Nature Of Local Magma Storage Zones and Geometry Of Conduit Systems Below Basaltic Eruption Sites: Pu'u 'O'o, Kilauea East Rift, Hawaii, Example

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The fluid dynamics of the well-documented eruptive episodes at Pu'u 'O'o, Kilauea (Wolfe et al., 1987) are used to investigate quantitatively the size and shape of the shallow conduit system beneath the vent. Cooling calculations are employed to study the long-term survival of conduits and the consequences of multiple dike injection events. We find that the subvent conduit must have a planar geometry at depths greater than a few tens to at most a few hundreds of meters, with a width of less than a few meters, a length of the order of 100 m, and a height of the order of 1 km. This structure is clearly the residue of the preeruption dikes. Although such a feature can be widened somewhat by repeated dike emplacement events, there is no evidence to suggest that a larger and much more equant magma reservoir should develop at shallow depth. The extensive degassing often occurring after single eruptions or between repeated eruptions, previously thought to imply the presence of a large equidimensional shallow magma chamber, can be readily explained by the volume of magma in a dikelike subvent planar magma storage zone. On the basis of these observations, which we infer to apply commonly to basaltic eruption sites, we suggest that the terms "chamber" or "reservoir" be used with caution, because they most often connote a relatively large equant magma body, distinct from the feeder dike, lying just below the vent. We find that the subvent region has the same basic geometric characteristics as the parental dike, and we propose that it be referred to as a "planar magma storage zone."

1. INTRODUCTION

The well-documented nature of the recent activity at the Pu'u 'O'o vent on Kilauea's east rift zone [Wolfe et al., 1987] has concentrated attention on a number of aspects of the supply of magma from summit areas to flanks of shield volcanoes. The Pu'u 'O'o eruptive episodes took place near the center of a zone (Figure 1) that had been the target of numerous dike injection episodes from the summit region of Kilauea over a number of years [Dzurisin et al., 1984], with a major injection event just prior to episode 1 of the eruption [Wolfe et al., 1987]. Seismic and deformation evidence summarized by Wolfe et al. [1987] suggests that magma from earlier intrusions was still present in the rift zone at the time of this major injection event, and that the main vent for subsequent eruptive activity (Pu'u 'O'o) was located above the site of this magma.

The detection of gas release from the main vent [Casadevall et al., 1987] during the approximately 3-week intervals between subsequent eruptive episodes [Wolfe et al., 1987] strongly suggests that at least part of the major preepisode 1 dike retained a considerable volume of magma at the end of the episode 1 eruption. This magma body continued to be present both between and during the subsequent eruptions, located at a sufficiently shallow depth to exsolve volatiles during the repose periods and having dimensions which enabled it to avoid excessive cooling and solidification.

In this paper we consider the possible geometry of this region beneath the vent, through which magma rises mainly vertically during eruptive episodes from the deeper dike system that, in turn, conveys melt laterally from the summit magma reservoir about 18 km away. In particular, we use the dynamics of the eruptive episodes to place restrictions on the size and shape of this region, and use thermal calculations to show that the geometry is consistent with the region being the fluid residue of the partially cooled, major preepisode 1 dike. We use the Pu'u 'O'o example to illustrate some general properties of shallow magma storage zones.

2. PRODUCTION OF SMALL MAGMA BODIES: EAST RIFT ACTIVITY PRIOR TO THE Pu'u 'O'o EVENTS

Activity on Kilauea's east rift zone between 1955 and 1982 has been summarized by Wolfe et al. [1987] drawing on analyses of both intrusive [Dzurisin et al., 1980; 1984] and extrusive [Moore
et al., 1980] activity. For example, between 1976 and the end of 1982, nine small intrusions occurred in the east rift zone and three eruptions took place. Dzurisin et al. [1984] provide data implying that just over 200 x 10^6 m^3 of magma was injected into the rift zone during this period. However, geodetic or seismic data are insufficient to allow the dimensions of any of the discrete magma bodies to be deduced. To investigate the possible long-term survival of these bodies against cooling, we assume that the total volume of magma was shared equally between the nine bodies formed, giving each a volume of about 25 x 10^6 m^3.

