most likely to be affected are the olivine phenocrysts and microphenocrysts (fig. 2).

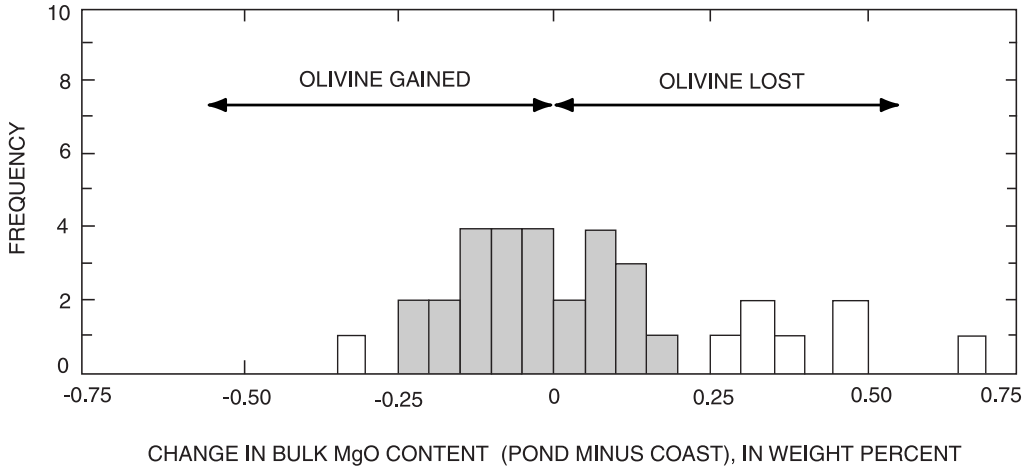

Figure 5. Histogram showing change in bulk MgO content for 35 pond-coast sample pairs, for which whole-rock analyses are available. Most sample pairs (shaded) show less than 0.25 weight per cent change in MgO content during transport in lava tube between pond and coast.

Table 4. Changes in crystal content for pond-coast pairs from the Kūpaianaha tube system, estimated from thermodynamic calculations and from changes in glass composition.

	Early pairs	Later pairs
Number of pairs	12	25
Average ΔT value and range, in degrees Celsius	12.25 (7–17)	8.4 (7–11)
Theoretical increase in crystal content and range, in weight percent	9.8 (5.6–13.6)	6.7 (5.6–8.8)
Average ΔTiO_2 and range, in weight percent	11.3 (5–20)	4.8 (1–9)
Average ΔK_2O and range, in weight percent	12.3 (4–27)	4.1 (0–11)
Average ΔP_2O_5 and range, in weight percent	13.5 (4–23)	8.2 (0–24)

Let us first consider whether bulk composition has been affected by either gain or loss of crystals between the flowing lava and the floor or walls of the lava tube. The difference in bulk MgO content between pond and coastal samples is shown in figure 5, for all pairs for which bulk analyses exist (as noted in table 1). We chose to plot MgO because it has the largest range in Kīlauea lavas (Wright, 1971) and because it will most strongly show the effects of any olivine fractionation in these samples.

In figure 5, the shift in bulk MgO for most of the 34 analysis pairs falls within $\pm$0.25 weight percent MgO (corresponding to $\pm$0.6 weight percent olivine), and the distribution of these samples is random around zero. We suggest that this part of the variation is dominated by sampling noise, presumably reflecting heterogeneity in bulk lava composition, perhaps with slight mismatches caused by varying time gaps in sampling. Seven analysis pairs have larger ΔMgO shifts, consistent with loss of 0.7–1.3 weight percent olivine, but this process is not pervasive. Changes in other aspects of bulk composition are small, so we conclude that lava bulk composition does not change significantly because of flow through the tubes.

The most straightforward way to estimate how much glass crystallizes during transport in the lava-tube system is to calculate the increase in the less-compatible oxides in glasses from coastal samples relative to glasses from pond samples. The most suitable oxides, based on looking at the microprobe analyses (Helz and Hearn, 1998), are TiO_2, K_2O and P_2O_5. Because the absolute values of some of these oxides are low, and the differences between analyses in the sample pairs are small, the differentials (such as ΔK_2O) from one pair of analyses to the next can be quite variable. Therefore, in table 4, we have shown averages for the differences in oxide concentration for the 12 early pairs and for the 25 later pairs, with the ranges of values for individual pairs indicated in parentheses. The increases in crystal content predicted by thermodynamic modeling, for the average ΔT of the early and later sample pairs, are included in table 4 for comparison. These results show that TiO_2, K_2O and P_2O_5 contents of glasses in coastal samples are consistently higher than those of the pond samples; therefore, the coastal samples have crystallized during transport by approximately the percentages shown. Glasses in the early coastal samples (with larger average ΔT) have crystallized more extensively than glasses in the later coastal samples. Finally, for both data sets, the estimated increases based on ΔTiO_2 and ΔK_2O are broadly consistent with theoretical predictions. The observed increase in P_2O_5 is larger than the theoretical prediction, but this difference (especially for the later pairs) is probably insignificant because the absolute P_2O_5 contents of these glasses are low (0.21 to 0.29 weight percent), and ΔP_2O_5 is very small and may be dominated by analytical noise.

