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MAGMA ACCUMULATION IN HAWAIIAN PLUME SOURCES

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ABSTRACT. The steady state eruptions of Kilauea, Hawaii cannot be related to a percolative plume source, because continuous melt migration results in overly high overpressures in the upper part of the partially molten plume. The ascent of melt must instead be discontinuous. It is suggested that melt initially accumulates in conduits, and that melt episodically intrudes upward from one system of conduits to others situated above. The ascent is thus envisaged to occur by disperse multiple mini-intrusions.

INTRODUCTION

Basaltic magmas are generated in the mantle by partial melting of lherzolite. Initially minute drops of melt form interstitially between the grains of lherzolite. The extrusion of basaltic lava flows with volumes exceeding 1.0×10^8 m^3 and the presence of large intrusions with volumes of tens of cubic kilometers and more show that the interstitial melt somehow must accumulate into large batches of magma. Direct evidence of the structure of the partially molten mantle cannot be obtained, so the nature of the accumulation process must be inferred from available field evidence of peridotites and theoretical approaches.

The partially molten mantle may exist in two different structural states. The melt may remain interstitially between the grains, in which case the melt percolates upward between the grains due to its buoyancy. The melt may then accumulate in a sill-like magma chamber situated above the partially molten region (fig. 1). The sill would then supply a feeder dike that conducts the melt toward the surface. The other is a state where the melt initially accumulates in conduits. These conduits could either be veins filled with melt or porous channels with a relatively high porosity. The vertical migration of the melt is then caused by the upward intrusion or flow of melt from one conduit to another situated above it. The conduits may supply a sill with melt that in turn supplies a feeder dike. Alternatively, a system of connected conduits could form a disperse source region that supplies a feeder dike directly. The present work attempts to evaluate the structure of the partially molten plume using a theoretical approach. The presence or absence of conduits is estimated by considering the relationship between melt flow and pressure using the differential equation for permeable flow.

Several experimental estimates have shown that the melt formed in partially molten lherzolite has a small wetting angle, which suggests that the mantle could become permeable at degrees of melting of 0.1 to 2 percent (Waff and Bulau, 1979; Riley, Kohlstedt, and Richter, 1990; Kohlstedt, 1992; Faul, 1997). The onset of permeability at small degrees of partial melting is suggested by the uranium series disequilibria according to Cohen and O'Nions (1993) and Chabaux and Allègre (1994). While the interstitial distribution of melt appears well known, the nature of the subsequent processes that lead to magma accumulation is still under debate. The field evidence from the peridotites of ophiolite complexes suggests that melt can accumulate due to both shear strain and compaction. However, as emphasized by Ceuleneer and Rabinowicz (1992), the small scale structures of ophiolite complexes and alpine peridotites provide evidence for various types of melt accumulation processes but do not show which of these types account for melt accumulation at great depths. Models for the initial stages of melt accumulation are therefore speculative and several hypotheses have been proposed. The driving force for accumulation has been suggested to be shear stress (Nicolas, 1986, 1989, 1990) or compaction due to the buoyancy of the melt (Waff, 1980; Maaløe, 1981; McKenzie, 1984; Riley, Kohlstedt, and Richter, 1990). The accumulation models can be divided into matrix percolation models, where the flow is entirely interstitial (McKenzie, 1984), and conduit models, where the source region consists of intercalated segregation

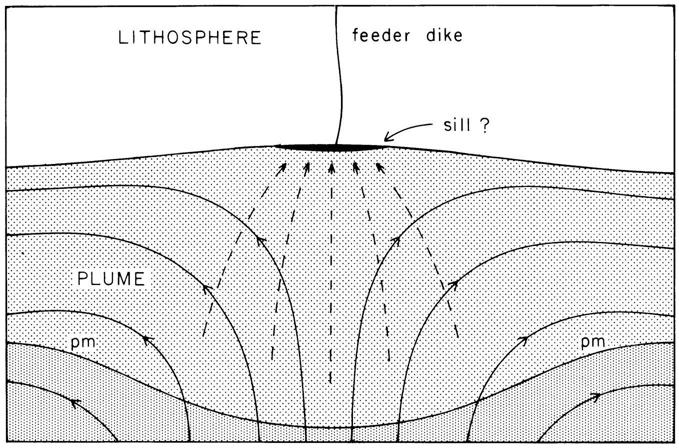

Fig. 1. The diagram shows a hypothetical cross section of a plume oriented perpendicular to the inferred long axis of the plume striking east-northeast west-southwest. The plume ascends and spreads out beneath the lithosphere. The solid curves show the flow of the plume. The onset of partial melting at depth is shown by the curve pm. The melt generated must somehow flow from the partially molten region toward a central feeder dike or sill as indicated by the stipled curves. This flow could occur either by interstitial pecolation or through conduits formed in the partially molten plume. The partially molten region is here modeled by a rectangular region as shown in figure 3.

veins and residuum (Wood, 1979; Shaw, 1980; Maaløe, 1981; Sleep, 1988; Ceuleneer and Rabinowicz, 1992; Ceuleneer, Monnereau, and Amri, 1993). A model based on field evidence from ophiolite complexes implies that the melt instead percolates through porous dunite channels (Quick, 1981). Theoretical models based on the dynamics of a porous matrix have suggested that the melt moves as solitons (Scott and Stevenson, 1984; Spiegelman, 1993; Richardson, Lister, and McKenzie, 1996). However, as mentioned by Fowler (1990) the soliton models ignore melting, and it is questionable if the soliton model can be applied to melt migration in mantle undergoing melting.

The percolation and conduit models differ in one important aspect, specifically the potential flow rate of the melt. The conduit models allow for fast flow rates, while models based on matrix and channel percolation must imply small flow rates. In principle the uranium series activity ratios should be able to define the transfer time from the mantle to the surface, but the interpretation of the activity ratios is presently model dependent. Flow of melt in conduits is suggested by Cohen and O'Nions (1993), Richardson and McKenzie (1994), and Chabaux and Allègre (1994). According to Spiegelman and Elliot (1993) and Lundstrom and others (1994) the formation of conduits is not required, interstitial percolation could also cause secular disequilibria with extraction times as long as 10^6 yr. Iwamori (1994) modeled both percolation and channel flow and concluded that the Hawaiian isotope activity ratios allow for both channel flow and percolation. The percolation models assume a porosity less than 0.2 percent. This in accordance with an experimental estimate by Kohlstedt (1992), but according to Faul (1997) the porosity required for permeability is more likely to be 2 to 3 percent.

Using widely different geochemical models, it has been shown that the trace element concentrations of accumulated melts have concentrations similar to those calculated for simple batch melting (Plank and Langmuir, 1992; Eggins, 1992; Maaløe, 1995), whereas the depletion of the residuum depends on the fractional character of the melting process. It was shown by Iwamori (1993) and Maaløe (1995) that equilibration between the ascending melt and the residuum must be limited, which suggests a fractional type of melting. The geochemical evidence for fractional melting may be related to the formation of conduits that allow for a more or less isolated flow of the melt.