Magma intruded into rift zones typically produces narrow, steeply dipping bladelike dikes [Rubin and Pollard, 1987]. Assuming, on the basis of recent estimates of east rift dike shapes [Dvorak et al., 1986] that a typical dike is 5 to 10 km long and 1 to 3 km high, the above volume implies dike widths in the range from 1 to 5 m. The intervals between successive dike injections during the 1976–1982 period ranged from 11 to 840 days. We therefore need to investigate the ability of planar dikes with this width range to remain partly liquid for a wide range of time intervals, and to account for the fact that a given dike may be intruded into cool rocks that have not been associated with dike injection for some years, or into a region that is hot and was itself a dike quite recently.

The time taken to freeze a dike of a given width (in the sense that the center reaches the solidus) has been calculated as a function of the initial temperatures of the fresh dike and the country rocks and the thermal properties of each using the method and physical properties given by Turcotte and Schubert [1982]. Full account is taken of heat transfer by conduction and latent heat release. Figure 2 shows some results for dikes intruded at the liquidus temperature into country rocks which are cooler by 100, 200, and 800 K. It is clear that, depending on the time since the most recent previous intrusion, a 1-m-wide dike freezes in a time ranging from 1 week to 2 months. A 5-m-wide dike can remain partly liquid from 4 months to 2.5 years. Thus no dike with a width in the 1- to 5-m range is likely to have been able to retain melt for the longest of the quiet periods (840 days) in 1976–1982, whereas essentially any of them would have been liquid for the shortest quiet period (11 days).

To gain an idea of the possible patterns of development of accumulations of magma in the east rift, we conducted a series of 1000 Monte Carlo-type simulations. In each case, nine dikes were assumed to be injected laterally along the same vertical plane in a hypothetical rift zone at intervals corresponding to those actually recorded between 1976 and 1982. The dike centers were located at random in the downrift direction, and the width of each dike was assigned randomly from the range quoted above. The calculations illustrated in Figure 2 were used to determine the width of the residual molten zone at the time of the next dike injection. At each new intrusion, an estimated value was assigned to the temperature of the surrounding country rocks based on the length of time since the last intrusion. These simulations imply that a residual unsolidified body of width 1.9 ± 2.3 m may have been present in the rift by the end of the 1976–1982 sequence. This width is significantly less than the accumulated thickness of all the dikes, about 15 m using average dike length and height values. The analysis highlights the point that a single long repose period between east rift intrusions can effectively solidify small magma bodies intruded frequently over a long period of time. Nevertheless, we shall see shortly that there is evidence for the presence of a large amount of melt in the rift at the start of the Pu‘u ‘O‘o eruption series. We note for later reference that a dike (residual melt) width of 1.9 m, combined with the average height, 2 km, given by the Dvorak et al. [1986] estimates, implies that a maximum magma volume close to 4 x 10^6 m^3 per km of horizontal extent is stored in the rift.

3. Eruption Precursor Activity: Dimensions and Emplacement Conditions of the Preepisode 1 Dike

The geophysical observations relating to the emplacement of a major dike into the east rift be-
...from Pu'u Kamoamoa to between Pu'u Kahaualea by Wolfe et al. [1987]. An important aspect of the subsequent major eruption sites' Pu'u 'O'o and the bodies will have produced a thicker and slightly vent complexes (designated 0740, 1123, and 1708). and Kalalua (see Figure 1), and contained all of the deformation was close to instantaneous [Wolfe et al., across which the pressure increase causing surface dike eraplacement was essentially aseismic and event is the evidence for a zone within which the sion occurred along the same plane within the rift region as a result of the earlier dike injection events of which only a segment is shown.