Petrographic examination of the samples (fig. 2) suggests that the estimated 5 to 12 percent increases in microlite contents are reasonable. These results, together with earlier observations that bulk lava composition does not vary significantly during transport, lead us to conclude that (1) the melt in these lava tubes follows the equilibrium crystallization path quite closely and (2) the theoretical model (SILMIN) adequately models the actual (near-equilibrium) crystallization processes in the episode 48 lava tubes.

Heat Versus Temperature

The crystallization model of Ghiorso (1985) gives an approximate rate at which heat must be evolved during equilibrium crystallization of these basaltic liquids. Various studies of the actual cooling behavior of basalts (Keszthelyi, 1994, 1995a, b) have shown that heat loss is not uniquely related to temperature changes. For example, Keszthelyi (1995a) was able to detect increases in temperature following sudden bursts of nucleation of crystals. Such complications, though they do not occur under near-equilibrium conditions, nevertheless show directly that evaluating the thermal efficiency of lava tubes may involve more than simply monitoring changes in glass composition and inferred temperature.

We can illustrate one complication with some simple calculations. We assume a 12-km-long lava tube, with a total heat loss of 2,930 J (equivalent to ΔT= -7°C along the three-phase cotectic). However, for this calculation we assume that the input at the upper ("pond") end of the tube has a melt like the most magnesian glasses of the current Kīlauea east rift zone eruption, namely glasses from episode 30 at Pu‘u ‘Ō‘ō, with MgO contents of 7.90 to 7.99 weight percent and quenching temperatures of 1,174–1,175°C. We use the actual phase relations of the lava (plagioclase-in at 1,155°C for Pu‘u ‘Ō‘ō; Helz and others, 1991) to determine when the lava shifts from stage 1 (olivine-only) parameters to stage 2 (multiphase crystallization) parameters, as given above.

Because this hypothetical lava starts in stage 1, we begin our calculation with the change in heat content (ΔH) of -176 J/°C (versus -418 J/°C for stage 2). To lose 2,930 J of heat per 100 g of lava, the temperature must decrease by 16.6°C. Thus, the temperature at the coast for our hypothetical lava would be 1,158°C, still 3° above plagioclase-in. The lava will have been in stage 1 for the entire 12-km trip and will have undergone a total temperature decrease 2.4 times as great as a three-phase lava would, *at the identical thermal efficiency of the lava-tube system*.

If we assume a total heat loss of 3,760 J (corresponding to ΔT= -9°C under stage 2 conditions), we find that the entire range of 20°C at stage 1 does not quite consume all of the heat that must be dissipated. The remaining 242 J requires a further stage 2 temperature decrease of 0.6°C, for a total temperature decrease in the lava tube of 21°C. The change from one-phase to three-phase crystallization occurs 11.2 km down the lava tube, so the hypothetical lava will have been crystallizing only olivine for almost the entire distance, and the temperature change is more than twice what we observe in the actual samples. In evaluating the significance of rates of temperature decrease in tubes (here or as in Thornber, 2001), clearly the nature of the crystallizing assemblage is an important variable that must always be verified by inspection of the samples.

Results for Skylight Samples

Results comparing skylight samples to corresponding pond and/or coastal samples are summarized in figures 6 and 7, which plot the elevation at which the samples were collected against MgO content of glass. Figure 6 shows the early skylight samples (from Nov. 1987 to Jan. 1988), plus a larger set of skylight samples taken at the very end of the Kūpaianaha lava pond activity (see table 2). The shaded area in each set of data shows the range of slopes (ΔMgO, which is effectively ΔT) versus elevation generated by the range of MgO contents of glass in the pond and coastal samples. Other lines connect individual samples collected closest to each other in time. The most obvious feature in figure 6 is that some of the skylight samples do not fall on straight lines between the population of pond and coastal samples. The MgO contents of half the skylight glasses are low by 0.20 to 0.45 weight percent absolute, at the elevation of collection. The total range in replicate MgO analyses of glasses is 0.15 weight percent absolute, so these larger deviations in MgO content should be analytically significant, as are the deviations in temperature from straight-line decreases.

