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Mantle heterogeneity beneath oceanic islands: some inferences from isotopes

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Radiogenic isotopes in oceanic basalts are extremely useful as tracers of long-lived heterogeneities in the Earth's mantle. Helium isotopes provide unique information in that high $^3\text{He}/^4\text{He}$ ratios are indicative of relatively undegassed mantle reservoirs (i.e. mantle with high time-integrated $^3\text{He}/(\text{Th} + \text{U})$ ratios). An alternative hypothesis is that high $^3\text{He}/^4\text{He}$ ratios may have been produced by ancient melting events, if the solid/melt partition coefficient (K_d) for He is greater than that for Th and U (i.e. yielding relatively high $\text{He}/(\text{Th} + \text{U})$ in the residue of melting). However, the distribution of helium within basaltic phenocrysts, and olivine/glass helium partitioning within mid-ocean ridge basalts, suggest that helium behaves as an incompatible element during melting (K_d (olivine/glass) < 0.0055), which strongly supports the hypothesis that high $^3\text{He}/^4\text{He}$ ratios are derived from undegassed mantle reservoirs.

Isotopic measurements of He, Sr, and Pb in Hawaiian volcanoes lavas demonstrate that the mantle sources have changed on extremely short timescales, between 100 and 10000 years before present. The preferred explanation for these variations is that they represent heterogeneities within the Hawaiian mantle plume, combined with late stage melting in the lithosphere for post shield alkali basalts. Helium isotopic data from Kilauea, Hualalai and Mauna Loa suggest that the plume is presently located beneath Kilauea (and Loihi seamount), and constrain the melting zone of the Hawaiian plume to be less than 40 km in radius.

1. Introduction

Basalts erupted at oceanic islands provide important constraints on the nature and origin of heterogeneity within the Earth's mantle. Isotopic variations between oceanic islands demonstrate that the mantle is heterogeneous on various length scales, and that the heterogeneities have persisted for at least 10^9 years (Hofmann & Hart 1978). However, it is not possible to infer accurately the geometry of mantle isotopic heterogeneities because basaltic lavas do not retain information about the original depth of their sources; at best, one can infer the depth of melt segregation. Because geochemical heterogeneities have persisted in the mantle on 10^9 year timescales, many geochemists favour layered mantle convection models (e.g. Allegre & Turcotte 1985). However, whole mantle convection models can also allow such long-lived heterogeneities, if for example the lower mantle is not well stirred (Davies 1984; Gurnis & Davies 1986). Although the geochemical data are important constraints, they alone cannot distinguish between these two distinct models for mantle structure (Silver *et al.* 1988).

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Helium isotopic data are important because high $^3\text{He}/^4\text{He}$ ratios in basalts are generated by high $^3\text{He}/(\text{Th} + \text{U})$ ratios in the mantle, and thus must be derived from the least degassed mantle. However, some recent laboratory partitioning experiments indicate that the helium solid/melt partition coefficients could be higher than for Th and U. If so, high $^3\text{He}/^4\text{He}$ ratios could be produced by early melting events in the mantle, which would leave the residue of melting with high $\text{He}/(\text{Th} + \text{U})$ ratios. High $^3\text{He}/^4\text{He}$ ratios cannot be produced by recycled materials, such as subducted oceanic crust sediments or ancient continental lithosphere, which have low $^3\text{He}/(\text{Th} + \text{U})$ ratios and hence low $^3\text{He}/^4\text{He}$ ratios (Mamyrin & Tolstikhin 1984). The simplest way to explain a relatively undegassed terrestrial reservoir is to assume that it has remained far from the surface. Indeed, most studies of ^3He flux from the mantle require some source within the lower mantle (O'Nions & Oxburgh 1983; Kellogg & Wasserburg 1990). Many oceanic islands, such as Hawaii, Iceland, Reunion, and Samoa have $^3\text{He}/^4\text{He}$ ratios higher than MORB, and based on helium alone, these islands would be derived from the least degassed mantle sources, which could reside in the lower mantle. The helium solid/melt partitioning is critical to these inferences.

Several processes may complicate such simple interpretations of isotopic data. Recent fluid dynamical studies have emphasized the importance of entrainment processes within the mantle (Griffiths & Campbell 1990; Griffiths & Campbell 1991). Entrainment of surrounding mantle can result in significant heterogeneity within individual mantle plumes. In addition, there is evidence that small melt fractions are rapidly removed from the mantle, suggesting that many basaltic liquids are produced by fractional melting (McKenzie 1985; Richter 1986; Riley & Kohlstedt 1991). Therefore, the basaltic liquids that are extruded on the Earth's surface may be hybrids of many different melt fractions. Interaction between the rising melt and the mantle may also result in geochemical changes resulting from 'percolation' effects (Navon & Stolper 1986; McKenzie & O'Nions 1991). If the geochemistry of erupted volcanic rocks is to be effectively used as a 'window into the mantle', criteria must be developed to distinguish between effects caused by melting processes and those related to source characteristics. The hypothesis advanced here is that one important way to constrain this problem is with data from individual volcanoes, where geological and temporal constraints are available. The emphasis here is on the isotopes of helium, because of the unique constraints high $^3\text{He}/^4\text{He}$ ratios can place on mantle derived rocks, and because of the large helium isotopic variability that is found within individual volcanoes.

2. Helium partitioning

Although there have been few studies of crystal-melt noble gas partitioning, most workers in this field have tended to assume that helium is an incompatible element within mantle minerals (Kurz *et al.* 1982*a*; McKenzie & O'Nions 1991). However, several laboratory equilibration studies have yielded noble gas crystal/melt partition coefficients significantly higher than Th and U and other incompatible trace elements (Hiyagon & Ozima 1986; Broadhurst *et al.* 1992). Hiyagon *et al.* (1986) found helium olivine/melt partition coefficients close to 0.1, with higher values for the heavy noble gases. If helium is a mildly compatible element ($K_d \approx 0.1$), then melting could conceivably leave the residue of melting with higher $^3\text{He}/(\text{Th} + \text{U})$ ratio, because the

Table 1. *Olivine/glass partitioning for MORB glass sample ALV 526-1*

(This sample is an olivine rich glass (*ca.* 30% olivine), and approximately 1% vesicles; further petrographic and geological details can be found in Bryan & Moore (1977).)

glass ^4He concentration ^a	3.117×10^{-6} cc (STP) g ⁻¹
olivine ^4He concentration ^a	2.875×10^{-8} cc (STP) g ⁻¹
olivine ^4He within inclusions ^b	1.136×10^{-8} cc (STP) g ⁻¹
$K_d(\text{ol/melt}) \leq \frac{(^4\text{He}_{\text{olv}} - ^4\text{He}_{\text{olv,crush}})}{^4\text{He}_{\text{glass}}}$	
$K_d(\text{ol/melt}) \leq 0.0058$	

^a Total concentrations for olivine and glass were obtained by melting of 0.5 to 2 mm grains in an ultra-high vacuum furnace.

^b Helium contained by inclusions is inferred to be that released by crushing, although this is a minimum because some helium may also be dissolved within the glassy inclusions (see text).