Fig. 3. Schematic diagram of the evolution of a magma body in a rift zone. (a) A new dike is intruded, (b) this dike cools and the molten zone shrinks (dashes show original size), (c) another new dike intersects the older melt body (dotted line shows contact between old and new liquid), and (d) composite body cools and the molten zone shrinks, producing somewhat more equant body than at stage in Figure 3b. Note that, for simplicity, only the evolution of the vertical extent of these dikes is shown; similar developments take place in the horizontal direction, of which only a segment is shown.

between January 2 and 7, 1983, have been described by Wolfe et al. [1987]. An important aspect of the event is the evidence for a zone within which the dike emplacement was essentially aseismic and across which the pressure increase causing surface deformation was close to instantaneous [Wolfe et al., 1987]. This zone extended for about 5 km downrift from Pu'u Kamoamoa to between Pu'u Kahaulea and Kalalua (see Figure 1), and contained all of the subsequent major eruption sites: Pu'u 'O'o and the vent complexes (designated 0740, 1123, and 1708). The geophysical evidence implies that a quantity of at least partially molten magma was present in this region as a result of the earlier dike injection events discussed in the previous section. If the new intrusion occurred along the same plane within the rift zone, the combination of the old and new magma bodies will have produced a thicker and slightly more equant body of melt as suggested qualitatively in Figure 3.

Dvorak et al. [1986] used deformation data associated with the major dike injection to deduce the size and shape of the dike immediately after its emplacement. To make the problem tractable, they idealized the dike shape as a rectangular slab, for which they found a vertical height of 2.4 km, a horizontal length of 11.4 km, and a thickness of 3.6 m, with its top located just below the ground surface. These dimensions imply a volume of just less than 100 x 10^6 m^3 (about 4 times larger than the average volume associated with individual dikes during the 1976–1980 period). Wolfe et al. [1987] note that the volume of lava erupted during episode 1 was 14 x 10^6 m^3, and that Dvorak et al. [1986] have estimated a value of 60 x 10^6 m^3 for the amount of magma leaving the summit reservoir in association with the episode 1 events. The residual dike volume of 100 x 10^6 m^3 then suggests that, in addition to the magma from the summit reservoir, about 54 x 10^6 m^3 (i.e., (100 – 60 + 14) x 10^6 m^3) of magma were incorporated into the preepisode 1 dike from that part of the rift system uprift of the dike's eventual location.

The site of the start of seismic activity in the rift marking the injection of the new dike was located [Wolfe et al., 1987] about 1 km uprift of Makaopuhi Crater (Figure 1), i.e., about 11 km from the summit. Using our earlier estimate of the typical magma content of the rift, 7 x 10^6 m^3/km, the upper 11 km of the rift system may contain about 45 x 10^6 m^3 of magma at any one time. This value is of the same order as the 54 x 10^6 m^3 required by the above calculation. Although there are inevitably large uncertainties in all of these volume estimates, we conclude that the preepisode 1 event marked the substantial depletion of magma in the upper part of the rift zone and its concentration into the zone of subsequent eruptions.

We can obtain the most likely maximum total thickness of the magma body underlying the eruption zone at the end of the episode 1 eruptions by combining the 3.6-m width calculated by Dvorak et al. [1986] for the preepisode 1 dike injection with the 1.9 ± 2.3 m width found in the previous section for the accumulated magma width likely to be present in the rift at any random time; this yields an estimate of 5.5 ± 2.3 m.

4. Evidence for the nature of the local magma storage zone: episode 2 onward

We assume that eruptive activity in the vicinity of Pu'u 'O'o from episode 2 onward was due to fresh magma from the summit reservoir connecting with and flowing through the region representing the uncooled portion of the preepisode 1 dike. That the episode 1 pathway was reused is indicated by the lack of seismic swarms associated with the onset of the second and later episodes [Wolfe et al., 1987]. Over the course of the next few episodes, eruptive activity settled into a fairly repeatable pattern, suggesting a nearly constant conduit geometry and stress field. The time and volume relations between summit tilt and effusion rate from the vent [Wolfe et al., 1987] indicate continuity of magma through the system, so that at any instant there is a well-defined mass flux everywhere and we can use continuity and other dynamic arguments to investigate the geometry of the plumbing system. We have considered elsewhere (L. Wilson and J. W. Head, Geometry of volcano rift dike systems and the dynamics of magma supply to flank eruptions: Pu'u 'O'o and the Kilauea East Rift zone, submitted to Journal of Geophysical Research, 1988; hereafter referred to as WH88) the lateral connection between the summit reservoir and the near-vent region, and concentrate here on that part of the system located below the Pu'u 'O'o vent.

