Magma transport at Hawaii: Inferences based on igneous thermobarometry

Keith Putirka
Lamont-Doherty Earth Observatory of Columbia University, Palisades, New York 10964

ABSTRACT

Pyroxene + liquid equilibrium in Hawaiian lavas occurs at a range of pressures for each volcano. Ranges are systematic and may be related to the stage of development of the magma conduit system. Kilauea, which is in its shield-building phase, yields relatively shallow storage estimates. Loihi and Mauna Kea, which are in the early and late stages of volcano growth, respectively, yield deeper storage estimates. Shallowest depth estimates at Loihi and Mauna Kea are similar to estimates of elastic plate thickness, suggesting that the mechanical behavior of the lithosphere, rather than density contrasts at the Moho, regulates magma delivery. Apparently, a large increase in fracture energy below the brittle-ductile transition inhibits transport at depth, whereas magma transport by fracture propagation is rapid through the brittle lithosphere. Some shallow depth estimates at Kilauea support the hypothesis that the strength of the unbuttressed southeast flank influences magma storage. Kilauea transport depths correlate with an eruption sequence, which illustrates a top-to-bottom emptying of the conduit system. Successively deeper reservoirs at Kilauea were tapped within 300 days, indicating that magma is stored at a range of depths, including in the mantle, rather than at a single level within the lithosphere.

INTRODUCTION

Magma transport and storage are influenced by density variations (Stolper and Walker, 1980), stress (Dieterich, 1988), and elastic rock properties (Shaw, 1980). Pressures of crystallization estimated from igneous thermobarometers can be translated into magma staging depths and used to constrain magma transport and storage. This paper presents a reconnaissance study of *P-T* estimates from Hawaiian lavas using new pyroxene-liquid thermobarometers.

For some volcanoes at Hawaii, the composition, structure, and seismic activity have been well studied. Such observations were integrated with pressure (*P*) and temperature (*T*) estimates of pyroxene + liquid assemblages in lavas from Mauna Kea, Kilauea, and Loihi volcanoes to sort out controls of magma transport. As these volcanoes span much of the history of a single Hawaiian edifice (Moore et al., 1982), the resulting *P-T* estimates may yield a time-integrated view of the Hawaiian magma conduit system.

METHODS

P-T estimates were computed from the pyroxene + liquid thermobarometers of Putirka et al. (1996), which are based on the jadeite content of pyroxene. To avoid pyroxene–whole-rock pairs unlikely to yield valid *P-T* estimates, several data filters were employed. Most samples were nearly aphyric. *P-T* estimates were for pyroxenes described as euhedral phenocrysts, when that information was given. Pyroxenes were checked to yield four cations per six oxygens and minimal Fe^{3+}. Samples were also avoided in which geochemical data indicated magma mixing. Most rocks contained olivine, and a test for closed-system behavior was made on the basis of $K_D^{\text{ol-liq}}$[Mg-Fe] (= 0.30 ± 0.03; Roeder and Emslie, 1970). For five samples (2A, 3A, H26, H1, MU-9) $K_D^{\text{ol-wr}}$[Mg-Fe] did not equal 0.30 (wr = whole rock), and the bulk composition was corrected by adding or subtracting the reported olivine-core composition. If corrections exceeded 10%, or required subtraction of more olivine than the reported modal abundance, the *P-T* estimate was considered unreliable. In several cases, more than one pyroxene analysis was available; in these cases *P-T* estimates were individually calculated and averaged.

Temperatures for olivine + liquid equilibrium were determined. An olivine + liquid thermometer, applicable to pressures up to 30 kbar (cf. Roeder and Emslie, 1970), was calibrated from existing experimental olivine + glass pairs. The resulting equation is

$$\frac{10^4}{T} = 6.41 + 0.75\ln\left[\frac{\text{Mg}^{\text{ol}}}{\text{Mg}^{\text{liq}}}\right] - 18.1\text{NaAl}^{\text{liq}} - 1.42\ \text{Si}^{\text{liq}} - 0.026P \qquad (1)$$

where *T* is in Kelvins and *P* is pressure in kilobars. The superscript "liq" refers to the cation fraction of the indicated element in the liquid, and Mg^{ol} is the cation fraction of Mg in olivine. The term NaAl^{liq} is the product of the cation fractions of Na and Al in the liquid. Experimental data used for calibration are from sources cited by Putirka et al. (1996). The thermometer has an error of ±31 K.