Th and U crystal/melt partition coefficients are very low ($\lesssim 10^{-3}$ (La Tourrette *et al.* 1992)). As a result, mantle reservoirs that were melted early in Earth's history could have high $^3\text{He}/^4\text{He}$ ratios, which would invalidate the assumption that high $^3\text{He}/^4\text{He}$ ratios indicate undegassed mantle sources.

There is some evidence that helium is an incompatible element during partial melting. The glassy rims of mid-ocean ridge basalts typically have helium concentrations of 10^{-5} – 10^{-6} cc(STP) g⁻¹ (Kurz & Jenkins 1981; Kurz *et al.* 1982*b*; Sarda & Graham 1990). Basaltic phenocrysts have significantly lower helium concentrations, typically 10^{-8} cc(STP) g⁻¹ (Kurz *et al.* 1982*a*), which would imply bulk partition coefficients less than 0.01. Unfortunately, large olivine phenocrysts are rare in MORB, and most of the phenocryst concentration measurements have been performed on ocean island basalts, where there are few other gas containing phases.

Table 1 lists new helium measurements on coexisting olivine and glass from the FAMOUS area at 37° on the mid Atlantic ridge. The concentration data provided by this experiment suggest that the olivine/glass partition coefficient (K_d) is significantly less than 0.0058. This value is an upper limit for several reasons. First, even submarine glasses are known to have undergone some degassing before, or during, eruption on the sea floor (Kurz & Jenkins 1981; Sarda & Graham 1990); hence the measured helium concentration in the glass must be considered to represent a *lower* limit estimate for the magma at eruption. On the other hand, the concentration in the olivine is an *upper* limit due to the presence of melt inclusions within olivine. Even though the calculation of the partition coefficient (see table 1) includes a correction for the helium released by crushing, there could still be considerable amounts of helium trapped within the glass inclusions that would not be released by crushing in vacuum.

There is also evidence, from the distribution of helium within olivine phenocrysts, that helium is incompatible in the olivine matrix. Most of the helium contained by olivine is released by crushing *in vacuo* (Kurz *et al.* 1982*a*), which implies that the helium resides within melt or fluid inclusions. This is illustrated by figure 1, which shows the helium distribution within olivine crystals from a number of Mauna Loa (Hawaii) lava flows. These data were obtained by crushing the samples in *vacuo*, followed by melting of the powder (Kurz *et al.* 1990). In all cases, most of the helium is released by crushing, as indicated by ratios of $^4\text{He}_{\text{olv}}/^4\text{He}_{\text{tot}}$ always significantly less than unity. The data define a trend whereby higher total concentrations are

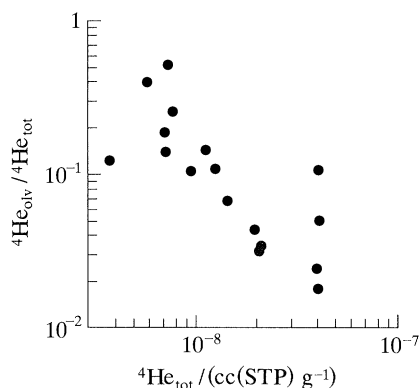


Figure 1. Partitioning within olivine crystals from Mauna Loa Volcano (Kurz *et al.* 1987, 1990). The total concentrations reflect the sum of helium extracted by crushing and melting of 1–2 mm olivines *in vacuo*. The ratio of $^4\text{He}_{\text{oliv}}/^4\text{He}_{\text{tot}}$ is the fraction that is contained by melt inclusions; $^4\text{He}_{\text{oliv}}$ is the amount of helium obtained by melting the previously crushed olivine powder, and $^4\text{He}_{\text{tot}}$ is the sum of crushing and melting the olivines. This fraction decreases with increasing total concentration, and is inferred to be related to abundance of melt inclusions (see text). Note that all of the samples have $^4\text{He}_{\text{oliv}}/^4\text{He}_{\text{tot}}$ significantly less than one, which suggests that helium is not contained significantly by the olivine matrix, but is held within the inclusions. The variability in the $^4\text{He}_{\text{oliv}}/^4\text{He}_{\text{tot}}$ ratio at high total ^4He concentration is primarily related to incomplete crushing of several samples.

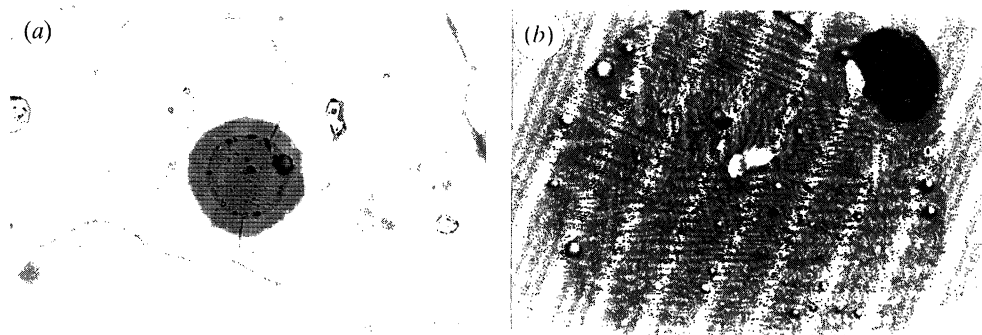


Figure 2. Photographs of an olivine grain containing a large melt inclusion exposed on a polished surface, from the 1868 lava flow of Mauna Loa, in transmitted and reflected light. (a) The spherical melt inclusion in the centre of the field is approximately 200 μm (width of field 1.2 mm, magnification 50 \times). The lighter area surrounding the inclusion is the olivine host crystal; small dark patches are smaller melt and spinel inclusions within the olivine. (b) This higher magnification photograph (field of view *ca.* 250 μm , magnification 250 \times) shows that the melt inclusion has undergone post-entrapment crystallization, as evidenced by skeletal pyroxene crystals. The large round dark area at the top of the field is a bubble, as are the smaller ones dispersed throughout the inclusion. The angular dark area near the centre of the inclusion is an ion microprobe crater.

accompanied by greater fraction released by crushing (i.e. lower $^4\text{He}_{\text{oliv}}/^4\text{He}_{\text{tot}}$). This suggests that high helium concentrations are associated with inclusions (which release helium upon crushing).

Microscopic examination of the olivine crystals supports the hypothesis that melt inclusions control the helium concentrations in olivine. Figure 2 shows photographs of one crystal containing melt inclusions, with an inclusion exposed on the polished surface. The photograph demonstrates that the melt inclusion contains microscopic bubbles which are the most likely site for helium and other gases. These bubbles

could be produced by post-entrapment crystallization (as suggested by the spherulitic crystal growth in the glass inclusion), or may exist before entrapment. The existence of a gas phase in these melt inclusions is probably related to confinement pressure on eruption, and speed of quenching, so they may not be present in all olivine crystals. Microscopic examination of the samples shown in figure 2 also suggests that fluid inclusions (i.e. containing liquid CO₂) are rare. Therefore, the features shown in figure 2, combined with the data in figure 1 strongly suggest that melt inclusions dominate the helium abundance in olivine phenocrysts, which is consistent with the interpretation that helium is an incompatible element in olivine.