MgO contents of glasses from skylight and littoral samples from an episode 48 tube of February 1991, a single sample pair from episode 51 (November 1992) and a last

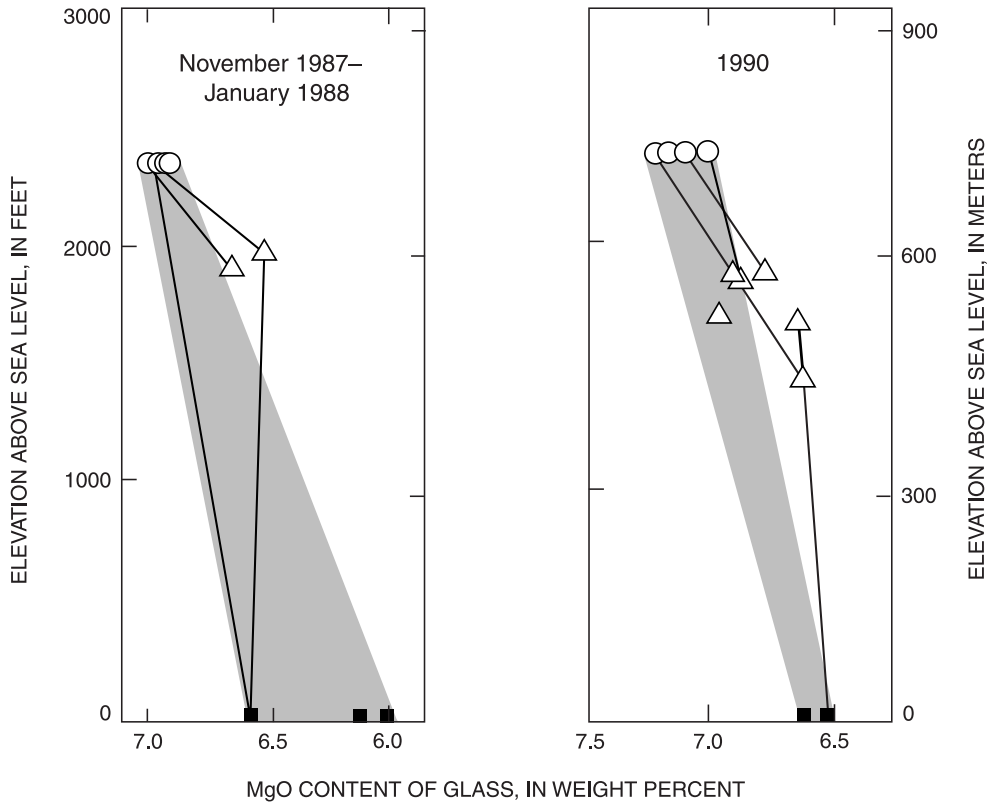

Figure 6. MgO content of glass from pond samples (circles), skylight samples (triangles), and coastal samples (squares) versus elevation at which samples were collected, from two different time periods. Heavy lines connect samples closely associated in time. Shaded areas indicate range of slopes shown by pond-coast data alone.