The volatile content of the magma erupted during the Pu'u 'O'o episodes was typically about 0.03 wt % CO₂, 0.1 wt % SO₂, and 0.3 wt % H₂O [see...
Fig. 4. Essential features of the geometry of the subvent magma body acting as a storage zone between eruptions and a magma pathway during active episodes. The width, W, horizontal length, X, and vertical extent, Z, are defined. The connection feeding magma from the summit reservoir, and also the connection to the surface vent, are indicated schematically.

Greenland, 1987]. Based on volatile solubility data [e.g., Wilson and Head, 1981], the major phase, H\textsubscript{2}O\textsubscript{2}, would start to exsolve at a depth of about 230 m, and so a rapid increase in speed and decrease in bulk density would occur in magma moving up to depths less than this. The eruption dynamics in this shallow part of the system, and in particular the relations between the morphology of the fire fountain, the surface vent size, and the amount of volatile phases exsolved from the magma are dealt with by Head and Wilson [1987]. Here we are considering the somewhat deeper zone, where volatile release during the eruptive episodes is minimal, but slower long-term degassing takes place during repose periods. The important aspects of the geomelry of the system are shown in Figure 4, and can be related to the eruption dynamics as follows.

An incompressible magma with density \( \rho \) subjected to a driving pressure gradient \( dP/dz \) in a fissure of width \( W \) (see Figure 4) flows upward at a constant speed \( u \) given by

\[
W(dP/dz) = \rho fu^2
\]

[e.g., Knudsen and Katz, 1958], where \( f \) is the Fanning friction factor. In laminar flow,

\[
f = \frac{24}{Re}
\]

where \( Re \) is the Reynolds number defined as

\[
Re = \frac{2Wup}{\eta}
\]

Here \( \eta \) is the apparent viscosity of the magma (assumed to be a Newtonian fluid), for which we adopt the value 200 Pa s, typical of basalts erupted at high rates from the east rift zone of Kilauea [Shaw, 1969]. If the Reynolds number is large enough (greater than about 2000) that the motion is incipiently or fully turbulent, a more complex expression must be used for \( f \) [Knudsen and Katz, 1958]; however, we shall show later that this is not necessary for the Pu‘u ‘O‘o eruptions. The volume flux passing through the system, \( F \) (which must be equal to the dense rock equivalent volume flux emerging from the vent), is given by

\[
F = uWX
\]

where \( X \) is the horizontal extent of the region through which magma moves vertically beneath the vent. If the vertical extent of this region is \( Z \), its volume, \( V \), is given by

\[
V = WXZ
\]

The above equations can be combined to yield

\[
W^2 = \frac{12\eta FZ}{V(dP/dz)}
\]

We now demonstrate that values can be assigned to all of the variables in the above equations except \( W \), \( X \), and \( dP/dz \). Thus solutions can be found by assuming values for \( dP/dz \); (6) then provides the consequent value of \( W \), and (5) the corresponding value of \( X \).

At a depth of 1 km (approximately the depth to the center of the preepisode 1 dike) a batch of east rift magma will have exsolved most of the available CO\textsubscript{2} and SO\textsubscript{2}, a total of about 0.1 wt % [Gerlach and Graeber, 1985; Greenland, 1987]. Calculating gas density as a function of temperature, pressure, and molecular weight [e.g., Wilson and Head, 1981, equation 8], it can be shown that this gas represents only about 5% of the volume of the gas–liquid mixture. At these pressure levels the gas volume is essentially inversely proportional to the pressure and hence the depth. Thus in the region of interest, changes in the amount of exsolved gas exert a minimal influence on the bulk density of the magma, and we can assume the magma to be incompressible with a density of, say, 2700 kg m\textsuperscript{-3}.