Although the data presented here only provide an upper limit K_d for olivine/melt, they do illustrate that helium partition coefficients inferred from natural crystals are significantly lower than those found by laboratory experiments (Hiyagon & Ozima 1986). One explanation is that the laboratory experiments were also affected by the presence of microscopic inclusions, as discussed by Hiyagon & Ozima. Additional data will be necessary to resolve this discrepancy. Because all existing data from naturally formed olivines suggests that helium behaves as an incompatible element, it is assumed that high $^3\text{He}/^4\text{He}$ ratios relate to degassing history.

3. Global isotopic patterns

Isotopic variations in basalts are generally interpreted to reflect mixing between different reservoirs; various efforts to explain the isotopic variability in basalts have suggested the necessity of four or five distinct mantle sources (White 1985; Zindler & Hart 1986*a*; Hart 1988; Hart *et al.* 1992). Figure 3 summarizes most of the existing coupled He, Sr and Pb isotopic data for oceanic islands. Hawaii, Iceland and Samoa have the highest $^3\text{He}/^4\text{He}$ ratios, and by the reasoning adopted here, their lavas are derived from the least degassed mantle sources. The tholeiites from Loihi seamount, which have the highest $^3\text{He}/^4\text{He}$ ratios, have Sr and Nd isotopic compositions which differ significantly from those expected for bulk Earth. In fact, Hawaii is a geochemical end-member only with respect to helium; as illustrated in figure 3, data from Hawaiian basalts fall within the centre of the Sr–Pb diagram. Those oceanic islands having $^3\text{He}/^4\text{He}$ ratios lower than MORB values (i.e. lower than *ca.* $7 R/R_{\text{atm}}$), such as Tristan da Cunha, Gough and Sao Miguel, may be derived from mantle sources having significant contributions from recycled oceanic crust or sediments (Kurz *et al.* 1982*a*).

One problem that is well illustrated by figures 3 and 4 is that there are significant local variations in many of the oceanic islands. At Hawaii for example, two volcanoes Loihi and Mauna Loa, which are only 40 km apart, define most of the global variability in $^3\text{He}/^4\text{He}$ ratios, ranging from 8 to 32 times atmospheric. One explanation for this variation in helium, accompanied by small variability in Sr and Pb, is that helium is decoupled from the other isotopes by accumulation of ^4He during residence in a magma chamber (Condomines *et al.* 1983; Zindler & Hart 1986*b*), or by metasomatism in the mantle (Vance *et al.* 1989). However, in most cases where He, Sr, Nd and Pb are available for the same samples, there are correlations between them. This includes Haleakala and Mauna Loa on Hawaii (Kurz *et al.* 1987; Kurz & Kammer 1991), Samoa (Farley *et al.* 1992), Sao Miguel (Kurz *et al.* 1992). In Reunion basalts none of the isotopic systems, including helium, show any significant variability (Staudacher *et al.* 1990; Graham *et al.* 1990). There are few general correlations between any isotope systems in lavas from Iceland and

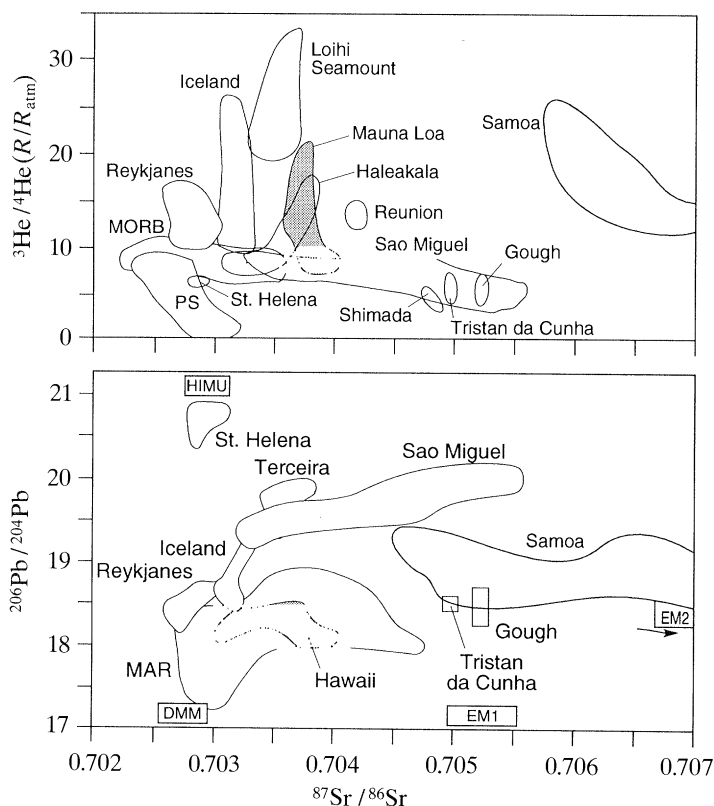


Figure 3. Helium, lead, and strontium isotopic data summary for selected oceanic islands (Kurz *et al.* 1982*a*, 1983, 1992; Farley *et al.* 1990; Graham *et al.* 1990; Poreda *et al.* 1986; and references in the text). PS refers to Pacific seamounts (Graham *et al.* 1988). Note that in the Sr–Pb diagram, Hawaiian basalts display a limited variation compared to other oceanic islands. The labels for distinct mantle reservoirs (HIMU), and enriched mantle (EM), proposed by Zindler & Hart (1986*a*) are shown for references.

Loihi seamount (Condomines *et al.* 1983; Staudigel *et al.* 1984), but this incoherence may relate to the lack of age information for most of the samples. Therefore, the systematic relationships between He, Sr and Pb in oceanic island basalts, particularly those from Hawaii, argue strongly against decoupling of helium from the other isotopes by magma chamber production of ^4He , or by metasomatism, and these processes will not be considered in any detail here (see Kurz & Kammer (1991) for discussion).

Because high $^3\text{He}/^4\text{He}$ ratios indicate undegassed mantle sources, helium is central to understanding the origin of mantle heterogeneity. Advocates of whole mantle convection have suggested that ^3He is coming from the core (Davies 1990), and others have used high $^3\text{He}/^4\text{He}$ ratios as indicators of a lower mantle origin for volcanism (Allegre *et al.* 1983; Hart *et al.* 1992). In evaluating such hypotheses, it is critical to evaluate the relationships between helium and the isotopes of Sr, Nd, and Pb, particularly with respect to small-scale heterogeneities, in order to discern possible mixing end-members.