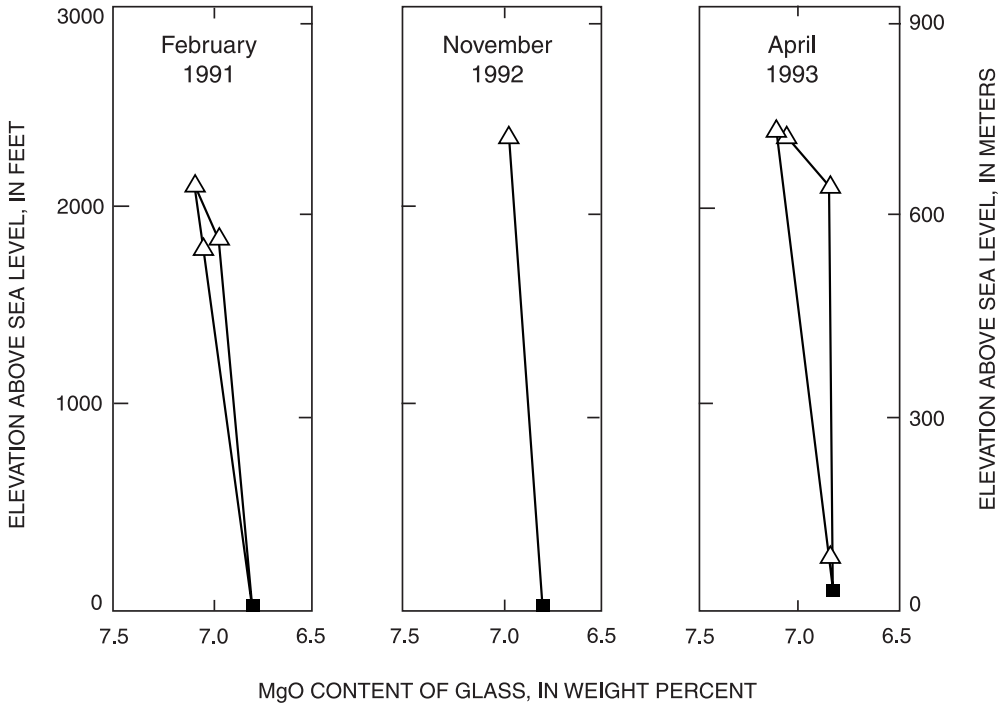

Figure 7. MgO content of glass from skylight samples (triangles) and coastal samples (squares) versus the elevation at which the samples were collected, during three different time periods. MgO content of glass in coastal samples is virtually constant throughout the entire period.

sample set from episode 53 (April 1993) are shown in figure 7. The latter two episodes were fed from Pu‘u ‘Ō‘ō itself, not from Kūpaianaha (fig. 1). The 1991 and 1992 data sets were previously described by Cashman and others (1994); the analyses used here are those of Helz and Hearn (1998). The various analyses, regardless of source, are very similar; in particular MgO contents differ by no more than 0.1 weight percent absolute in all instances. The skylight and coastal data in figure 7 resemble those in figure 6: the lines are steep to vertical, with a spread in MgO content in the upper skylight samples.

What is causing the nonlinearity evident in figure 6 (and, to a lesser extent, in fig. 7), such that so many skylight-littoral analysis pairs have calculated ΔT values in the range -1 to 4°C (see tables 2 and 3)? Clearly the behavior of the lava-tube system is unchanged: MgO in the skylight glasses has deviated on the low side from early in the lifespan of Kūpaianaha. Two possible explanations are (1) the upper part of the tube system loses heat much faster than the lower part, in which case the kinks seen in figures 6 and 7 are real, or (2) the skylight samples, which are difficult to retrieve, are less well quenched than the pond samples.

What could make the upper 2- to 4-km section of the lava-tube system thermally less efficient than the lower 6 to 8 km? One possibility would be enhanced loss of volatiles (mostly H_2O) because of the loss of the heat of vaporization that would accompany H_2O loss. However, the lava in the pond already had the opportunity to degas as it flowed under Pu‘u ‘Ō‘ō on its way to Kūpaianaha; also, the Kūpaianaha pond itself was effectively an immense skylight and must have allowed for extensive H_2O loss before lava entered the upper part of the tube system. In fact, Helz and others (1991) and Mangan and others (1995) noted the extreme dryness of the pond samples, as evidenced by their elevated plagioclase-in temperature (7.12 weight percent MgO, T=1,157°C) relative to that of Pu‘u ‘Ō‘ō samples and to samples taken after the pond crusted over (7.00 weight percent MgO, T=1,155°C). Finally, it is difficult to see why spontaneous degassing should become more efficient after the lava reenters a confined passageway. We believe that variable degassing rates are not responsible for the nonlinearity in glass MgO content.

A related possibility, suggested by Keszthelyi (1994), is that higher rainfall at higher elevations may cause more cooling in the uppermost part of the lava tube. However, if the rain falling on Pu‘u ‘Ō‘ō, on the 3 km of ground between Pu‘u ‘Ō‘ō and Kūpaianaha, and on Kūpaianaha itself was not enough to affect the condition of the pond samples, why should rainfall on the next 2 to 4 km of the tube system have so much effect on the skylight samples? In a more detailed analysis, Keszthelyi (1995b) divided the tube path into five segments and calculated their thermal budgets separately. His results, which consider many possible parameters, show that the rate of heat loss should be similar along most of the length of the tube, except for a small segment in the middle.