Loihi lavas consist of four tholeiites (Hawkins and Melchior, 1983; Garcia et al., 1995), a transitional basalt (187-7), a hawaiite (186-14), and a basanitoid (186-11) (Garcia et al., 1995). Kilauea lavas are from the East Rift Zone vent, Puʻu Oʻo (Garcia et al., 1989, 1992). Puʻu Oʻo lavas have been divided into episodes; episode 1 began on January 3, 1983. The *P-T* estimate for episode 1 was determined from an average of three lavas. *P-T* estimates for Mauna Kea (Frey et al., 1991) have been discussed previously (Putirka et al., 1996); a picrite and a basalt (Frey et al., 1990) have been added to this data set.

P estimates were converted to depth from second-order polynomial depth vs. *P* curves derived from seismic- and gravity-constrained density profiles of Hawaii (Hill and Zucca, 1987). Depths to Moho used for Mauna Kea, Kilauea and Loihi were 18, 15, and 12 km below sea level, respectively (Hill and Zucca, 1987). Maximum regression error on the depth vs. *P* curves was <0.6 km. Small errors in density estimates and placement of density boundaries changed the curves very little. The most significant source of error on depth estimates was from error on *P* estimates.

RESULTS

Most *T* estimates from equation 1 were equivalent to pyroxene *T* estimates within error (Fig. 1). In three cases (3-A, 186-7, and 186-11), multiple pyroxene compositions yielded significantly different *P-T* estimates; pyroxene *T* estimates closest to those calculated by using equation 1 were selected. For all other samples, agreement of *T* estimates indicated that the estimates were not spurious. The average difference between olivine and pyroxene temperatures was 31 K, perhaps indicating that olivine saturation preceded pyroxene crystallization. Application of a liquid line of descent program (Longhi,

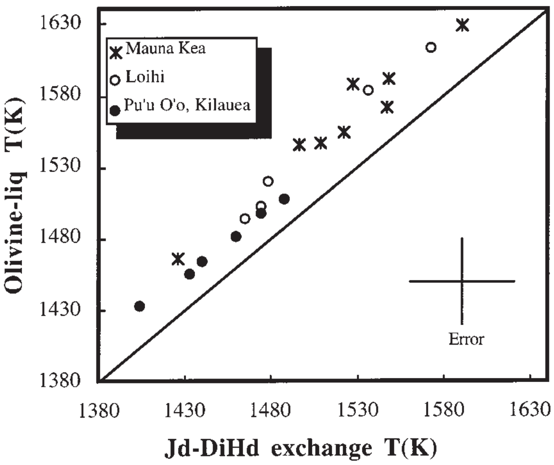

Figure 1. Temperatures from equation (1) are compared to *T* calculated from pyroxene–whole-rock compositions (model T2 of Putirka et al., 1996). A one-to-one correlation line is shown for reference; error for individual *T* estimates is indicated.

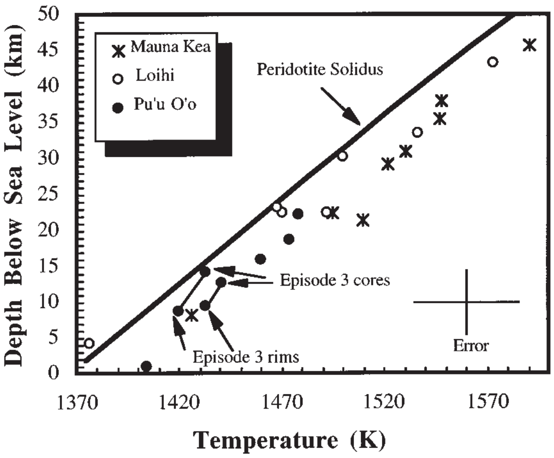

Figure 2. Temperature and depth estimates calculated from pyroxene + liquid thermobarometers of Putirka et al. (1996) and pressure to depth conversion discussed in text. McKenzie and Bickle's (1988) mantle solidus is shown for reference. Also illustrated are depth estimates computed from pyroxene-rim compositions for episode 3 lavas at Pu'u O'o, Kilauea. Depth-estimate error is approximated from error on *P* estimates.