4. The local context: inferences from Hawaiian volcanoes

There are many oceanic islands with large isotopic variations over small time or length scales, and a proper understanding of the causes of these variations is of fundamental importance. Because basalts from Hawaii have the highest $^3\text{He}/^4\text{He}$ ratios, understanding the local variations shown in figure 3 is particularly important to noble gas work. As one of the largest, longest-lived and presently most active hotspots, Hawaii, provides an important record of melting in the mantle. The fact that Hawaii constitutes the highest known mass flux at a hotspot (Sleep 1990) may suggest a relationship between high $^3\text{He}/^4\text{He}$ and material flux from the lower mantle. This relationship does not necessarily extend to other islands, because many of the islands having $^3\text{He}/^4\text{He}$ higher than MORB have low calculated mass fluxes (for example Iceland, Bouvet, and Reunion (Sleep 1990)). However, it is worth noting that Hawaii is the oceanic island that has both the highest mass flux and $^3\text{He}/^4\text{He}$ ratios.

At Hawaiian volcanoes it is well established that shield tholeiites are isotopically distinct from the post-erosional alkali basalts (Chen & Frey 1983, 1985). The Hawaiian shield building tholeiites have systematically higher $^3\text{He}/^4\text{He}$ than younger alkali basalts (Kurz *et al.* 1983; Kurz *et al.* 1987). These data are consistent with the mantle plume model for Hawaiian volcanism because they suggest an evolutionary trend, whereby the shield building stages are produced by melting of plume material with higher $^3\text{He}/^4\text{He}$ (Kurz *et al.* 1983; Kaneoka 1983; Kurz *et al.* 1987; Garcia *et al.* 1990). The later stages then are presumed to have $^3\text{He}/^4\text{He}$ similar to MORB due to a melting of MORB related upper mantle. The mechanism by which the later stages have different isotopic signatures, often presumed to result from melting of the lithosphere, could involve melt percolation because all the Hawaiian extrusives must pass through a substantial thickness of lithosphere (50–100 km).

An important aspect of the Hawaiian isotopic variations is the short timescale over which they occur at Mauna Loa volcano (Kurz *et al.* 1987; Kurz & Kammer 1991). Kurz & Kammer showed that the isotopic composition of Sr, Pb and He in Mauna Loa basalts decreased significantly 10 000 years ago, and there was another change in Sr and Pb isotopes approximately 600 years ago, with $^3\text{He}/^4\text{He}$ remaining essentially constant. The timescale is provided by geological mapping of historical and radiocarbon dated lava flows (Rubin *et al.* 1987). This complex isotopic evolution is demonstrated in figure 4*a* along with He and Sr data from some other Hawaiian volcanoes. The data from Mauna Loa provides the important time information, and demonstrates that alkali basalts and tholeiites, even from the same volcanic edifice, may not necessarily be connected by mixing processes, unless the sampling is on a short enough timescale.

There are two alternative explanations for the isotopic variations shown in figure 4*a*. Kurz & Kammer suggested changes in the source materials, as Mauna Loa migrates away from the hotspot. The isotopic end-members required by this model are shown sequentially in figure 4*b*, and include the plume, another distinct source, perhaps from a different part of the plume, and finally the lithosphere. Heterogeneity within the plume could be produced by entrainment processes (Griffiths & Campbell 1991). The lithospheric involvement becomes important only after the volcano is removed from the plume melting conduits, and some of the heat from the plume melts the lithosphere (Liu & Chase 1991).

McKenzie & O'Nions (1991) suggested that the isotopic variations within Mauna

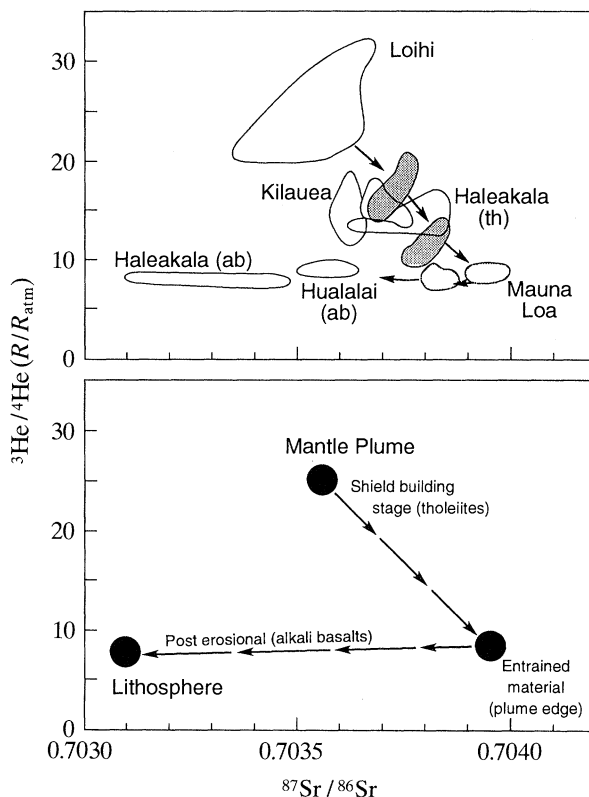


Figure 4. (a) Sr and He isotopes for Hawaiian basalts with the assumed temporal evolution pattern denoted by the direction of the arrows. The Mauna Loa basalts (shaded) are critical to this evolution scheme because the 'bend' in the evolution trend is observed between 600 and 2000 years before present (Kurz *et al.* 1983, 1987; Staudigel 1984; Kurz & Kammer 1991; and unpublished data, this laboratory). (b) Inferred isotopic compositions for the different components in the mantle beneath Hawaii that are required to explain the variability in the top figure. The highest $^3\text{He}/^4\text{He}$ ratios are inferred to be related to the mantle plume itself. The decrease in $^3\text{He}/^4\text{He}$ combined with increased $^{87}\text{Sr}/^{86}\text{Sr}$ is assumed to be related to heterogeneity within the plume, perhaps by entrainment affects that could lead to such axial variations (Griffiths & Campbell 1991). The lowest $^3\text{He}/^4\text{He}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, found in the alkali basalts, are inferred to be related to melting in the lithosphere.

Loa (figure 4a) could be explained by mixing between two mantle sources via a melt percolation model. In their model, helium is removed from the plume first due to its extreme incompatibility within silicates, while the other elements such as strontium are released into the melt more slowly. The predicted time evolution is not unlike that shown in figure 4a. In the context of the Hawaiian isotope data, there are several important problems with the McKenzie & O'Nions model. The model predicts that radiogenic isotopes and incompatible trace element abundances will decrease with time at any volcanic shield. As pointed out by Frey & Rhodes (this symposium), this prediction is at odds with existing Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and $^{143}\text{Nd}/^{144}\text{Nd}$ data for Kilauea and Mauna Loa. In addition, preliminary rare earth data for the Mauna Loa samples shown in figure 4a, does not display a systematic variation with time, as would be predicted by the model (Kurz & Shimizu, unpublished ion microprobe data for melt inclusions).