A quite different approach is the dynamic one. The melt dynamics for Hawaiian plume sources were considered by Maaløe (1998a), who concluded that Hawaiian source regions may have permeabilities within the range 10^{-4} to 10^{-2} cm^2, which imply that their sources contain conduits. Alternatively, the source could be a sill situated in the upper plume, which is supplied with melt from below by percolation. The present work considers in addition the migration of melt in the entire partially molten plume from the onset of partial melting to the intrusion into a feeder dike. This dynamic approach involves assumptions but may nevertheless add useful constraints on accumulation models.

PLUME MODEL

The dynamic aspects of melt accumulation in a plume source region cannot be estimated without knowing parameters such as the time-integrated melt flux to a volcano and the dimensions of the partially molten region of the plumes. These parameters have been estimated or modeled with satisfactory accuracy for some of the Hawaiian shield volcanoes, and Hawaiian parameters are therefore used here. The tiltmeter measurements of the Kilauea volcano indicate that the supply rate of magma to this volcano is rather constant and about 0.067 km^3/yr (Dzurisin, Koyanagi, and English, 1984). The constant supply rate suggests that the melt generation of the Kilauean plume source is in a steady state, which furnishes important constraints on the melt dynamics of the source region. The nature of the melt accumulation processes that precede intermittent eruptions like the Hawaiian post-erosional ones is less restricted.

Numerical modeling of the mantle plume underlying the island of Hawaii suggests that the partially molten region is about 50 km high, and that the ascent rate is about 30 cm/yr so that the ascent time for a 50 km high source region is 183 ky (Watson and McKenzie, 1991). The partially molten region here was therefore modeled by a 50 km wide and 50 km high rectangular 2D cross section of the plume.

Taking the degree of partial melting of the mantle as 20 percent by weight for tholeiitic primary magma, the generation rate is calculated as $w = 1.143 \times 10^{-13}$ g/gsec (Maaløe 1998). The volumetric generation rate is similarly calculated as $v = 3.812 \times 10^{-14}$cm^3/cm^3 sec. The volumetric generation rate of melt within a 1 cm thick unit slice of the entire 2D partially molten region is 0.95 cm^3/sec or 30 m^3/yr, which is the volume of melt that should leave the source region in a steady state. This generation rate should be compared with the annual supply rate to Kilauea. The supply volume of 0.067 km^3/yr can be obtained if the length of a feeder dike is 22.3 km above a 2D-source region. This length appears somewhat excessive and may suggest that the melt is focused toward a central source region before it ascends.

The part of the plume that surrenders melt during an eruption is here called the source region in order to distinguish it from the entire partially molten part of the plume.

MODEL CONSTRAINTS

The melt structure of a partially molten plume and its variation in permeability with depth are unknown, but there are various constraints that could lead to an approximate estimate of the nature of magma accumulation. The constraints listed below are what is available and may appear rather arbitrary, but they are nevertheless useful.

1. *Buoyancy gradient.*–The buoyancy gradient is here taken as -30 bar/km. This presumes that the strain rate of the residuum is large compared to the generation rate of melt. According to the strain rates estimated by Hirth and Kohlstedt (1995) for an isostatically compressed olivine matrix containing melt this presumption is fulfilled. The

strain rate ϵ sec^{-1} is estimated from:

$$\epsilon = A\sigma^n d^p \exp\left(\frac{-Q}{RT}\right) \quad (1)$$

Where $A = 1.82 \times 10^{-6}$, a materials constant; σ = deviatoric stress in bars; n = 1 for diffusion creep; d = grain diameter in centimeters; p = −3 for diffusion creep; Q = 315 kJ/gmol °K, the activation energy; R = 8.314 joule/gmol °K; T = degrees Kelvin. For a grain diameter of 0.1 cm, T = 1873 °K, and a deviatoric stress of 1 bar, the strain rate is calculated as $\epsilon = 3 \times 10^{-12}$ sec^{-1}, and the viscosity as 1.1×10^{17} poise. In the stationary state the generation rate of melt is 3.8×10^{-14} sec^{-1}, which is nearly 100 times less than the strain rate. The flow rate of the melt is therefore controlled by the permeability and not the strain rate of the residuum. According to (1) the residuum can undergo compaction and expel the generated melt for a deviatoric stress of only 3×10^{-2} bar. In comparison, estimates of the grain sizes of subsolidus lherzolite suggest that the deviatoric stresses at 100 to 200 km depth and 1200° to 1500°C is from 30 to 50 bars (Ave'Lallement and others, 1980). It is therefore considered justified to assume the buoyancy gradient approaching −30 bar/km. A similar conclusion was obtained by Ribe and Smooke (1987) who considered the convection patterns of a plume and its melt.

2. *Overpressure.*–Melt pressure need to exceed least compressive stress to open a dike or a conduit. The overpressure P_o is defined as the difference between the pressure of the melt P_m and the lithostatic pressure P_l:

$$P_o = P_m - P_l \quad (2)$$

If P_o is positive the melt has an overpressure. If the interstitial melt situated in a residual matrix is unable to migrate because the residuum is impermeable then the pressure of the melt will be identical to the lithostatic pressure, so that $P_o = 0$. When the residuum becomes permeable the melt can move upward with a rate determined by the vertical pressure gradient and the permeability.

The modeling of the partially molten region is severely constrained by the condition that an ascending melt should have an overpressure at all depths. If the pressure of the melt is less than the lithostatic pressure then the melt is unable to ascend. Let the upper boundary of the partially molten region be at a depth y = 0 with y increasing downward, and the reference pressure at this depth be zero. The reference pressure is chosen as zero at y = 0, because it is the pressure differences that are important here, not the absolute value of the pressure. For a melt density of 3.00 g/cm^3 and mantle density of 3.3 g/cm^3 the pressure gradient caused by buoyancy is −30 bar/km (Boyd and McCallister, 1976; Scarfe, Mysen, and Virgo, 1979). If melt was able to flow with negligible resistance from a depth of 50 to 0 km the overpressure at y = 0 km would be 1500 bar. The mantle situated at 50 km depth therefore acts as a source with a melt pressure of 1500 bar relative to the melt situated at y = 0. Similarly a source at a depth of y = 20 km is a source with a source pressure of 600 bar. Note that at each of these depths the mentioned pressures of the melt are identical to the lithostatic pressure. Generally the source pressure is given by $\Delta P = 30y$. If the model calculations result in a source pressure less than 30y at depth y, then the pressure of the melt is less than lithostatic pressure and the melt is unable to ascend. The general condition for ascent is shown in figure 2, which forms a convenient diagram for the evaluation of various permeability models. If the calculated pressure of the melt is situated to the left and below the diagonal reference line given by $\Delta P = 30y$ ascent is impossible. Realistic pressure curves for the melt should be situated to the right and above this reference line.