The alternative explanation (occasional quenching problems) is suggested by the fact that temperatures within the lava tube and even in air above skylights are very high (800–1,080°C; Peterson and others, 1994; Kauahikaua and

others, 1998). The process of withdrawing a sample from the lava stream, up through the airspace within the tube and out the skylight, is not instantaneous but takes minutes. The pond samples may not all be perfectly quenched, but they probably were more rapidly and more consistently quenched than the skylight samples.

This latter interpretation makes sense for the whole array of data in figures 3, 6, and 7, especially because half of the skylight data points in figure 6 are colinear with those of the pond-coast pairs. The idea that temperature decline in these lava tubes is approximately linear with distance is supported by the results of Thornber (2001), who presented detailed data for tube samples from the later part of episode 53. In his study, five sets of samples crystallizing at or below plagioclase-in (with up to six samples collected along the tube on a single day) show linear decreases in temperature from the Pu‘u ‘Ō‘ō vent out to almost 10 km.

Thus, we assume that the decrease in temperature (and heat loss, for a given crystallizing assemblage) is nearly linear with distance, so that skylight samples colinear with the pond-coast pairs (or having parallel slopes) represent successful sampling through skylights. To compare the total temperature changes for skylight-coast sample sets from episodes 51 and 53 with those from episode 48, we need to recall that the later tube systems were shorter than those fed from Kūpaianaha (10 km versus 12 km). Thus, ΔT=6°C for the 10-km-long lava tube system of episode 53 corresponds almost exactly to ΔT=7°C for the 12-km long Kūpaianaha system: both show a temperature decrease of 0.6°C/km. Thornber's (2001) minimum observed temperature decrease is 0.55°C/km. The close agreement in minimum cooling rate from these three studies strongly suggests that there is a limiting thermal efficiency for the lava tubes of the current Kīlauea east-rift-zone eruption.

Discussion

Repeated attempts have been made to quantify the change in temperature during flow of lava in tubes or along flows at Hawaiian volcanoes. In particular, the lava tube systems of the Mauna Ulu eruption were studied by Swanson (1973), who used an optical pyrometer to measure temperatures through skylights. He reported that "the highest optical-pyrometer temperatures of lava visible in each window of the tube system were 1,150° to 1,155°C, with no recognizable tendency for cooling downslope." Similarly, in their study of the 1984 Mauna Loa channelized ‘a‘ā flow, Lipman and Banks (1987) presented infrared-pyrometry and thermocouple data that showed unsystematic temperature variations along the flow, apparently all within 10–15°C of each other. The question posed by these and other early studies was, is the transport of lava really isothermal, or is the temperature variation too small to be defined by such measurements?

The consistent results of the present study and others (Cashman and others, 1994; Thornber, 2001) using glass geothermometry make it clear that the problem was the limited precision of field temperature determinations and that temperature decreases in lava tubes are readily quantifiable, given sets of well-quenched samples and an appropriate calibration. The calibration of Montierth and others (1995) will allow similar studies to be made on Mauna Loa lava, in the event of renewed eruptive activity and the formation of active tube systems on that volcano.

We have already noted that the minimum temperature loss per kilometer is the same in all three glass-geothermometry-based studies of the present Kīlauea eruption. We further note that the observed range of this parameter is fairly narrow (0.6–0.9°C/km for practically all sample pairs in this study from January 1988 through November 1989), and that this range of values has also been replicated by Thornber (2001), even though the sampling strategies used in the two studies were very different. We conclude that (1) there is a consistent and reproducible range of operating efficiency for Kīlauea lava tubes, at the magma-production rate of the ongoing east-rift-zone eruption, which can be documented over at least 10 years and (2) the sporadic lower ΔT values observed for skylight-coast pairs (tables 2, 3) are very unlikely to be significant.

Three additional features of these results deserve comment. First, the higher temperature decreases observed for pond-coast pairs collected before January 1988 result from variable but low quenching temperatures of the early littoral samples, at more or less constant temperatures for pond samples. We believe that these are caused by (unseen) variations in the thickness of the floor under the lava tubes in the new, relatively thin lava apron. The flows were thick enough for stable roofs to develop all the way to the coast; however, observers had no way of knowing whether the floor of a particular tube was entirely underlain by new lava, or by cold, older flows. We attribute the lower temperature decreases and more consistent littoral quenching temperatures observed after January 1988 to the development of a consistent floor across the area of the flow field.