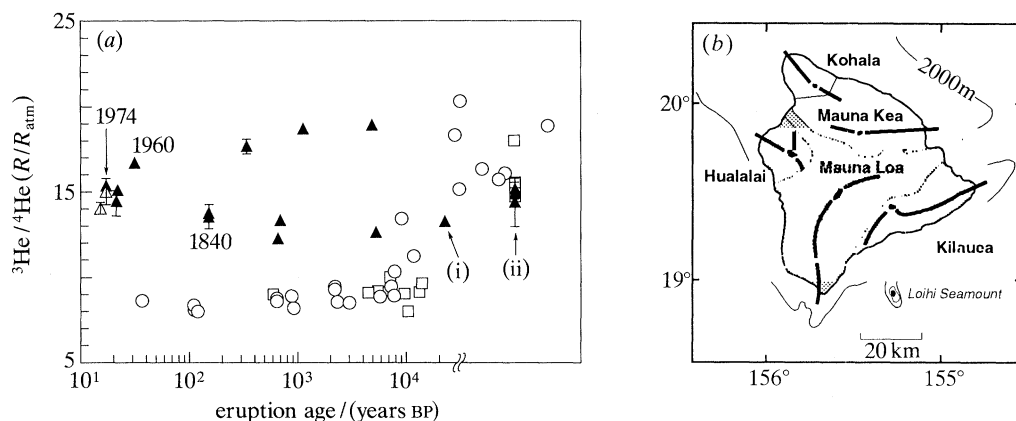


Figure 5. Helium isotopic evolution for three Hawaiian volcanoes, Hualalai (\square), Kilauea (\blacktriangle), and Mauna Loa (\circ). The timescale is provided by historical flows, radiocarbon dates on some lava flows (Rubin *et al.* 1987), and stratigraphy. There is no timescale beyond 30000 years because this is the age limit of the radiocarbon dates (Kurz *et al.* 1983, 1987, 1990; Kurz & Kammer 1991; and unpublished data this laboratory). Hollow triangles are volcanic gases from Kilauea (Craig & Lupton 1976; Jenkins *et al.* 1977). (i) Hilina series; (ii) dredged tholeiites.

Mauna Loa is the largest volcano on the island of Hawaii (and on Earth): an important question is whether the basalts erupted from the two smaller adjacent active volcanoes, Hualalai and Kilauea, have displayed similar isotopic variations. Figure 5 shows the temporal helium isotopic record from these two volcanoes (Kurz *et al.* 1990). As discussed earlier, the main feature of the Mauna Loa temporal record is the marked decrease in $^3\text{He}/^4\text{He}$ roughly 10000 years ago. For the Hualalai basalts, the temporal helium isotopic evolution is similar to Mauna Loa; the dredged tholeiites have higher $^3\text{He}/^4\text{He}$ than the historical alkali basalts (note that all surficial Hualalai eruptions are alkali basalts). As mentioned above, this is consistent with previous data from Haleakala (Kurz *et al.* 1987) and with the idea that the alkali basalts are derived from melting of MORB related lithosphere/asthenosphere, because the $^3\text{He}/^4\text{He}$ converge to MORB values (roughly $8 R/R_{\text{atm}}$ in figure 5).

During the past 10000 years, basalts from Kilauea have consistently higher $^3\text{He}/^4\text{He}$ ratios than the other two volcanoes (between 12 and $20 R/R_{\text{atm}}$). The consistently higher $^3\text{He}/^4\text{He}$ ratios suggest that Kilauea eruptions are more directly associated with the mantle plume, and therefore that the mantle plume contributes material (i.e. ^3He) to Kilauea, but not to Hualalai or Mauna Loa. This places a limitation on the size of the plume at approximately 40 km (see also Frey & Rhodes, this symposium). In addition, the data from Kilauea are significantly more variable, with helium isotopic differences between the 1840 and 1960 flows implying variability on the 100 year timescale. During the past 10000 years Mauna Loa and Hualalai basalts have had $^3\text{He}/^4\text{He}$ ratios which are indistinguishable from MORB values (ca. $8 R/R_{\text{atm}}$). There are several possible causes for the short-term helium isotopic variation in Kilauea basalts. Based on trace element abundance variations, Rhodes *et al.* (1989) proposed that there may be periodic invasion of the Kilauea plumbing systems by Mauna Loa magma. Because the helium isotopic composition of the two volcanoes is so distinct, this could yield a significant change in the isotopic composition. This is feasible because Mauna Loa magma batches are so much larger than Kilauea; Klein (1982) estimated that the Mauna Loa magma reservoir is

approximately five times larger than for Kilauea. This explanation for the variations at Kilauea would require that the magmatic transfer is essentially one way: from Mauna Loa to Kilauea.

An alternative interpretation for the Kilauea helium isotopic variation is that the fluctuations represent changes in the source chemistry over this timescale. For example, in the case of continuous magma supply, the McKenzie & O'Nions (1991) percolation model might be applicable to individual magma batches. If this is the correct explanation for the variations shown in figure 5, then a new magma batch is produced every 100–300 years in the mantle beneath Kilauea. Based on the magma supply rate to Kilauea (roughly $7 \times 10^6 \text{ m}^3 \text{ month}^{-1}$; Dzurisin *et al.* 1984), this would require individual magma batches of approximately 1 km^3 , which is roughly the predicted size of magma pulses (solitons) in the mantle (McKenzie & O'Nions 1991).

At present it is difficult to choose between the two hypotheses for the Kilauea helium isotopic variability. The relative sizes of the two volcanoes, coupled with the trace element evidence for magmatic exchange between them (Rhodes *et al.* 1989), suggests that magma mixing plays at least some role in the helium isotopic variability. Additional data will be required to confirm this. However, the data demonstrate that detailed studies of individual volcanoes may, in the future, place firm constraints on models for melting in mantle plumes.

In summary, the data presented in figures 4 and 5 demonstrate that isotopic variations within Hawaiian volcanoes can occur on extremely short timescales. With respect to the global isotopic compilations, the critical question is whether this type of variation is found at all ocean island volcanoes. Although there are too few isotopic studies of this kind, with the appropriate geological controls, to properly answer this question, existing data from other oceanic islands suggests that Hawaiian volcanoes are unique. Helium isotopic studies of Piton de la Fournaise (Reunion) demonstrated that the $^3\text{He}/^4\text{He}$ ratios (*ca.* $13 \times R_a$) in the erupted basalts have remained essentially constant for the past million years (Staudacher *et al.* 1990; Graham *et al.* 1990). Studies of the radiocarbon dated lava flows on Sao Miguel (Kurz *et al.* 1992), have demonstrated large variations over similar length scales as observed between Hawaiian volcanoes (*i.e.* *ca.* 20 km). However, in this case there is an isotopic gradient from one side of the island to the other, with little temporal control on the geochemistry of the basalts. There are many parameters which may differ between Reunion, the Azores, and Hawaii, including mantle composition, mantle structure, lithospheric thickness, depth of melting, and perhaps the extent of melting. The best constrained of these parameters is lithospheric thickness. Hawaii rests on relatively old oceanic crust and has thick lithosphere, which is circumstantial evidence that the lithosphere plays a key role in the temporal isotopic variations within Hawaiian volcanoes, as first suggested by Tatsumoto (1978).

5. Summary

New olivine/melt helium partitioning data presented here suggest that helium is an incompatible element with respect to melting, which contrasts with the conclusions of previous studies (*e.g.* Hiyagon & Ozima 1986). This supports the hypothesis that high $^3\text{He}/^4\text{He}$ ratios indicate undegassed mantle sources.