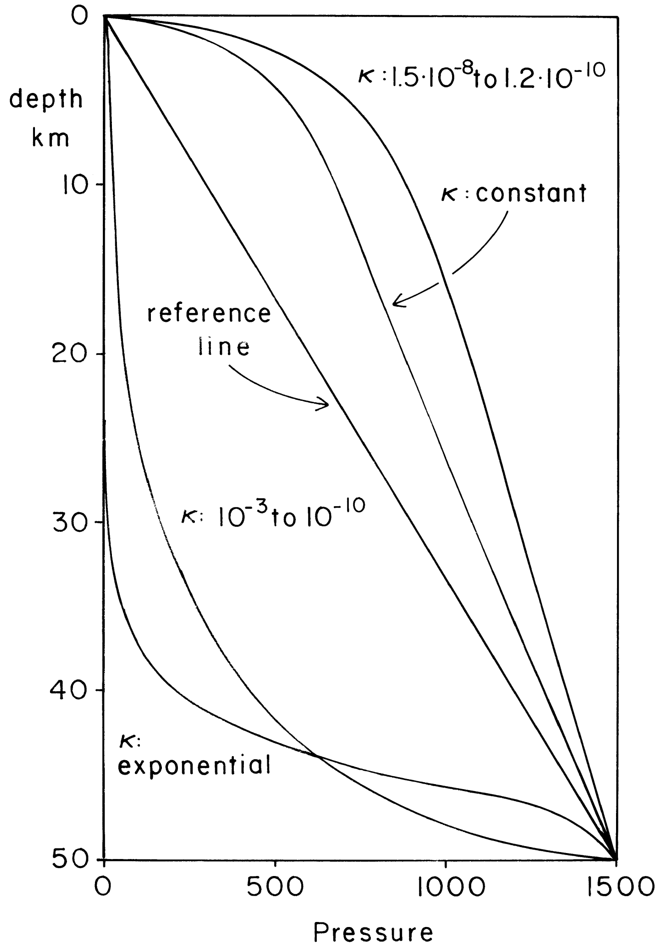

Fig. 2. The diagram shows the condition for overpressure of melts in the partially molten region. The reference line shows the increase in source pressure with depth given by $\Delta P = 30\,y$, where y is the depth. If the melt has an overpressure its pressure should be situated to the right of the reference line. The calculated pressures of melts for linear and exponential permeability variations from 10^{-11} to 10^{-3} cm^2 are less than the pressure of the reference line for all depths. The melts will therefore be unable to ascend. Constant permeability (κ: constant) and a linear variation in permeability from 1.2×10^{-10} to 1.5×10^{-8} cm^2 with decreasing depth (κ: variable) result in overpressure.

3. *Central focusing.*–The presence of shield volcanoes with a large single magma chamber and the seismic activity in the mantle beneath Kilauea suggest that the major part of the melt ascends through a feeder dike or system of feeder dikes (Klein and others, 1987). This suggest that the major part of the melt is focused into the central region of the plume. The modeling therefore assumes that the melt leaves the partially molten region in a central feeder dike.

The central calderas of the Hawaiian shield volcanoes are aligned in a northwest-southeast direction and the distance between the calderas is from 35 to 50 km. This distance is smaller than the perpendicular east-northeast west-southwest submarine extension of the islands of about 150 km (Watts, 1975). This suggests that the plumes have an elliptic cross section with thelong axis oriented in the same direction. The flow of the plumes in the upper asthenosphere is controlled by the presence of neighboring plumes, with the result that they spread out in a direction being perpendicular to the line joining the plume centers. It is therefore considered that the plume source can be modeled with acceptable accuracy by a 2D-model.

4. *Horizontal pressure variation.*–The pressure should decrease toward the central part of the partially molten plume so that the melt tends to flow from the sides of the plume toward its central part.

5. *Overpressure.*–The overpressure required for dike generation in the mantle is one of the most fundamental parameters for eruption dynamics but is poorly known. When the overpressure of a source region reaches a presently unknown threshold value, a dike will form above the source region. The overpressure along the roof of the source region increases with the height and permeability of the source region, so that the threshold value determines the maximal height of the source region.

Estimates of the shear strength of the asthenosphere based on seismic data suggest that the tensile strength of the asthenosphere could be within the range 10 to 100 bar (Shaw, 1980). Unfortunately, seismic data yield only evidence for the stress release during an earthquake, about 20 bar, and not evidence for the absolute stress (Wyss, 1970). The overpressures estimated for a generation of harmonic tremor beneath Kilauea are from 20 to 40 bar (Aki, Fehler, and Das, 1977; Chouet, 1981). Since the magmas intrude fractured lava flows and scoria, this range may appear of little relevance. However, the structure of layered intrusions show that they have a marginal border group which is partly consolidated during cooling of the intrusions (McBirney, 1995). The magma extruding at the Kilauea volcano stem from a magma chamber situated at 3 to 7 km depth, and the magma must first intrude through the upper border group of this chamber. The pressure of the intruding magma may either remain fairly constant or decrease with distance from the magma chamber, so that the range of overpressure from 20 to 40 bars may be considered somewhat lower than the initial overpressure required for intrusion of magma. An overpressure of about 100 bars may therefore approach the overpressure for intrusion into a rock containing interstitial melt.

The overpressure for dike formation could be different at high pressures. However, measurements of the grain size of lherzolite nodules suggest shear deformation at deviatoric stresses of about 40 bars at 100 to 200 km depth (Ave'Lallement and others, 1970). These measurements were made on lherzolite that equilibrated at temperatures less than 1400°C. Partially molten lherzolite generating tholeiite has higher temperatures. The mantle will therefore undergo deformation at a smaller deviatoric stress than 40 bars. The tensional strength of partially molten peridotite is not known, but the tensional strength of the refractory periclase containing melt is only 2 bar at 1400°C and 1 bar (Allison, Brock, and White, 1959). Much higher overpressure for dike generation in the mantle of 400 and 500 bar has been suggested by Nicolas (1990) and Kelemen and others (1997), respectively. It has been assumed here that overpressures in excess of about 100 bar result in dike formation.

6. *Melt flow rate.*–The generation rate in a 50 km × 50 km × 1 cm unit slice through the source region is $V = 0.95\ cm^3/sec$. For a steady state situation the flow of melt into a 1 cm wide slice of a feeder dike should be the same.

7. *Linear increase in melt flow rate.*–The flow rate of the melt should increase linearly with height for some distance above the onset of partial melting. The reason is that the flow of melt first becomes focused toward a feeder dike within the uppermost half of a partially molten region. The flow is therefore vertical within the lower part of this region.