Second, we should consider whether the differences in quenching temperatures shown in figure 3 are maximum or minimum values. That is, do the pond samples represent the average temperature in Kūpaianaha or just its surface temperature? We cannot be completely sure of the extent to which the temperature in the Kūpaianaha lava pond may have varied with depth. We suggest that any temperature gradients were small because the pond was constantly circulating and overturning, and because the range of MgO content in the pond glasses is quite limited (fig. 3). Moreover, Pu‘u ‘Ō‘ō spatter samples collected during episode 48 (from July 1991 on) are no hotter than the hottest of the Kūpaianaha samples (Helz and Hearn, 1998). The pond samples are surface samples, however, so to the extent that any bias exists, the ΔT values given in table 1 and figure 3 are minimums.

The coastal samples must be strongly biased toward sampling the hottest, fastest-moving part of the lava stream in the tube. The observed variation in stream velocity in the lava tube (Kauahikaua and others, 1998) may reflect variations in temperature across the lava stream; thus, the temperature decreases observed are (again) minimums for the system.

Table 5. Data on ʻAilāʻau selvage glasses, with estimated source temperatures.

[Parenthetical values indicate that the initial temperature estimate was higher than 1,155°C (plagioclase-in). For these samples, temperatures above 1,155°C are assumed to correspond to single-phase (olivine-only) crystallization, and the initial estimates were corrected accordingly. The resulting temperatures imply original (vent) MgO contents of 6.87–7.97 per cent by weight. Data from Clague and others (1999)]

Sample batch	No. of specimens	Distance from inferred vent	T_{MgO}°C for average selvage glass	Estimated source T_{MgO}°C (minimum)
1	5	0–5 km	1,153	1,153–(1,165)
2	2	5–10 km	1,145	1,148–1,151
3	10	10–15 km	1,143	1,149–1,152
4	4	15–20 km	1,150	(1,165)–(1,171)
5	6	20–25 km	1,142	1,154–(1,160)
6	2	25–30 km	1,143	(1,162)–(1,169)
7	9	30–35 km	1,142	(1,167)–(1,174)
8	15	35–40 km	1,139	(1,167)–(1,174)

Third, it is important to remember that the observed temperature decreases are all for lava undergoing multiphase crystallization (olivine + augite + plagioclase). The rate of temperature decrease for lavas crystallizing only olivine will be 2.4 times greater: the corresponding ΔT values would be 1.44–2.16°C/km, at exactly the same rate of heat loss along a lava tube. The controlling variable is heat loss, not temperature decrease, so petrographic examination of samples is required.

Other Studies of Lava Temperatures at Kīlauea

Studies of variations in lava temperature at Kīlauea have not focused exclusively on evaluating lava-tube systems. Cashman and others (1999) reported data on samples collected along an open lava channel, to elucidate factors controlling the transition from pāhoehoe to ʻaʻā observed in the channel. Application of glass geothermometry to these samples showed temperature decreases of 4.4–6.8°C/km, almost an order of magnitude higher than those observed in lava tubes, as would be expected from field observations. Modal analysis of these samples showed that the crystal content along the flow channel increased by 0.83 percent crystals per degree, as temperature decreased. This rate is similar to crystallization rates inferred in the present study (table 4), and also compares well with results of theoretical calculations (fig. 4), which showed an increase in crystal content of 0.8 weight percent per degree. Deviations from equilibrium are apparently not large in the hottest part of active lava flows, even in open channels. This observation is important, because the MgO glass geothermometer is valid only at equilibrium (Helz and others, 1995).

Glass geothermometry can be applied to older lava flow or tube systems, but the results must be interpreted carefully. For example, Clague and others (1999), in their excellent study of the ʻAilāʻau lava flow field (dated at A.D. 1445±30), reported average glass analyses of selvages from various parts of the flow field and translated the MgO contents of those glasses to temperature. Their results (summarized in table 5 by sample group, inferred distance from vent, and apparent quenching temperature of the glasses) show a relatively coherent decrease in the MgO content of glass with distance from the inferred vent (near the present Kīlauea Iki Crater), with a corresponding inferred temperature decrease of 0.35°C/km. The authors correctly note that this is much less than the observed cooling rate in tubes, being half the minimum rate of temperature decrease observed in the lava tubes of the current eruption.

The probable explanation for the limited compositional range of the ʻAilāʻau glasses is that they are not well-quenched samples; they are merely naturally occurring flow selvages, subject to a period of undercooling and disequilibrium crystallization prior to becoming rigid. Undercooling in pāhoehoe toes and flow margins has been observed in sampling of active flows (Helz and others 1995); under these conditions, MgO glass geothermometry (controlled by olivine crystallization) failed to keep up with actual measured temperature decreases. This process is probably responsible for the convergence in selvage glass compositions in the ʻAilāʻau lava flows. The sample base of Clague and others (1999) may include samples equivalent to the distal (littoral) samples presented here. For an eruption that occurred in A.D. 1445, however, we cannot obtain samples equivalent to pond spatter, tears from Puʻu ʻŌʻō, or dip samples from skylights or other active flow channels. Such samples require diligent real-time collection, an artificial quench, or both.