L. P. Greenland (personal communication, 1986) has used SO\textsubscript{2} release rates measured during interepisode periods [Greenland et al., 1985; Casadevall et al., 1987] to estimate the total volume of magma which resides between the pressure levels 40 MPa (above which pressure, at depths greater than about 1650 m, SO\textsubscript{2} release is negligible) and 1.5 MPa (this lower limit representing an estimated 50-m–thick plug of degassed magma left at shallow depth within the vent as an eruptive episode subsides). The measurements indicate that the volume of magma contained in the 1600-m vertical extent of the storage zone implied by these depth limits averaged 7.5 x 10\textsuperscript{8} m\textsuperscript{3} over eight episodes (seven values ranged from 3.6 to 13.6 x 10\textsuperscript{8} m\textsuperscript{3}, and one very small value of 0.9 x 10\textsuperscript{8} m\textsuperscript{3} was recorded). We therefore adopt values of \( Z = 1600 \text{ m} \) and \( V = 7.5 \times 10^8 \text{ m}^3 \) in the subsequent analysis. We explicitly assume that all of the SO\textsubscript{2} production occurs within the region where magma rises nearly vertically during eruptions. This is suggested by the fact that the deeper dike connecting the preepisode 1 dike to the summit appears, from seismic evidence [see
Wolfe et al., 1987, Figure 17.8], to lie at depths greater than 2 km, preventing it from contributing to interepisode SO₂ degassing.

The volume effusion rate, \( F \), is known for the Pu'u 'O'o eruptions: a typical value for episodes near the middle of the series is 160 m² s⁻¹ dense rock equivalent [Wolfe et al., 1987]. Since we shall be solving the above equations using the pressure gradient applicable at the beginning of an eruptive episode, and since the pressure gradient must decrease during an eruption as the summit reservoir is depleted, we assume that the initial value of \( F \) is about double the average value during an episode, and so adopt \( F = 320 \text{ m}^3\text{g}^{-1} \) in what follows.

Decker et al. [1983] examined the relationship between the summit tilt variations at Kilauea and the elevations on the volcano of eruptive vents. They assumed that the magma reservoir pressure head at the end of an eruptive episode was equal to the elevation of the vent below the summit and used this to relate changes in tilt to changes in the summit reservoir pressure, finding that 1 microradian of tilt change corresponded to a pressure change of about 8.5 x 10⁴ Pa. The tilt change during an eruption then corresponds to the change in reservoir pressure driving magma through the pipe system.

The summit tilt typically changed by 15 microradians during the repose periods between later eruptive episodes at Pu'u 'O'o, suggesting that the pressure difference driving magma through the pipe system was about 1.275 MPa at the start of a typical eruptive episode. If this pressure difference gave rise to a uniform pressure gradient throughout the approximately 15-km length of the pipe system, the gradient would be about 70 Pa/m. However, there is clearly the possibility that a large part of the total pressure change occurs over a small part of the system, giving rise to locally larger and smaller pressure gradients than the mean value.

This may be particularly true at the start of an episode, while the plug of magma that has cooled in the vent during the previous repose period is being discharged, or at the end of an episode, when the overall pressure difference is small. We discuss elsewhere [WH88] the possibility that eruptive episodes may be terminated when a relatively cool batch of magma, which resided in a narrow part of the pipe during an earlier repose period, encounters another narrow zone in such a way that its small but finite yield strength is able to balance the residual pressure gradient. We take as a lower limit on the size of a region of locally high pressure gradient beneath the vent a linear (vertical) extent of 1600 m, the extent of the magma storage zone implied by the interepisode degassing. This corresponds to an upper limit of the pressure gradient of about 800 Pa/m, an order of magnitude greater than the mean value; as the analysis proceeds, it will emerge that nothing new is established by considering values larger than this.

It is more difficult to establish a priori a likely lower limit on any locally small pressure gradient within the pipe system. We consider values down to 3 Pa/m (one and a half orders of magnitude smaller than the mean) in what follows and show later that such small values in the near-vent region imply a wide vent for which there is no geodetic evidence.

Using values of \( dP/dz \) between 3 and 800 Pa/m as discussed above, we have solved (5) and (6) for the consequent values of \( W \) and \( X \) (Table 1). All of the solutions lead to the same value for \( u \) (0.6827 m/s), as can be seen by eliminating the product \( WX \) from (4) and (5) and noting that \( V, F, \) and \( Z \) are all fixed in this analysis. Using (3), values were computed for the Reynolds number, \( Re \), for each line of the table, and all of the values are seen to correspond to laminar magma flow conditions, thus justifying the use of (2) for \( f \).