Data from the three Hawaiian volcanoes demonstrates how eruption age combined with isotope geochemistry can be used to place constraints on the melting dynamics beneath hotspots. A proper understanding of these local isotopic variations remains

an important aspect of interpreting global isotopic variations, particularly in the context of mantle convection models. Helium isotopic data indicate little contribution of mantle plume material to present day eruptions of Hualalai and Mauna Loa volcanoes. In the past, both volcanoes have displayed higher $^3\text{He}/^4\text{He}$ ratios, which suggests that they have recently been removed from the actively melting hotspot mantle beneath Hawaii. In contrast, Kilauea has markedly higher $^3\text{He}/^4\text{He}$ ratios over the entire time sequence of subaerial eruptions, and dredged basalts from Loihi seamount have the highest $^3\text{He}/^4\text{He}$ ratios of any oceanic rocks. These data suggest a significant contribution from undegassed, plume-type mantle to Loihi and Kilauea, and restrict the size of the actively melting plume to a radius less than 40 km (roughly the distance between Loihi seamount and Mauna Loa), and greater than 30 km (the distance between Loihi seamount and Kilauea) (see figure 5). The complex temporal evolution of these volcanoes, illustrated in figure 4, is attributed here to heterogeneity within the mantle plume.

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References

- Allegre, C. J., Staudacher, T., Sarda, P. & Kurz, M. 1983 Constraints on evolution of Earth's mantle from rare gas systematics. *Nature, Lond.* **303**, 762–766.
- Allegre, C. J. & Turcotte, D. L. 1985 Geodynamic mixing in the mesosphere boundary layer and the origin of oceanic islands. *Geophys. Res. Lett.* **12**, 207–210.
- Allegre, C. J., Staudacher, T. & Sarda, P. 1987 Rare gas systematics: formation of the atmosphere, evolution and structure of the Earth's mantle. *Earth planet. Sci. Lett.* **81**, 127–150.
- Broadhurst, C. L., Drake, M. J., Hagee, B. E. & Bernatowicz, T. J. 1992 Solubility and partitioning of Ne, Ar, Kr, and Xe in minerals and synthetic basaltic melts. *Geochim. cosmochim. Acta* **56**, 709–723.
- Bryan, W. B. & Moore, J. 1977 Compositional variations of young basalts in the Mid-Atlantic Ridge rift valley near 37 degrees N. *GSA Bull.* **88**, 556–570.
- Chen, C. Y. & Frey, F. A. 1983 Origin of Hawaiian tholeiite and alkali basalt. *Nature, Lond.* **302**, 785–789.
- Condomines, M., Gronvold, K., Hooker, P. J., Muehlenbachs, K., O'Nions, R. K., Oskarsson, J. & Oxburgh, E. R. 1983 Helium, oxygen, strontium and neodymium relationships in Icelandic volcanics. *Earth planet. Sci. Lett.* **66**, 125–136.
- Craig, H. & Lupton, J. E. 1976 Primordial neon, helium and hydrogen in oceanic basalts. *Earth planet. Sci. Lett.* **31**, 369–385.
- Davies, G. F. 1984 Geophysical and isotopic constraints on mantle convection: an interim synthesis. *J. geophys. Res.* **89**, 6017–6040.
- Davies, G. F. 1990 Mantle plumes, mantle stirring and hotspot chemistry. *Earth planet. Sci. Lett.* **99**, 94–109.
- Dzurisin, D., Koyanagi, R. Y. & English, T. T. 1984 Magma supply and storage at Kilauea Volcano, Hawaii. *J. Volcan. geotherm. Res.* **21**, 177–206.
- Farley, K. A., Natland, J. H. & Craig, H. 1992 Binary mixing of enriched and undegassed (primitive?) mantle components (He, Sr, Nd, Pb) in Samoan lavas. *Earth planet. Sci. Lett.* **111**, 183–199.
- Garcia, M. O., Kurz, M. D. & Muenow, D. 1990 Mahukona: the missing Hawaiian volcano. *Geology* **18**, 1111–1114.
- Graham, D. W., Zindler, A., Kurz, M. D., Jenkins, W. J., Batiza, R. & Staudigel, H. 1988 He, Pb, Sr and Nd isotope constraints on magma genesis and mantle heterogeneity beneath young Pacific seamounts. *Contrib. Mineral. Petrol.* **99**, 446–463.

- Graham, D., Lupton, J., Albarede, F. & Condomines, M. 1990 Extreme temporal homogeneity of helium isotopes at Piton de la Fournaise, Reunion Island. *Nature, Lond.* **347**, 545–548.
- Griffiths, R. W. & Campbell, I. H. 1990 Stirring and structure in mantle starting plumes. *Earth planet. Sci. Lett.* **99**, 66–78.
- Griffiths, R. W. & Campbell, I. H. 1991 On the dynamics of long-lived plume conduits in the convecting mantle. *Earth planet. Sci. Lett.* **103**, 214–227.
- Gurnis, M. & Davies, G. F. 1986 The effect of depth-dependent viscosity on convective mixing in the mantle and the possible survival of primitive mantle. *Geophys. Res. Lett.* **13**, 541–544.
- Hart, S. R. 1988 Heterogeneous mantle domains: signatures, genesis and mixing chronologies. *Earth planet. Sci. Lett.* **90**, 273–296.
- Hart, S. R., Hauri, E. H., Oschmann, L. A. & Whitehead, J. A. 1992 Mantle plumes and entrainment: isotopic evidence. *Science, Wash.* **256**, 517–520.
- Hiyagon, H. & Ozima, M. 1986 Partition of noble gases between olivine and basalt melt. *Geochim. cosmochim. Acta* **50**, 2045–2057.
- Hofmann, A. W. & Hart, S. R. 1978 An assessment of local and regional isotopic equilibration in the mantle. *Earth planet. Sci. Lett.* **38**, 44–62.
- Jenkins, W. J., Edmond, J. M. & Corliss, J. 1978 Excess ^3He and ^4He in Galapagos submarine hydrothermal waters. *Nature, Lond.* **272**, 156–158.
- Kaneoka, I. 1983 Noble gas constraints on the layered structure of the mantle. *Nature, Lond.* **302**, 698–700.
- Kellogg, L. H. & Wasserburg, G. J. 1990 The role of plumes in mantle helium fluxes. *Earth planet. Sci. Lett.* **99**, 276–289.
- Klein, F. W. 1982 Patterns of historical eruptions at Hawaiian volcanoes. *J. Volcan. geotherm. Res.* **12**, 1–35.
- Kurz, M. D. & Jenkins, W. J. 1981 The distribution of helium in oceanic basalt glasses. *Earth planet. Sci. Lett.* **53**, 41–54.
- Kurz, M. D., Jenkins, W. J. & Hart, S. R. 1982a Helium isotopic systematics of oceanic islands: implications for mantle heterogeneity. *Nature, Lond.* **297**, 43–47.
- Kurz, M. D., Jenkins, W. J., Schilling, J. G. & Hart, S. R. 1982b Helium isotopic variations in the mantle beneath the North Atlantic Ocean. *Earth planet. Sci. Lett.* **58**, 1–14.
- Kurz, M. D., Jenkins, W. J., Hart, S. & Clague, D. 1983 Helium isotopic variations in Loihi Seamount and the island of Hawaii. *Earth planet. Sci. Lett.* **66**, 388–406.
- Kurz, M. D., Garcia, M. O., Frey, F. A. & O'Brien, P. A. 1987 Temporal helium isotopic variations within Hawaiian volcanoes: basalts from Mauna Loa and Haleakala. *Geochim. cosmochim. Acta* **51**, 2905–2914.
- Kurz, M. D., Colodner, D., Trull, T. W., Moore, R. B. & O'Brien, K. 1990 Cosmic ray exposure dating with *in situ* produced cosmogenic ^3He : results from young Hawaiian lava flows. *Earth planet. Sci. Lett.* **97**, 177–189.
- Kurz, M. D. & Kammer, D. P. 1991 Isotopic evolution of Mauna Loa Volcano. *Earth planet. Sci. Lett.* **103**, 257–269.
- Kurz, M. D., Moore, R. B., Kammer, D. P. & Gulesserian, A. 1992 An isotopic study of dated alkali basalts from Sao Miguel, Azores: Implications for the Origin of the Azores Hot Spot. *Earth planet. Sci. Lett.* (Submitted.)
- La Tourette, T. Z. & Burnett, D. S. 1992 Experimental determination of U and Th partitioning between clinopyroxene and natural and synthetic basaltic liquid. *Earth planet. Sci. Lett.* **110**, 227–244.
- Liu, M. & Chase, C. G. 1991 Evolution of Hawaiian basalts: a hotspot melting model. *Earth planet. Sci. Lett.* **104**, 151–165.
- McKenzie, D. 1985 The extraction of magma from the crust and mantle. *Earth planet. Sci. Lett.* **74**, 81–91.
- McKenzie, D. & O'Nions, R. K. 1991 Partial melt distributions from inversion of rare earth element concentrations. *J. Petrol.* **32**, 1021–1091.
- Navon, O. & Stolper, E. 1986 Geochemical consequences of melt percolation: the upper mantle as a chromatographic column. *J. Geol.* **95**, 285–307.