What have we learned about the ʻAilāʻau eruption from these glass data? First, the general decrease in MgO content of glasses with distance from the vent should be real and would reflect the decline in maximum MgO in the melt in

the original tubes/channels. Second, the existence of sample group 4 (table 5), with distinctly more magnesian glasses for their distance from the summit, suggests that the temperature of ʻAilāʻau lavas erupted at the vent varied over the course of the eruption or, conceivably, that some of the lava was erupted from a different vent.

What were source temperatures likely to have been? If we assume the maximum observed thermal efficiency for well-developed lava tubes of the present eruption (0.6°C/km for multiphase crystallization, or 1.4°C/km for olivine-only crystallization), we can use the distances from the inferred vent to arrive at minimum source temperatures for the lavas. These temperature estimates, given in the last column of table 5, range from 1,153 to 1,174°C. Most of the more distant samples (groups 4–8) have probable temperatures above plagioclase-in, at the low end of olivine-only crystallization. This temperature range is plausible for lavas originating at Kīlauea's summit (Wright, 1971).

We have a modern analogue that supports this interpretation. Melts in the same temperature range inferred for the ʻAilāʻau lavas were actually erupted during the high-fountaining episodes at Puʻu ʻŌʻō; however, such magnesian melts were not recovered from flow margins, even close to the vents. Instead all flow (selvage) samples contained the three-phase assemblage olivine + augite + plagioclase (Helz and others, 1991; Helz and Hearn, 1998). If these hotter melts are not recoverable from flow margins, even for samples collected and quenched by human intervention, they almost certainly would not be preserved in flow margins collected centuries after the eruption.

Other Aspects of Lava Transport in Tubes

Most studies of lava tube systems focus on the question, what is the maximum distance that tube systems can deliver lava? Keszthelyi (1995b) provides a good summary on this subject, reviewing many factors that go into quantifying the thermal budget of tube systems. A more specific question that we consider here is, what more can we learn from the active tube systems at Kīlauea?

We have consistent determinations of the maximum thermal efficiency for transport of lava in the present tube system. However, lava at the coast has not reached a limiting crystal content (figs. 2*B*, 2*D*), so no direct estimation of maximum possible length can be made. Even longer Hawaiian lava-tube systems, such as the 1859 Mauna Loa system (see Greeley, 1987), are stopped by the ocean, not by high viscosity produced by high crystal contents (Helz and others, 1993, and R.T. Helz, unpub. data, 1995). Also, our ability to evaluate the effect of variation in slope is limited by the small range of slopes found on tholeiitic shield volcanoes, in addition to the relatively short distances involved. In fact, extremely long lava flows, such as those of the Columbia River Basalt Group (Tolan and others, 1989) and those of North Queensland (Stephenson and Griffin, 1976) have traveled many times farther on lower slopes than the longest Hawaiian lava flows or tubes. This observation suggests that regional slope is not particularly critical in limiting lava tube lengths.

Other parameters of interest that have been investigated include actual structure, flow velocity of lava tubes, and the cross-sectional area of the lava stream. Thermal imaging obtained by Realmuto and others (1992) shows an intricate and fairly regular pattern of sinuosity and bifurcation of the Kūpaianaha lava-tube system of October 1988. Analysis of the observed structure by Keszthelyi (1994, 1995b) suggests that bifurcations have little effect on overall heat loss. The regularity in pattern implies a fairly predictable difference between net length and actual length of the lava tubes. Although length is not a major source of uncertainty in thermal efficiency, knowing true length would enable us to refine the estimates of temperature decrease and rate of heat loss. Kauahikaua and others (1998) and Kauahikaua and others (this volume) present unique data on velocity variations and on cross-sectional areas of the lava stream that are essential to increasing our understanding of the dynamics of the whole lava-tube system.

Another dynamic parameter, which remains relatively unknown, is lava flux rate along the length of the tube. Specifically, we do not know what fraction of the lava that enters the top of the lava-tube system emerges at the bottom. Evidence for downcutting of the tube floor (Kauahikaua and others, 1998) documents the erosion of material and its incorporation into the lava stream. Yet it is thermodynamically impossible for more lava to come out the bottom of the tube than flowed in at the top, so any erosion must be more than compensated by loss of material from the lava stream to the walls or floor elsewhere along the tube.