5. LOCAL MAGMA STORAGE ZONES: DISCUSSION OF PERMISSIBLE GEOMETRIES

The results given in Table 1 show that for large pressure gradients a narrow, planar geometry is deduced for the subvent storage zone. The predicted shape becomes more equant as the assumed pressure gradient is decreased, the extreme case of a square cross section occurring for \( dP/dz \approx 3.6 \text{ Pa/m} \), when \( W = X \approx 22 \text{ m} \). The equivalent solution for a circular tube has almost the same pressure gradient and a diameter of about 24 m.

We argued earlier that the maximum width of the magma body underlying the vent region could be estimated as 5.5 ± 2.3 m by adding the width of the preepisode 1 dike deduced from deformation measurements to a generous estimate of the width of any magma body likely to be present in the rift from much earlier and partly cooled intrusions. When ap-

<table>
<thead>
<tr>
<th>( (dP/dz) ), Pa/m</th>
<th>( W, \text{ m} )</th>
<th>( X, \text{ m} )</th>
<th>( Re )</th>
</tr>
</thead>
<tbody>
<tr>
<td>800</td>
<td>1.43</td>
<td>327.5</td>
<td>26</td>
</tr>
<tr>
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<td>31</td>
</tr>
<tr>
<td>400</td>
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<td>231.6</td>
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<tr>
<td>200</td>
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<td>163.8</td>
<td>53</td>
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<td>141</td>
<td>3.41</td>
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<tr>
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<td>13.88</td>
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<tr>
<td>6</td>
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<td>4.2</td>
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<tr>
<td>3</td>
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</table>

\( W \), the mean width of an elongate rectangular zone through which magma rises; \( X \), the horizontal extent of the zone required to permit the mean observed volume flux; and \( Re \), the implied Reynolds number.
TABLE 2. Model Solutions for a Scenario in Which a Planar Subvent Storage Zone of Width 3.5 m is Connected to the Surface Vent by a Tube of Diameter 10 m

<table>
<thead>
<tr>
<th>L/m</th>
<th>$V_T$, x($10^8$ m$^3$)</th>
<th>$V_P$, x($10^8$ m$^3$)</th>
<th>$Z_P$/m</th>
<th>$u_P$, m/s</th>
<th>$(dP/dz)_P$, Pa/m</th>
<th>$X_P$/m</th>
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<tr>
<td>0</td>
<td>0.0</td>
<td>7.50</td>
<td>1600</td>
<td>0.683</td>
<td>133</td>
<td>134</td>
</tr>
<tr>
<td>30</td>
<td>2.4</td>
<td>7.48</td>
<td>1570</td>
<td>0.672</td>
<td>132</td>
<td>136</td>
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<td>7.9</td>
<td>7.42</td>
<td>1500</td>
<td>0.647</td>
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<tr>
<td>300</td>
<td>23.6</td>
<td>7.26</td>
<td>1300</td>
<td>0.573</td>
<td>112</td>
<td>160</td>
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<td>700</td>
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<td>6.95</td>
<td>900</td>
<td>0.414</td>
<td>81</td>
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<td>6.48</td>
<td>300</td>
<td>0.148</td>
<td>29</td>
<td>617</td>
</tr>
</tbody>
</table>

For tube lengths, $L$, values are given for $V_T$, the volume of magma in the tube; $V_P$, the volume of magma in the planar zone required to produce a total volume of $7.5 \times 10^8$ m$^3$ (see text); $Z_P$, the vertical height of the planar zone; the magma rise speed, $u_P$, and the pressure gradient, $(dP/dz)_P$, needed in the planar zone to produce the required flux of 320 m$^3$/s; and the horizontal width, $X_P$, of the planar zone.

As a result we infer that solutions with $W$ greater than about 6 m should be rejected at a very high confidence level, so that $(dP/dz)$ is likely to be greater than about 45 Pa/m and $X$ to be greater than about 80 m. The most likely solution is one with $W \leq 3.5$ m, for which $(dP/dz) \geq 133$ Pa/m and $X \geq 134$ m.