TABLE 1. *P-T* DEPTH ESTIMATES

Sample no.	*T* (K)	*P* (kb)	Depth below sea level (km)
Mauna Kea			
LP-5	1530	10	31
MK1-8	1496	7.5	16
MK6-18	1509	7.2	21
MU-2 rim	1402	0.1	–2.7
MU-2 core	1426	3.2	8.3
MU-8	1522	9.6	29
MU-9 corrected	1547	12	36
H1	1590	15	46
H26	1548	12	38
Loihi			
2-A	1474	6.5	22
3-A	1479	6.5	27
4	1544	11	37
186-3	1573	13	43
187-7	1468	6.8	24
186-11	1500	9	30
186-14	1376	0.9	4
Pu'u O'o, Kilauea			
Episode 1	1404	0.4	1
3-88	1441	3.7	13
3-88 (rim)	1433	2.7	9.5
3-96	1433	4	14
3-96 (rim)	1420	2.5	8.7
9-175	1460	4.5	18
9-178	1474	5.3	20
10-187	1488	6.3	21

1991) to several bulk compositions showed that 4% olivine should crystallize over a *T* interval of 29 K prior to pyroxene saturation. This percentage of olivine was less than the reported modal olivine for the samples tested. It is thus concluded that higher-*T* estimates from olivine are plausible and possibly real.

Most *P* estimates fall to the high-*T* side of the mantle solidus (Fig. 2). Most depth vs. *T* estimates from Mauna Kea and Loihi also overlap, whereas lavas from Kilauea yield lower depth and *T* estimates. Two discrete depths have possible relevance to magma transport: 10 km, the shallow limit for most Kilauea lavas, and 20 km, the approximate deep limit for Kilauea lavas and the shallow limit for Mauna Kea and Loihi tholeiites.

P estimates greater than 1 bar (Table 1) indicate that magma transport was rapid enough that pyroxenes did not reequilibrate at low pressure. On the basis of a diffusion coefficient (*D*) for pyroxene of 7.0×10^{-16} cm^2/s (1200–1250 °C; Brady and McCallister, 1983) and the relationship $t = d^2/D$ (t = time; d = distance), a crystal with a radius of 0.25 mm (average pyroxene size for Pu'u O'o pyroxenes) will require 35 400 yr for reequilibration. For a 10^2-fold increase in *D*, reequilibration will still be on the order of hundreds of years. Preservation of high *P-T* information is thus unsurprising, even for small crystals, since residence times of magma in shallow reservoirs are likely shorter than the time required for pyroxene reequilibration (Dvorak and Dzurisin, 1993). The limiting condition for extracting *P-T* information is that magma and coexisting crystals have approximated a closed equilibrium system which has recorded a period of magma storage and pyroxene growth.

DISCUSSION

Constraints may be placed on transport processes if the distribution of depth estimates correlates with transport barriers. Below, potential transport barriers are discussed and compared to depth estimates from thermobarometry.

Depth Estimates of 0–15 km

Depth estimates of less than a few kilometers likely reflect storage in summit or rift-zone reservoirs and can be attributed to a shallow level of neutral buoyancy (Ryan, 1987). Depth estimates for episode 3 lavas at Kilauea may be consistent with ponding of magma at the base of the volcanic edifice or perhaps near the Moho (Hill and Zucca, 1987). Although the Moho may occur at 10 km beneath Kilauea (Hill and Zucca, 1987), the density above the Moho at all three volcanoes is 2.9 g/cm^3 (Hill and Zucca, 1987) and does not provide a level of neutral buoyancy to magma with a density of 2.7 g/cm^3. A highly fractured edifice may yield a density barrier regardless of Moho placement. Closure of fractures is expected to be complete between 9 and 10 km (Ryan, 1987); above this depth, bulk density may be reduced. Also, closure of fractures will require a magma overpressure for dike propagation—without magma accumulation, transport at depths of >10 km by fracture may thus be inhibited.

Kilauea's unbuttressed southeast flank may also influence magma transport (Denlinger and Okubu, 1995; Delaney et al., 1990; Ando, 1979). Dieterich (1988) noted a decrease in the proportion of extrusive to intrusive igneous activity following the 1975 Kalapana earthquake and postulated that igneous activity is modulated by the strength of the detachment beneath Kilauea's southeast flank. Between eruptive episodes, rift-zone intrusions increase the horizontal stress (σ_h). Magma pressure may become great enough to support a column of magma equal to the height of the edifice, and eruptions occur. Failure along the detachment reduces the supporting stress, and subsequent magma pulses are trapped. Depth estimates of 10 km may thus reflect a period of temporary magma storage following failure of the detachment.