For a steady state all melt generated beneath a certain depth should ascend past this depth. Thus if the height above onset of partial melting is z, then the volumetric flow rate should be

$$J = \nu z \tag{3}$$

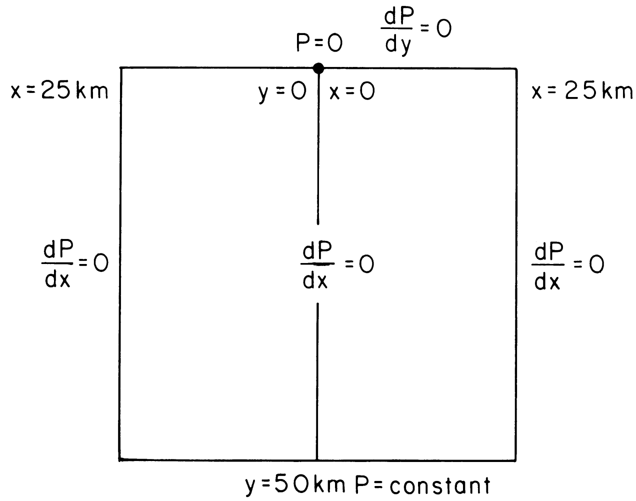

Fig. 3. The boundary conditions for a partially molten region. P is the source pressure, and x and y are the horizontal and vertical coordinates, respectively. The region is 50 km wide and 50 km high. Melt leaves the region through a central feeder dike situated at x = 0 and y = 0. No flow occurs across the roof so that dP/dy = 0. Also no flow occur across the sides, hence dP/dx = 0 at x = 25 km.

Using the melt generation rate $\nu = 3.182 \times 10^{-14}$ cm^3/sec, and Darcy's law and $\eta = 10$ poise, it is evident that the permeability of the lowermost part of the partially molten region must be about 10^{-10} cm^2, the porosity being about 0.5 percent after 1.5 km ascent above onset of partial melting. In comparison productive oil fields have porositys from 5 to 20 percent and permeabilities from 10^{-10} to 10^{-8} cm^2 (Levorsen, 1967).

8. *Tiltmeter measurements.*–The tiltmeter measurements of Kilauea from 1957 to 1983 and those of Mauna Ulu in 1969 show that the tilt increases within a day after the cessation of an eruption (Dvorak and Okamura 1987; Decker 1987; Koyanagi, Chouet, and Aki 1987). The instantaneous increase is only possible if the subvolcanic magma chamber is fed continuously from below as suggested by Eaton and Murata (1960).

STEADY STATE MODELS

The partially molten region of a plume is represented here by a rectangular region with a height of 50 km and a width of 50 km (fig. 3). The model assumes that all the melt leaves the region through a central feeder dike situated at x = 0, so that the flow across the roof of the source region is zero (fig. 3). There is no flow across the sides since no melt is generated outside the source region. The flow is symmetric with respect to the central plane through x = 0, so that the pressure gradient dP/dx = 0 for all values of y at x = 0. The pressure and flow rate variation within this region depends on the variation in permeability with depth. The purpose of the modeling that follows is to evaluate which permeability variation is consistent with the constraints listed above.

The transfer of melt toward the source region situated beneath a feeder dike is due to both the ascent of the plume and the ascent of melt within the ascending plume. The main concern here is the flow of melt within the plume, which implies that the ascent of the plume need not be taken into account. The flow rate of the melt relative to the plume may be estimated using Darcy's law:

$$J = -\frac{\kappa}{\eta}\left(\frac{dP}{dy} + \rho g\right) \tag{4}$$

where J cm^3/cm^2sec = flux of melt, κ cm^2 = permeability, η poise = viscosity, P bar = pressure of the melt, y cm = vertical coordinate, ρ = density of melt, and g = gravity constant. Note that J is the flux of melt. The flow rate of melt through the matrix is given by J/ϕ, where ϕ is the porosity. The permeability constant κ is normally applied for interstitial percolation. The permeability constant may also be applied for a source region with conduits, in which case the permeability is an apparent permeability of a partially molten region, the difference being that a source with conduits may have permeability's orders of magnitude larger than a percolative source. The melt flow for a steady state was estimated using the differential equation for permeable flow with a variable permeability with depth (Matthews and Russell, 1967):

$$\kappa_x \frac{\delta^2 P}{\delta x^2} + \kappa_y \frac{\delta^2 P}{\delta y^2} + \frac{\delta P \delta \kappa_y}{\delta y \delta y} + \frac{\eta}{\rho} A = 0 \qquad (5)$$

where κ_x and κ_y are the permeabilities in the x and y directions, respectively, P = pressure, η = viscosity, ρ = density, x and y the horizontal and vertical coordinates, respectively. The source term A represents the volumetric generation rate of melt. The effect of the source term is important as it increases the overpressure. Eq (5) was solved using variable permeability versus depth variations, and the numerical method of successive overrelaxation was applied for a 50×25 grid representing half the region (Strikwerda, 1989).

The variation in permeability with depth is not known, and different permeability depth functions may be considered. The supply rate of magma to the Kilauea volcano suggests a permeability of 10^{-2} to 10^{-4} cm^2 for the source region situated near the feeder dike if the source region supplies the feeder dike directly (Maaløe, 1998a). Since the permeability must be about 10^{-10} cm^2 for the source region at y = 50 km depth the pressure variation was estimated for a permeability variation from 10^{-10} to 10^{-3} cm^2 between 0 and 50 km, respectively. As is evident from figure 2, this variation results in pressures below the reference line and therefore below the lithostatic pressure at all depths, both for linear and exponential variations. The plume supplying the source region for the Kilauean lavas must have much smaller permeabilities. The pressure variation is also shown for a constant permeability in figure 2. A constant permeability results in melt pressures larger than the lithostatic pressure, but a constant permeability does not allow for a linear increase in the flow rate with decreasing depth in the lowermost part of the partially molten region.

A linear increase in the flow rate in the lower part of the partially molten region, as well as an overpressure of the melt, was obtained for $\kappa(50\ \text{km}) = 1.2 \times 10^{-10}$ cm^2 and $\kappa(0\ \text{km}) = 1.5 \times 10^{-8}$ cm^2, the variation in permeability with depth being linear (fig. 4A). These permeability values were obtained by iteration. The conditions being a linear increase in flow rate at depth and an overpressure at all depths along the center line, that is, for x = 0. The overpressure variation for this model is shown in figure 5. While the model fulfills the criterion for a linear increase in flow rate at depth it is evident that the overpressure in excess of 900 bar at y = 0 and $x \approx 25$ km is far too high. Even though the overpressure required for dike generation is poorly defined this overpressure seems excessive. Such large overpressures can only be attained within the duration of the shield stage, that is, about 1 my, if the permeability is much larger than 10^{-8} cm^2, and if the melt is retained within the plume. The ascent time for the melt is shown in figure 4B. The total ascent time is about 10^6 yr which is about the same as the duration of the shield activity. The ascent time is, thus, far too long. If a sill is present at y = 0 with widths exceeding 1 km then the overpressures would be smaller, but if the overpressure along the roof of the partially molten region is decreased then the pressure within its lower part becomes smaller than the lithostatic pressure. The pressure conditions thus appear to exclude the presence of a sill. If a sill is present its horizontal width must be less than 2 km. This estimate of the maximum width is based on the grid size adopted here where the distance between the grid points is 1 km.