- O'Nions, R. K. & Oxburgh, E. R. 1983 Heat and helium in the Earth. *Nature, Lond.* **306**, 429.
- Poreda, R., Schilling, J. G. & Craig, H. 1986 Helium and hydrogen isotopes in ocean-ridge basalts north and south of Iceland. *Earth planet. Sci. Lett.* **78**, 1–17.
- Rhodes, J. M., Wenz, K. P., Neal, C. A., Sparks, J. W. & Lockwood, J. P. 1989 Geochemical evidence for invasion of Kilauea's plumbing system by Mauna Loa magma. *Nature, Lond.* **337**, 257.
- Ribe, N. M. 1985 The generation and composition of partial melts in the Earth's mantle. *Earth planet. Sci. Lett.* **73**, 361–376.
- Richter, F. M. 1986 Simple models for trace element fractionation during melt segregation. *Earth planet. Sci. Lett.* **77**, 333–344.
- Riley, G. N. Jr & Kohlstedt, D. L. 1991 Kinetics of melt migration in upper mantle-type rocks. *Earth planet. Sci. Lett.* **105**, 500–521.
- Rubin, M., Gargulinski, L. K. & McGeehin, J. P. 1987 Hawaiian radiocarbon dates. U.S. Geological Survey Prof. Paper 1350, 213–242.
- Sarda, P. & Graham, D. 1990 Mid-ocean ridge popping rocks: implications for degassing at ridge crests. *Earth planet. Sci. Lett.* **97**, 268–289.
- Silver, P. G., Calson, R. W. & Olson, P. 1988 Deep slabs, geochemical heterogeneity, and the large-scale structure of mantle convection: investigation of an enduring paradox. *A. Rev. Earth planet. Sci.* **16**, 477–541.
- Sleep, N. H. 1990 Hotspots and mantle plumes: some phenomenology. *J. geophys. Res.* B **95**, 6715–6736.
- Staudacher, T., Sarda, P. & Allegre, C. J. 1990 Noble gas systematics of Reunion island, Indian Ocean. *Chem. Geol.* **89**, 1–17.
- Staudigel, H., Zindler, A., Hart, S. R., Leslie, T., Chen, C. Y. & Clague, D. 1984 The isotope systematics of a juvenile intraplate volcano: Pb, Nd, and Sr isotope ratios of basalts from Loihi seamount. *Earth planet. Sci. Lett.* **69**, 13–29.
- Tatsumoto, M. 1978 Isotopic composition of lead in oceanic basalt and its implication to mantle evolution. *Earth planet. Sci. Lett.* **38**, 63–87.
- Vance, D., Stone, J. O. H. & O'Nions, R. K. 1989 He, Sr, and Nd isotopes in xenoliths from Hawaii and other oceanic islands. *Earth planet. Sci. Lett.* **96**, 147–160.
- White, W. M. 1985 Sources of oceanic basalts: radiogenic isotopic evidence. *Geology* **13**, 115–118.
- Zindler, A. & Hart, S. R. 1986a Chemical geodynamics. *A. Rev. Earth planet. Sci.* **14**, 493–571.
- Zindler, A. & Hart, S. R. 1986b Helium: problematic primordial signals. *Earth planet. Sci. Lett.* **79**, 1–8.