The fact that there is some latitude in specifying the size and shape of the storage region allows us to explore the possibility that the relatively equant shape of the surface vent in plan view (which, toward the end of the series of eruptive episodes, had a diameter of about 10 m) may continue to some depth. We have assessed some possible geometric scenarios by calculating the volume of magma stored between eruptions in a 10-m-diameter tube of a given length and then using the methods described in the previous section to find the size and shape of the planar reservoir below the tube needed to accommodate the residual volume.

Table 2 shows some results for tube lengths in the range zero to 1000 m. In all cases, $W$ is set equal to 3.5 m, the most probable value based on the earlier arguments. It is clear that the main effects of adopting this modification to the geometry of the subvent system are to increase the required horizontal extent of the planar region (though not to an extent which conflicts with the deformation evidence) and to decrease the pressure gradient required to drive magma at the required flux through the planar region. The changes in $X$ and $(dP/dz)$ are modest until the tube length becomes comparable with the 1600-m vertical extent of the storage region. We cannot exclude such scenarios on the basis of the available evidence but find it hard to imagine that the cooling of an initially planar dike, or the mechanical erosion of the country rock surrounding such a dike, could actually lead to such an extreme geometric shape for the system. Thus we prefer the solutions in which almost all of the 1600-m-high region contributing to the degassing has a planar shape (Figure 5).

Basaltic eruptions, especially in rift zones, commonly begin with the opening of a long fissure, a clear indication that they are due to the intersection of a dike with the surface. The concentration of magma flow early in an eruption from a long series of closely spaced active centers (the "curtain-of-fire" phase) to a few locations or a single center is readily understood as the consequence of large horizontal variations in the balance between heat supply (proportional to magma flux per unit horizontal length of fissure) and heat loss by conduction through the walls. Such variations can be caused by quite small variations in the fissure width, as discussed by Delaney and Pollard [1981, 1982]. These facts strongly suggest that the vast majority of basaltic vents are underlain by planar magma bodies as indicated in Figures 4 and 5. The residual bodies of magma from single dikes can become somewhat inflated by subsequent intrusions, as indicated in Figure 3, but the aspect ratios of most dikes in the shallow lithosphere (especially in rift zones [Rubin and Pollard, 1987]) are likely to be so great that the essentially planar shape is always retained.

6. CONCLUSIONS

1. Basaltic vents in volcanic rift zones commonly originate when dikes intersect the surface (Figure 1),
and so are initially underlain by narrow, nearly vertical, planar, magma-filled zones (Figures 4 and 5).

2. These shallow planar zones commonly persist for periods of months (Figure 2) and, if subsequent eruptions occur at the same site (as happened at Pu'u 'O'o), may widen slightly (Figure 3), providing a source of degassing between eruptions. They continue to act as magma pathways during active periods (Figure 4).

3. The shapes of the near-circular conduits commonly observed feeding the vents inside basaltic cones must at least in part be related to the building of the cone or to activity within a lava pond inside the cone; they must merge quickly with a more planar, dike-shaped structure at shallow depth (Figure 5).

4. The planar unsolidified residues of earlier shallow dikes can be rejuvenated if they are intersected by later dikes. The process is only important, however, if the interval between successive injections is sufficiently short. Figure 2 shows that the interval required to maintain a fluid core in a dike is about 1 week to 1 month for dikes 1 to 2 m wide injected into "cold" country rocks. Once some preheating by the first one or two injections has occurred, the interval need only be 2 to 9 months. However, net widening of dikes requires injections on significantly shorter time scales. There is no evidence for the growth in this way of very shallow magma reservoirs with near-equant dimensions.

5. On the basis of these observations and conclusions, we urge that the terms "chamber" and "reser-
voir" be used with caution, because they most often connote a relatively large equant body, distinct from the feeder dike. In this study we find that the subvent region has the same basic geometric characteristics as the dike system and is thus a "planar magma storage zone." We propose the nomenclature specified in Figure 6, in which the terms "deep dike system," "shallow subvent dike," and "conduit" describe the properties of the various elements of the system in commonly understood terms.

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