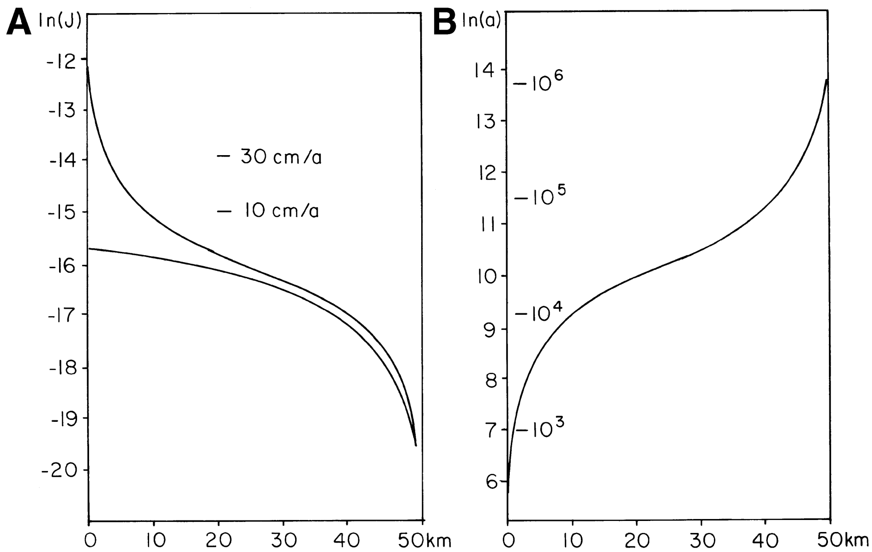

Fig. 4(A) The upper curve shows flow rate of melt in cm/sec for a linear permeability decrease with depth from 1.2×10^{-10} to 1.5×10^{-8} cm^2. The lower curve shows the flow rate for a constant permeability of 1.0×10^{-10} cm^2. The ascent rates for plumes of 30 cm/y and 10 cm/y are also shown. The calculated flow rates are smaller than the supposed ascent rates of the plume. (B) The period of time for ascent of melt generated at different depths in the central part of the plume. The ordinate shows the natural logaritm to the ascent time in years, while the ascent time is indicated in years inside the diagram. Melts formed in the center of the plume at its bottom have a total ascent time within the plume of about 1 my. This is a much longer period of time than the ascent of the partially molten part of the plume, which is estimated as 183.000 yr.

The melt may be expected to homogenizise in a sill, and homogeneous lava flows could suggest the presence of a sill. The lavas of the 1969 through 1971 eruption of Mauna Ulu were homogeneous when corrected for fractionation with respect to both major elements and isotope ratios (Hofmann, Feigenson, and Raczek, 1987). On the other hand, the lavas of the 1983 through 1985 flank eruption of Kilauea were heterogeneous, but this heterogeneity could be due to magma mixing in the magmatic system beneath the volcano (Garcia and Wolfe, 1988). In conclusion it appears difficult to estimate if a sill-like magma chamber is present in the upper part of the plume.

The present melt flow model shown in figure 5 differs from the model suggested by Ribe (1986). Both models assumes that melt from a permeable region is extracted from a central sink, but the present model includes melt generation. The implication of the Ribe (1986) model is that the porosity and permeability become nil near the sink after a miniscule amount of melt has been extracted, and it is therefore concluded that melt flow caused by pressure gradients cannot account for the accumulation of melts in a partially molten mantle. However, this situation arises when melt generation is omitted.

Excess overpressures appear to be intrinsic to migration models based on percolation and a stationary state, since other permeability variations with depth yield similar results. A stationary state can only be attained within 10^6 yr if the permeability is larger than 10^{-8} cm^2, which results in even higher overpressures. The excessive overpressures may be avoided if the melt either does not percolate at all or if it ascends discontinuously. If the melt resides where it is generated the melting is batch melting, but batch melting is inconsistent with the observed ($^{226}Ra/^{230}Th$) of Hawaiian tholeiites, which are within the range 1.1 to 1.2 (Cohen and O'Nions, 1993). Since the ascent of the plume takes 100,000 to 200,000 yr the melt will attain secular equilibria during ascent of the plume. Further, if the melt remained in situ it would follow the stream lines of the plume and not

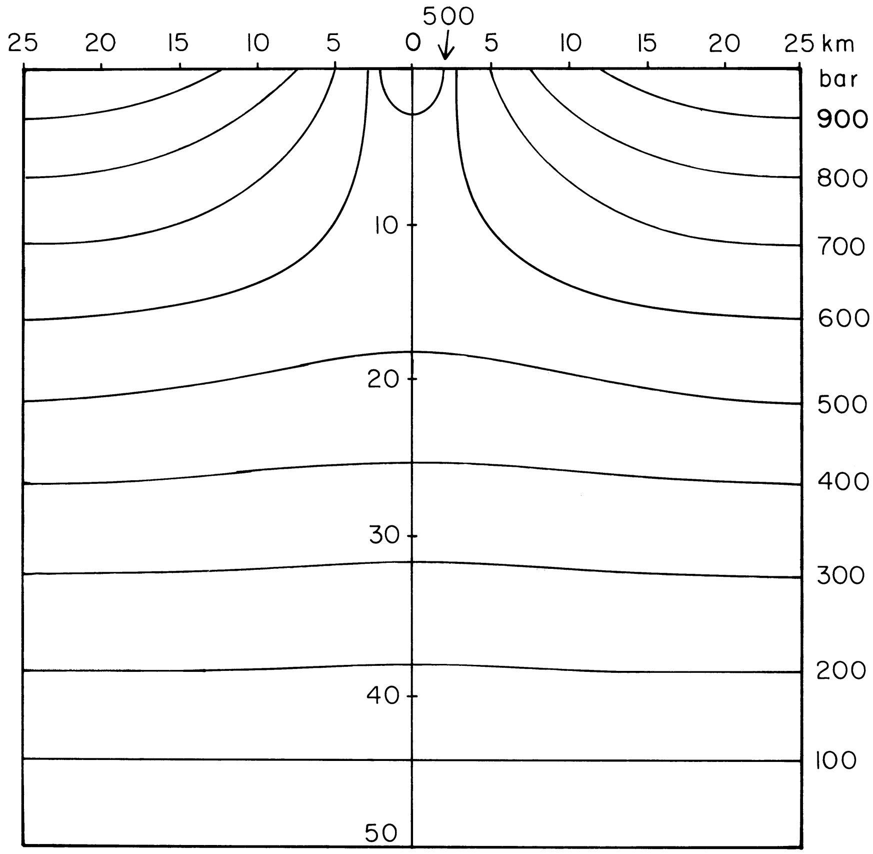

Fig. 5. The overpressures calculated for a steady state and a permeability variation of 1.2×10^{-10} at 50 km depth to 1.5×10^{-8} cm^2 at zero depth. The overpressure is in excess of 900 bar along the roof which is far too high. This overpressure would cause the formation of feeder dikes along the roof, and the melt would not leave the partially molten region through a central feeder dike.

accumulate at all (Ribe and Smooke, 1987). The melt must be able to flow relative to the plume for accumulation to occur.

