

SOLIDUS CURVES, MANTLE PLUMES, AND MAGMA GENERATION BENEATH HAWAII

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Abstract. The eruption of nepheline-normative lavas in the early and late stages of formation of the large Hawaiian tholeiite shields is well established, as is the conclusion that volatile components are involved in the genesis of these alkaline lavas. For magmas to be generated, the source materials must be transported across their solidus curves. Solidus curves for volatile-free peridotite and for peridotite-C-H-O provide the depth-temperature framework for the sites of magma generation. The assumption (controversial) that garnet remains in the source material locates the major melting region in plume material at depths of about 80 km, with isotherms in plume center exceeding 1500°C. The plume carries traces of volatile components from depths greater than 300 km. These dissolve in a trace of interstitial melt as the plume crosses the solidus for peridotite-C-H-O at depths decreasing from 350 km to about 150 km with distance from the plume axis. The volatile-charged melt, enriched in incompatible elements, is swamped by the picrites generated in the major melting region. From the outer portions of the plume, the volatile-rich melt enters the lithosphere at 80-90 km depth, where the change in rheology retards its upward percolation. This magma (remaining in equilibrium with peridotite) is carried toward the solidus for peridotite-C-H-O, changing composition toward nephelinite; evolution of vapor as magma approaches the solidus may facilitate intermittent crack propagation, releasing the nephelinitic magmas for eruption from depths of 75-85 km. Movement of the lithosphere plate over the rising plume establishes asymmetry. Eruption of nephelinitic magmas on the upstream side of the plume (early volcanism) may be suppressed or very close in time and space to eruption of alkaline lavas and tholeiites. On the downstream side (late volcanism), eruption of nephelinites is delayed by lateral transport away from the main melting region.

Introduction

The fundamental constraint for the generation of magmas is that the source material must be transported across its solidus curve by (1) local increase of temperature, through regional heating or stress, (2) physical movement of source material, e.g., by convection, (3) flux of volatile components lowering the solidus curve below the ambient temperature of the source

material, or (4) local decrease of pressure (for a source already near its solidus temperature). The fundamental variables to be determined in placing constraints on the conditions for melting of mantle are (1) the compositions of the source materials, (2) their solidus curves as a function of pressure, volatile components, and oxidation state, and (3) the physical history of the source materials, which involves tracking the flow lines of the materials within the mantle, and determining the changes in thermal structure arising from the motions.

Since Wilson [1963] first proposed that the Hawaiian Islands were formed by lavas generated in a fixed hot spot beneath a moving lithospheric plate, and Morgan [1972] developed the idea of deep mantle convection plumes in relationship to plate motions, there have been many papers dealing with the petrology and geochemistry of Hawaiian basalts, leading to the conclusion that two or more chemically distinct source materials are involved. There have been papers dealing with the geophysics and dynamics of mantle plumes, some of which attempted to correlate the geochemistry of magmas with the source materials involved in the plume, but there have been few papers dealing with the detailed consequences of a thermal plume impinging on the base of the moving lithosphere. Few of the geochemical papers or the physical papers have taken into consideration the actual positions of the solidus curves for the source materials, in relationship to flow lines and isotherms associated with the thermal plume assumed to rise beneath Hawaii. Wright and Helz [1987] reviewed recent advances in these topics.

In this contribution I adopt a somewhat arbitrary set of isotherms and flow lines for a mantle plume rising beneath the oceanic lithosphere, which brings into the melting region three source materials with distinct geochemical characteristics. Superposition of solidus curves for the materials in the pressure-temperature framework defines the sites of melting, and results from experimental petrology give the compositions of melts produced at depth. Some of the starting points adopted are still controversial, but similar frameworks can be constructed by starting with alternative assumptions. This scheme is not presented as an answer to what happens beneath Hawaii, but as a flexible physical framework that appears to accommodate much petrological and geochemical data, a framework that can be modified (1) to fit improved models of mantle convection and temperatures, (2) as experimental data on peridotite-C-H-O with variable oxygen fugacity become available, (3) as we obtain a clearer picture of the actual oxygen fugacity and its variation, and (4) as we determine the compositional variations within the mantle.

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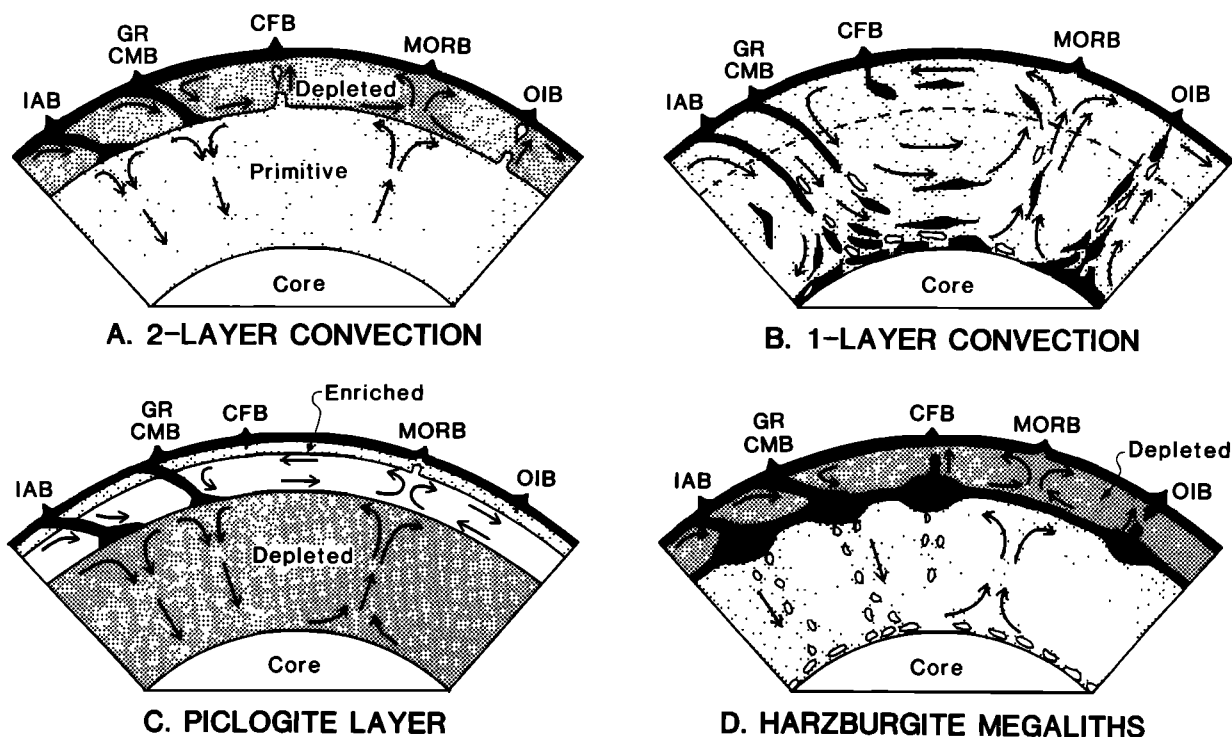