Whereas most core and rim analyses yield similar *P-T* estimates, some pyroxenes were zoned. Compared to cores, rim compositions from episode 3 and MU-2 yield distinctly lower *P-T* estimates (Table 1). These differences may indicate discrete ponding levels or crystallization during upward transport. Magma mixing may also cause pyroxene heterogeneity (sample 3-88; Garcia et al., 1992). Pyroxenes that nucleate after mixing, however, will yield valid *P-T* estimates; *P* estimates would then place mixing at nontrivial depths for some samples.

Depth Estimates of >15 km

The maximum depth estimate was 46 km for a lava from Mauna Kea. This depth reaches a localized zone of low seismic velocities (Ryan, 1988) and is close to the greatest depth recorded for seismic tremor (Aki and Koyanagi, 1981). This correlation may indicate that low velocities result from the presence of melt.

The depth to the Moho at Loihi is probably 12–13 km; at Mauna Loa, the Moho is at 18 km (Hill and Zucca, 1987) and should be similar at Mauna Kea. Most depth estimates at Mauna Kea and Loihi exceed these values. Depth estimates for Pu‘u O‘o episodes 9 and 10 also exceed depth to the Moho at Kilauea. These depth estimates, and the density relationships discussed above, show that the Moho does not provide an intrinsic barrier to magma transport beneath Hawaii and that magma storage does not necessarily occur at the crust-mantle boundary.

Depth estimates near 20 km are in the range of estimates of T_e, the elastic thickness of the lithosphere (Watts et al., 1985). Bodine et al. (1981) related the depth represented by T_e to the brittle-ductile transition. A change in the mechanical behavior of the lithosphere at 20 km is also indicated from earthquakes that occur as swarms at <20 km, but are unclustered at depths >20 km (Dvorak and Dzurisin, 1993). Earthquake swarms are observed from regions with recent volcanic activity (Sykes, 1970) and have been related to a change in strength and fluid flow (Scholz, 1990, p. 209). It is inferred that below depths of 20 km, ductile behavior inhibits fracture transport of magma. The absence of shallower depth estimates at Mauna Kea and Loihi may result from the facility with which fracture mechanisms can transport magma through a relatively brittle lithosphere.

Deep estimates from Loihi and Mauna Kea may indicate that magma transport above depths of 20 km is relatively rapid compared to rates of pyroxene equilibration. Flexural stresses, which result from loading of the oceanic crust by the growing volcanoes, vary with depth; part of the lithosphere is in compression and part is in tension (ten Brink and Brocher, 1987). Ten Brink and Brocher (1987) placed the neutral surface that separates these regimes at 20–25 km, where there is also an observed gap in seismic activity. During the early phases of volcanism, σ_h is tensile above 20 km and compressive below. Tensile

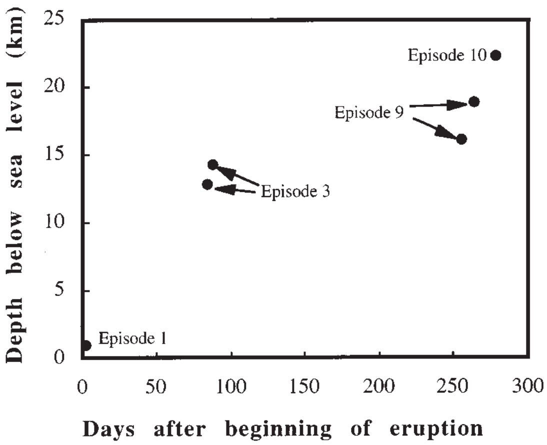

Figure 3. Depth estimates for Pu‘u O‘o samples are plotted against number of days after eruption began (January 3, 1983). Eruptive episodes are indicated.

stresses above 20 km should facilitate magma transport without stagnation; pyroxene reequilibration (of crystals formed at greater depths) may thus be limited. The upper 20 km of lithosphere, though, is also tensile at Kilauea. The lack of estimates >20 km at Kilauea may indicate that transport barriers occurring at 0–15 km (discussed in the previous section) are more efficient during the shield-building stage compared to other stages of volcano growth. It is also conceivable that 0–15 km depth estimates at Kilauea result from slow transport rates during the shield-building phase (high *P* pyroxenes are reset to lower *P*). Given the length of time required for reequilibration coupled with short magma processing times it is perhaps more likely that pyroxenes have nucleated between 0 and 15 km at Kilauea. A scenario for slow transport, though, is discussed below.