It may be assumed instead that the melt after some ascent accumulates in conduits. By increasing degree of melting the frequency and size of the conduits increase with the result that they become interconnected. When the vertical extent of a local system of conduits has become large enough the melt intrudes upward (fig. 6). During ascent the conduit looses melt to the permeable matrix due to its overpressure whereby its ascent is arrested. During subsequent melting new melt is added to the conduit whereafter a new ascent takes place. This model is similar to the models suggested by Shaw (1980) and Sleep (1988). Similarly, the initial accumulation mode of melt favored by Ceuleneer and Rabinowicz (1992) and Ceuleneer and others (1993) from field evidence is a veined network. The detailed stress fields related to such conduit interaction has been considered by Takada (1994). It should be mentioned that the condition of discontinuous ascent does not by itself specify the nature of the conduits, which could be veins or porous channels.

The evidence for conduit formation is not that surprising because the variation in composition of peridotite nodules shows that the mantle is heterogeneous. The MgO

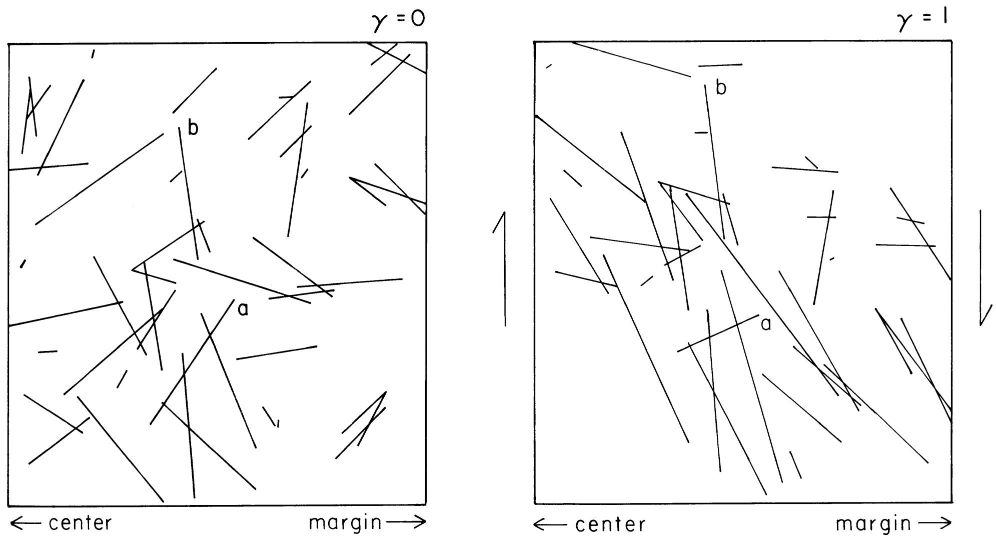

Fig. 6(A) A schematic diagram illustrating arbitrary oriented conduits. The orientation of conduits in a partially molten plume is not known. The diagram therefore shows melt conduits with arbitrary orientation and lengths estimated using the multiplicative linear congruential method. Melt of the local system beneath "a" may intrude upward to the next local system above "a". Similarly melt from the system beneath "b" may intrude upward to the next conduit above "b". (B) The diagram shows the same conduits as (A) that have undergone shear equal to 1.00, with left side up and right side down. The shear tends to tilt the layers away from the center of a plume that has the largest ascent rate, with the result that the ascent of melt is directed toward the center of the plume.

content of these nodules varies from that of lherzolite to harzburgite, that is, from 39 to 47 percent MgO (Maaløe and Aoki, 1977). The model shown in figure 5 assumes that the melt forms a continuous interstitial network throughout the partially molten region, which implies a homogenous lherzolite. Heterogeneous peridotite will result in local concentrations of melt and local variations in the permeability, which favor the formation of conduits.

The ascent rate of the central part of a plume with a radial decrease in temperature is larger than its marginal part. The conduits will therefore undergo shear with the result that conduits become tilted away from the center (fig. 6). This applies both for horizontal and arbitrary oriented conduits. The presence of conduits can therefore explain the focusing of melt toward the center of a plume (Maaløe, 1998b). Both horizontal and arbitrarely oriented conduits should be considered as melt accumulation in an heterogeneous mantle may result in an arbitrary orientation. A horizontal orientation would only occur if compaction occurs in a homogeneous mantle. The number of conduits must increase upward because the volume of melt increases with decreasing depth, and the conduits may become connected within the uppermost part of the partially molten region. The steady state supply to the Kilauea volcano shows that its plume source must be able to provide a continuous supply, so that melt can flow continuously from one conduit to the next one above. It is therefore suggested that the lower part of the source region contain conduits with melt that ascend periodically, while its uppermost part contain conduits that are continuously connected (fig. 7).

The conduits may become connected due to their increasing frequency, but connectivity may also arise in response to sheared flow. The upper part of the plume spreads out beneath the lithosphere, and this results in shear deformation of the plume (Ribe and Christensen, 1994). The field evidence from Ronda and other lherzolite