We presume this "lost" material is lodged in the inflating, thickening flow field (Hon and others, 1994), where it must represent a large part of the thermal budget of these lava-tube systems. Initial modeling by Keszthelyi (1994) led him to conclude that the temperature decrease during transport must be much greater than the 1°C/km that he used in his model, an estimate based loosely on the present results (as reported by Helz and others, 1991, 1993). However, if even a fairly small fraction of the lava that enters the tube never emerges, the ΔT values observed for the lava stream are not sufficient to define the thermal budget (integrated heat loss) for the whole tube system. Keszthelyi (1995b) subsequently recognized that lava effusion rate is a critical parameter, but his treatment of this parameter was necessarily theoretical.

We may be able to obtain data on lava flux in Kīlauea's lava tubes if the present activity continues, but can we recover any information applicable to the past 20 years' activity? One possible approach would be to look more closely for changes in bulk composition of the lava stream. We have looked briefly at changes in bulk MgO content of the pond-coast sample pairs and concluded that systematic changes in bulk composition, as a result of either cooling or flow differentiation in the lava tubes, appear to be small. Nonetheless, it might be fruitful to look more closely for changes in bulk composition at the trace element level, especially for anomalous increases in incompatible elements.

Summary and Conclusions

This chapter presents the results of applying the glass geothermometer of Helz and Thornber (1987) to a suite of 37 sample pairs of very glassy lava collected from the upper and lower ends of the episode 48 tube system, while the Kūpaianaha lava pond was open and accessible. We also include data on a smaller suite of skylight samples collected from early in the lifetime of Kūpaianaha through the end of episode 48 (in 1991), plus some data on skylight-coast sample pairs collected during episodes 51 and 53. The results are as follows:

(1) From November 1986 through January 1988 (15 months), the average change in inferred quenching temperature (for 12 sample pairs) is 12.4°C, and the average increase in crystal content (inferred from observed enrichment of TiO_2 and K_2O in the coastal glasses) is about 11 to 12 weight percent.

(2) From February 1988 through November 1989 (23 months), the average change in inferred quenching temperature (for 25 sample pairs) is 8.4°C, and the average increase in crystal content is 4 to 5 weight percent. Within this part of the data set, pond and coastal temperatures seem to rise and fall together much of the time, even though these temporal fluctuations are at or below the limit of resolution of glass geothermometry ($\Delta T \leq 3$°C).

(3) The minimum difference in temperature (ΔT) for any pond-coast pair is 7°C. Of 37 sample pairs, 24 have ΔT=7–9°C, over the entire 3-year period. These data, which form a coherent set, give us a well-defined range of temperature decreases in the episode 48 tube system, over its lifetime.

(4) The skylight samples are somewhat equivocal in their significance. About half the glasses in the skylight samples have MgO contents consistent with their position along the lava-tube system, and the other half have lower MgO contents. Apparently some of the skylight samples are not as well quenched as the pond and littoral spatter samples.

Other studies using glass geothermometry to evaluate temperature decreases in lava tubes of the ongoing Kīlauea eruption have produced similar results. The well-documented ΔT value of 6°C for some episode 53 samples (Cashman and others, 1994), collected from the 10-km long tube fed from Pu‘u ‘Ō‘ō, gives exactly the same temperature decrease with distance (0.6°C/km) as the limiting ΔT of 7°C observed for the 12-km tubes fed from Kūpaianaha. The range of efficiencies observed in the present study has been replicated by the results of Thornber (2001) for samples collected in 1996–98 during episode 53.

Better information on the actual length of the path traversed by lava in these tube systems would help establish the limits of overall thermal efficiency of Kīlauea lava tubes. However, the biggest gap in our understanding of these lava tube systems is lack of information on the efficiency of mass flux through the tubes. Without estimates of how much lava is retained in the lava-field system, the best-quantified changes in lava temperature give us only part of the story.

Acknowledgments

We wish to acknowledge the dedication of the HVO staff from 1986 through 1993. This chapter depends heavily on their continuous efforts to make rigorous field observations and to collect well-quenched and well-documented samples. In addition, we acknowledge B. Carter Hearn's meticulous electron microprobe analyses, which form part of the data base discussed here. The manuscript was greatly improved by reviews from Kathy Cashman and Jim Kauahikaua.

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