Controls on Fracture Transport of Magma

The correspondence of shallowest depth estimates for Mauna Kea and Loihi with elastic plate thickness implies a relationship between fracture mechanisms and lithosphere strength. Stresses are concentrated at crack tips because of their shape, and propagation occurs when the stress intensity factor, *K*, at the crack tip exceeds a critical value, K_c. The stress, σ, at a dike tip is given by $\sigma = \sigma_r(1 + 2h/w)$, where σ_r is the remote stress, *h* is the dike half-height, and *w* is dike width (Scholz, 1990). As a result, stress concentration is a function of aspect ratio, *h/w*: for a given dike width there will exist a dike height beyond which $K > K_c$. A transition from brittle to ductile behavior will cause an increase in K_c (Scholz, 1990, p. 37): first, the crack must overcome the additional work of plastic flow; second, plastic flow blunts the crack tip, which reduces stress concentration. A substantial increase in fracture energy below the brittle-ductile transition is thus anticipated. Such an inverse relationship between strength and fracture energy is observed for metals (Cherepanov, 1979).

At Mauna Kea and Loihi, ductile creep mechanisms at depths of $<T_e$ (20 km) are ineffective at relieving stress at crack tips; magma is transported to the surface by fracture mechanisms without shallow-level storage. Below T_e, dikes will stall or propagate at subcritical velocities until dikes become relatively large; a range of transport depths may reflect a range in dike *h/w*. Shallower transport depths at Kilauea may be due to the mature nature of the conduit system. A steady magma supply at Kilauea should cause greater thermal softening of the wall rock and local increases in K_c compared to Loihi and Mauna Kea. Slower transport rates at Kilauea may result. As frequency of eruptions wanes, elastic properties of the conduit walls relax back to regional values. Thermal softening of the conduit may also explain the range of depth estimates at Mauna Kea: rocks with $P < 8$ kbar are from the shield-building stage (high magma-supply rate) whereas rocks with $P > 8$ kbar are from transitional and post-shield stage (low magma-supply rate).

Reservoirs Tapped During Sustained Eruptions

Because deflation of the Kilauea summit is closely followed by eruptions at rift zones, Wolfe et al. (1987) concluded that the summit reservoir and rift zones are hydraulically coupled. When eruptive volumes exceed summit deflation volumes, tapping of magma stored below the summit reservoir is indicated (Dvorak and Dzurisin, 1993). The depth versus time relationships indicated in Figure 3 support this hypothesis and are consistent with the timing of changes in deflation and eruption volumes (see Table 1 of Dvorak and Dzurisin, 1993). A progressive emptying of the conduit system, from top to bottom, is thus indicated. Also, because a systematically deepening range of depths is sampled in less than 300 days, magma is likely stored over a range of depths rather than at a single level within the lithosphere.

SUMMARY

Pyroxene + melt thermobarometry applied to Hawaiian lavas produces linear *P-T* arrays parallel to the mantle solidus. A 0.4 kbar estimate for Pu‘u O‘o episode 1 at Kilauea indicates ponding near the summit. Pyroxenes from later episodes, and from Loihi and Mauna Kea, yield greater depth estimates. The shallowest recorded equilibration depths for most Loihi and Mauna Kea lavas are similar and exceed depth estimates from Kilauea.

Density contrasts at the Moho do not provide a level of neutral buoyancy. The mechanical behavior of the lithosphere, however, appears to play a significant role in controlling magma transport. During early and late stages of volcano evolution, the conduit is not well developed, and magma may be transported by fracture mechanisms through a brittle lithosphere. Below T_e, fracture energy increases and magma transport is inhibited; dikes must be large enough so that the stress intensity at their tips exceeds K_c. As a conduit system evolves and softens, magma may stall at shallower depths, as at Kilauea.

Barriers other than variable fracture energy may explain some transport depths. Flexural stresses are tensile above 20 km during the pertinent phases of volcano growth, thus facilitating transport through fractures at shallow depths. Some depth estimates at Kilauea are consistent with the hypothesis (Dieterich, 1988) that detachment strength beneath Kilauea influences magma transport. Magma-transport depths at Kilauea are also observed to have progressively increased with time for the Pu'u O'o eruption. This sequence shows that the conduit system is progressively emptied from top to bottom. The rapid change of depth estimates at Pu'u O'o also indicates that magma is stored over a range of depths within the lithosphere.

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