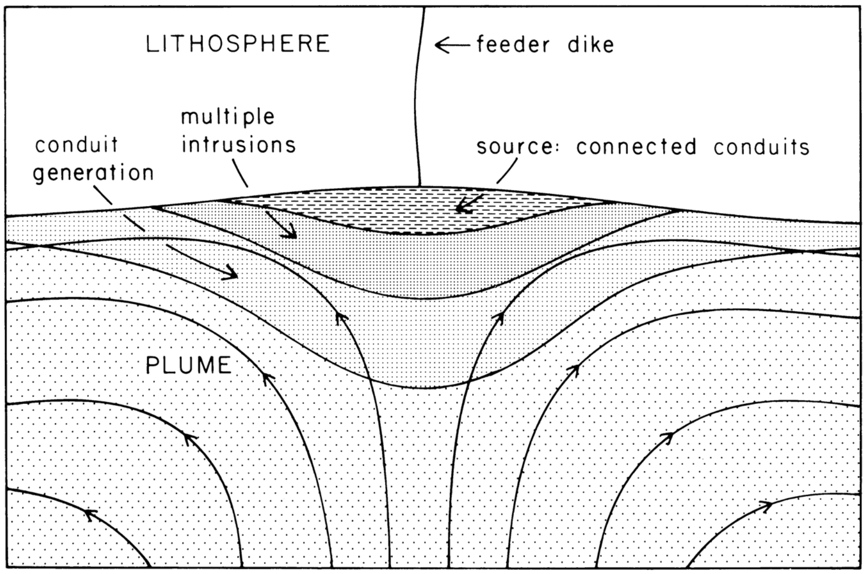

Fig. 7. A hypothetical sketch of a plume with conduits. Conduit formation begins after the plume has become permeable. At a higher degree of melting conduits have become more numerous and form local systems of conduits. When the systems of conduits has reached a certain height they begin to intrude upward due to the buoyancy of the melt. Melt therefore ascends by multiple mini-intrusions from one local system to others situated above. Finally, the conduits become continuously connected and form a source region for a feeder dike or a sill. The stream lines for the plume was adopted from Parmentier, Turcotte, and Torrance (1975).

massifs shows that shear deformation can result in the formation of melt layers from 0.1 to 5 cm thick (Dickey, 1970; Nicolas, 1990; and personal observation). It is thus possible that the divergent flow of the upper part of a plume also could result in formation of connected melt conduits.

The tiltmeter measurements of Kilauea and Mauna Ulu show that the magma supply must be continuous (Decker, 1987; Klein and others, 1987). The source region below the feeder dike must therefore be connected, and its permeability must be sufficiently large to generate the required supply rate. It is not possible to define the exact dimensions of such a connected source region, except that the overpressure along the roof must be less than the pressure generating dikes, here suggested as about 100 bar. As an example the height of the source region was here taken as 5 km. The isobars for such a source with a constant permeability of 10^{-4} cm^2 are shown in figure 8. The overpressure at the outer margin is nearly 115 bar. With these values the average pressure gradient for horizontal and vertical flow at the inlet of the feeder dike becomes 197 bar/km, and the flow rate of melt into the feeder dike 0.02 cm/sec. Since the melt production rate for a unit slice of the entire model source region is 0.95 cm^3/sec, a dike with a width of 47.5 cm can extract the melt produced.

Summarizing, the present modeling suggests the following steady state model for magma accumulation in a plume. After some initial melting the residuum becomes permeable, and at a higher degree of melting conduits begin to form. Initially the conduits are isolated from each other. As melting proceeds their number and size increase, and at some stage the conduits begin to become interconnected. When a local system of conduits has reached a certain vertical extension the overpressure becomes so high in the uppermost conduit that melt starts to intrude upward into peridotite or

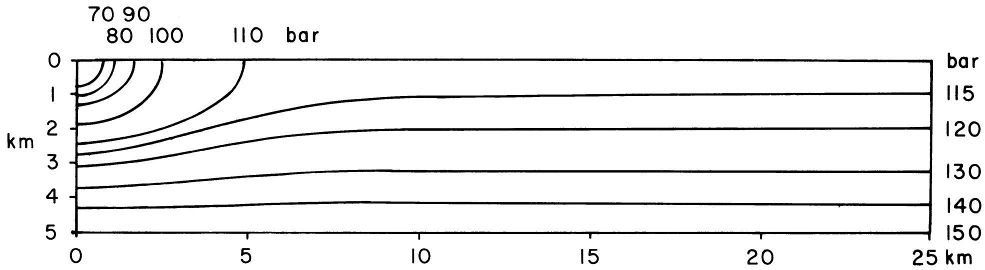

Fig. 8. The isobars for a 5 km high steady state source with connected conduits that is 50 km wide. The diagram show the right half of this source. The melt flows perpendicular to the isobars. Melt situated afar from the feeder dike would mainly move vertically toward the roof and thereafter horizontally toward the feeder dike. The source with connected conduits has here been assumed to be 50 km wide, but its width is not known and could be less.

another system of conduits situated above. The upward migration of the melt is thus envisaged to occur by multiple mini-intrusions. In the uppermost part of the source region the conduits become continuously connected due to an increasing frequency of conduits or increasing shear deformation, so that melt situated in the source region just below its roof attains an overpressure, which ensures that melt can form a feeder dike.

GEOCHEMICAL ASPECTS

The Hawaiian post-erosional lavas contain eclogite nodules, which indicate that melt veins can form in the mantle beneath Hawaii. However, the nodules do not provide an indication of the general structure of a partially molten plume. It is therefore relevant to consider the evidence from ophiolite complexes.

A general feature of the Oman ophiolite is the scarcity of intergranular melt relicts formed at depths greater than 10 km (Hopson and others, 1981; Ceuleneer and Rabinovicz, 1992; Ceuleneer and others, 1993). This may suggest that the width of the conduits is small. The average flow rate in a channel with parallel walls is given by (Kay and Nedderman, 1974):

$$u = \frac{-\Delta P b^2}{3\eta\Delta l} \tag{6}$$

where ΔP = pressure difference, b = half width of the channel, η = viscosity, and l = the length of the channel. Here $\Delta P/\Delta l = -30$ bar/km. Using eq (6) it is evident that the melt produced by a steady state can be extracted even when the conduits are veins a few millimeters wide. By the generation of tholeiitic melts and fractional melting of the residuum, the final interstitial melt would have harzburgitic composition. The melt left behind in the conduits could homogenize by diffusion with this interstitial melt so that the consolidated harzburgite becomes homogeneous, leaving virtually no trace of the initial melt accumulation. Subsolidus flow and recrystallization of the residual harzburgite after melt extraction could obliterate minor structures related to melt extraction. If this explanation applies it may be concluded that the thickness of the conduits must be small. If the melt conduits generally attained thickness up to a meter, then residual ophiolitic harzburgite should contain numerous deformed melt structures, but they are present only in small amounts in the Samail ophiolite (Hopson and others, 1981) and absent in the Trinity ophiolite (Ceuleneer and Rabinowicz, 1992).

The consequence of thin conduits being less than a few centimeters wide could be that the ascending melt to some degree equilibrates with the residuum. The ascending melt will both become mixed with melt generated from a depleted residuum and react with the residuum surrounding the conduits. The result could be that the composition of the melt approaches that formed by equilibrium melting. Its composition will depend on the period of time the melt remains stationary in a given depth (Bedard, 1989). A theoretical analysis of melt conduits in a porous matrix by Richardson, Lister, and McKenzie (1996) suggests that fast flowing melt should not interact greatly with the surrounding matrix, but a slow flowing melt might. Modeling of the chondrite normalized rare earth patterns of the tholeiitic and alkali olivine basaltic lavas from Mauna Kea suggest that the best fitting model is batch melting (Feigenson, Patino, and Carr, 1996), the precondition being that the source has a horizontal chondrite normalized rare earth element pattern.