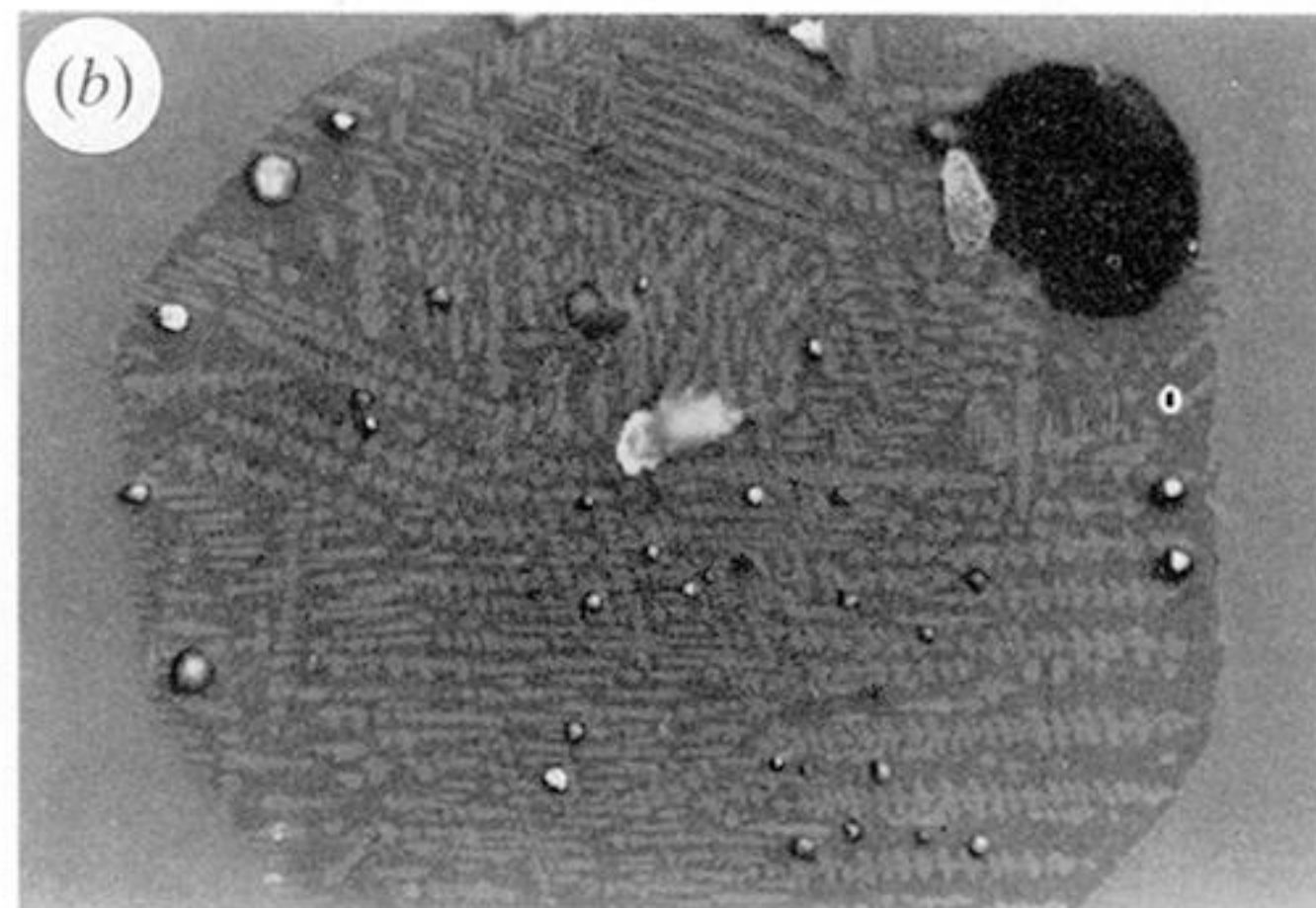
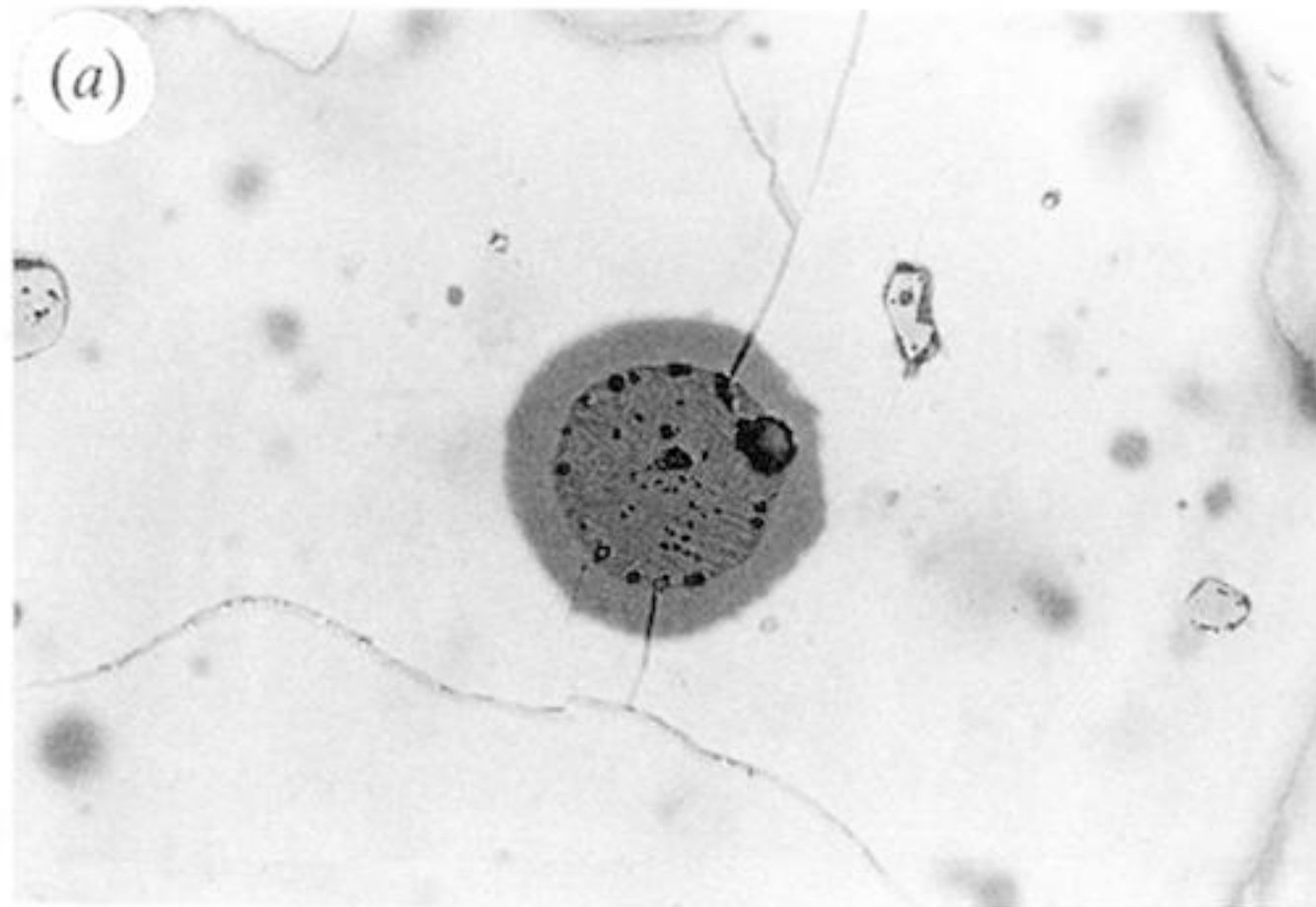


Figure 2. Photographs of an olivine grain containing a large melt inclusion exposed on a polished surface, from the 1868 lava flow of Mauna Loa, in transmitted and reflected light. (a) The spherical melt inclusion in the centre of the field is approximately $200\text{ }\mu\text{m}$ (width of field 1.2 mm , magnification $50\times$). The lighter area surrounding the inclusion is the olivine host crystal; small dark patches are smaller melt and spinel inclusions within the olivine. (b) This higher magnification photograph (field of view *ca.* $250\text{ }\mu\text{m}$, magnification $250\times$) shows that the melt inclusion has undergone post-entrapment crystallization, as evidenced by skeletal pyroxene crystals. The large round dark area at the top of the field is a bubble, as are the smaller ones dispersed throughout the inclusion. The angular dark area near the centre of the inclusion is an ion microprobe crater.