INTERMITTENT ERUPTIONS

The eruption frequency of the Kilauea volcano has been about 1 per yr. The eruption frequency of the older shield volcano Mauna Loa volcano has been one in 4 yrs (MacDonald and Abbot, 1970; Clague, 1987; Wright and Clague, 1989). The alkalic post-shield lavas of Mauna Kea have erupted about once in 4000 yr. The time intervals between the post-erosional eruptions on Oahu appear to be 10,000 to 20,000 yr. There is thus a considerable decrease in eruption frequency with time from the shield stage to the post-erosional stage. The duration of the post-shield and later eruptions is only known for the 1800 to 1801 eruption of Hualalai that lasted about 1 yr.

The long repose times between eruptions may be related to two features. Firstly, it takes an extended period of time before a large volume of melt accumulates in the source region. Secondly, the overpressure in the source region may have to attain a certain threshold value before a feeder dike forms. The overpressure is considered to increase with the volume of melt accumulated, because the height of the source region is expected to increase with the accumulated volume.

The total upward transfer rate of melt T toward a source region situated beneath the lithosphere is given by the sum of the ascent rate of the plume u multiplied with the fraction of melt f and the average flow rate of the melt J relative to its residuum:

$$T = fu + J \tag{7}$$

If the matrix surrender melt due to shear deformation caused by the divergent flow of the plume, then the matrix may contribute a fraction q of its melt so that the amount of accumulated melt A is given by:

$$A = qfu + J \tag{8}$$

The relative magnitude of f u and J may be estimated by considering a unit column extending from $z = 0$ to $z = h$. All melt from the column is accumulated at $z = h$. For an ascent rate u the total ascent time becomes $t = h/u$. The melt generation rate per unit volume and unit time is $m = f/t$. The volume of matrix passing $z = h$ in a year is u so that the amount of melt M accumulated due to the ascent of the plume matrix is given by:

$$M = qfu \tag{9}$$

Assuming a constant porosity in the column at a given depth the melt flux J from the column is:

$$J = hm = \frac{hf}{t} = fu \tag{10}$$

so that:

$$A = fu(1 + q) \tag{11}$$

For a constant ascent rate of the plume the volume of melt accumulated in a source region increases linearly with time. The overpressure of the melt in the source region will also increase, but its increase with time will to some degree depend on the shape of the source region. The long repose times can therefore be related to a slow rate of melt generation in a slowly ascending plume. The constraint of a steady state cannot be applied for intermittent eruptions, and it is therefore not possible to distinguish between percolative ascent and ascent in conduits.

In theory the melt could remain in situ where it was generated until shear deformation in the mushrooming plume caused melt accumulation. If the accumulation period of time is relatively short compared to the total ascent time the melting is essentially batch melting. The percentage of melting of melilitite was estimated at 4 percent (Maaløe, 1994). Using this result the percentage of melting of alkali olivine basalt is 10 percent, so that $f = 0.1$. The activity ratio $(^{226}Ra/^{230}Th)$ has not been measured for Hawaiian alkalic lavas but is about 1.3 for other alkalic lavas (Chabaux and Allègre, 1994). Using the batch melting equations for Ra and Th, one has

$$D_{Ra} = \frac{D_{Th}}{r} + \frac{f(1 - r)}{r(1 - f)} \tag{12}$$

where $r = (^{226}Ra/^{230}Th)$ and f = fraction of melting. The source has secular equilibria, hence the initial activity ratio of the source is 1.00. For $r = 1.2$ and $f = 0.1$ the right side of (10) gets negative for $D_{Th} < 0.022$, so that D_{Th} must be larger than 0.022, which is about 10 times larger than experimental values (Lundstrom and others, 1994). This suggests that batch melting cannot explain the observed activity ratios and that melt migration relative to the plume matrix must be involved.

In principle it could be possible to estimate the volume of melt in a source region from the secular disequilibrium of the uranium series. The larger the volume of the melt the larger the mean storage time would be. Consequently the activity ratio $(^{226}Ra/^{230}Th)$ of erupted lavas would decrease with increasing volume of the melt. However, since the initial ratio is unknown it is not possible to obtain an accurate estimate. Taking $(^{226}Ra/^{230}Th)$ of the melt entering a source region as 2.00, the maximum volume of the melt accumulated is estimated as three times the erupted volume for a repose time of 4000 yr and a ratio of 1.3 of the erupted magmas. This ratio is representative for alkalic basalts (Chabaux and Allègre, 1984). However, the volume of the melt could be nil after an eruption or somewhat larger than three times the erupted volume if the assumed initial ratio is larger than 2.00.

Using a dynamic approach it appears difficult to estimate if the source region is a disperse one or a sill. There is indirect evidence that the source in some cases is not a sill but a disperse source with conduits. The major element trend of the post-erosional lavas on Kauai cannot be related to fractionation but is consistent with partial melting, which suggests that their compositions are primary. The lava flows are heterogeneous on a meter scale (Maaløe and others, 1992). It appears inconceivable that such heterogeneity could be preserved in a sill that accumulates melt for some hundred or thousand years. Even though thermal convection is absent due to small temperature gradients, the influx of melt must cause some stirring in a sill. The heterogeneity could rather be related to a disperse source with conduits where the melt can reside insulated from other melt. Presently, it is not known if other Hawaiian post-erosional flows display the same heterogeneity. If other Hawaiian post-erosional lava flows are homogeneous then it is possible that the melt of their sources was accumulated in a sill before it ascended through a feeder dike.

CONCLUSIONS

For steady state eruptions like the Kilauean ones, it appears that the melt accumulates in local systems of conduits and ascends episodically. It is considered that the ascent occurs from one system of conduits to others situated above, as suggested by Shaw (1980). It is suggested the melt ascends by such multiple mini-intrusions toward a source region with a connected network of conduits situated within the upper part of the plume. The formation of conduits can also explain the focusing of melt into a central source region within the plume. Since melt is generated interstitially, initial melt migration must occur by percolation. The degree of melting at which conduits are formed has not been estimated. According to Iwamori (1994) ascent in conduits is in accordance with observed Hawaiian isotope disequilibria, but these can also be explained by porous flow.

The small accumulation rates preceding the intermittent eruptions with long repose times suggest that the upward transfer of melt occurs at a low rate due to both a small ascent rate of the plume and a small ascent rate of melt.

The present modeling suggests that conduits form in a partially molten plume, but the model does not distinguish between veins and porous channels. The presence of composite mantle derived nodules in alkalic lavas and dunite channels in ophiolite complexes show that both types of conduits exist in the mantle (Harte, Hunter, and Kinny, 1993; Kelemen, Shimuzu and Salters, 1995). It is possible that a better understanding of the isotopic secular disequilibria will make it possible to determine which of these two types of conduits is most pertinent for segregation processes in a plume.

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