Fig. 1. Four current interpretations of convection in the mantle, showing possible sources for the main basalt types. See text for details and source references. OIB is ocean island basalts.

Mantle Convection and Magma Sources

The observations of plate tectonics have led to general acceptance of mantle convection, but the details remain controversial, as illustrated by the sketches in Figure 1 representing four models currently in vogue. This figure is presented to illustrate the variety of source materials and processes that might be involved in the formation of a mantle plume beneath Hawaii, represented by OIB (ocean island basalt).

There is clear isotopic evidence from basalts for the existence of geochemical reservoirs in the mantle that have remained physically separate from each other for a billion years or so, with up to five different sources now being recognized [Carlson, 1987; McCallum, 1987]. The questions in chemical geodynamics [Allegre, 1982] concern the extent to which these distinct reservoirs represent layers convecting separately without significant mass transfer across boundaries, and the extent to which they represent heterogeneous masses or blobs within a convecting mantle.

Figure 1a illustrates model I of Jacobsen and Wasserburg [1981], with the lithosphere overlying two convective layers. The upper layer has been progressively depleted by the extraction of magmas to form oceanic and continental crust [Wasserburg and DePaolo, 1979]. Mantle-wide convection illustrated in Figure 1b has been advocated by Hager [1984] and Davies [1984], with impressive support provided by interpretations incorporating mantle tomography, fluid dynamics, and predicted geoids [Hager and Clayton, 1987]. In Figure 1b there is significant increase in viscosity through the transition region, which

means that flow in the lower mantle would be much slower than in the upper mantle. Anderson [1982, 1985] developed the model in Figure 1c, with the upper mantle crystallized from melts extracted from the depleted lower mantle early in Earth history, and the formation of a piclogite layer (olivine + eclogite) overlain by lherzolite enriched by kimberlite-like magma. OIBs are formed from the latter layer. Ringwood [1982] described a model illustrated in Figure 1d, where subducted oceanic lithosphere forms megaliths of depleted harzburgite between upper and lower mantle layers. Eventual warming of the megaliths causes partial melting of the enclosed eclogite from oceanic crust, and the refertilized harzburgite then yields rising blobs as plumes to feed IOBs.

Tomographic studies of the mantle are beginning to provide three-dimensional pictures of mantle convection [Dziewonski, 1984; Anderson, 1987; Hager and Clayton, 1987], and eventually we may know the form of current flow. Only from isotopic studies can we obtain information about ancient convection patterns. In the meantime, cases have been made for all examples in Figure 1.

Upper Mantle, Geotherm, and Peridotite Solidus

Figure 2 illustrates an average oceanic geotherm calculated by Richter and McKenzie [1981] for a specific mantle model, with three layers. The uppermost layer, the lithosphere, loses heat by conduction, whereas the superimposed layers 2 and 3 represent separate geochemical reservoirs for basalts, corresponding

to the depleted and primitive layers, respectively, in Figure 1a. The geotherm is adiabatic in each of these layers, with a marked increase in temperature between the two layers (determined from laboratory and numerical experiments). The asthenosphere-lithosphere boundary layer is the level where the mantle rheology changes from ductile to brittle. This is arbitrarily assumed to coincide with the 1200°C isotherm in the following treatments.

Volatile-Free Peridotite

Several solidus curves have been published for natural peridotites considered to be candidates for mantle material, with varied results, as illustrated by Wyllie [1984, Figure 2], who extrapolated the experimental data available (limited to pressures corresponding to depths of 160 km) down to depths of 700 km in the upper mantle [Wyllie, 1984, Figure 7b]. Takahashi [1986] has since published the phase relationships for a lherzolite up to pressures of 200 kbar, with results reproduced in Figure 2. The distinctive features compared with previous extrapolations from lower pressures, relevant for this paper, are the cusps on the solidus, associated with subsolidus phase changes, and the very small increase in solidus temperature between depths of about 150 and 480 km.

The geotherm does not intersect the solidus in Figure 2, indicating no magma generation under these conditions. The geotherm rises to higher temperatures in upwelling convection, or in a thermal plume, and melting begins where the geotherm first intersects the solidus. The concept of initiation and control of mantle melting at a solidus cusp was first developed by

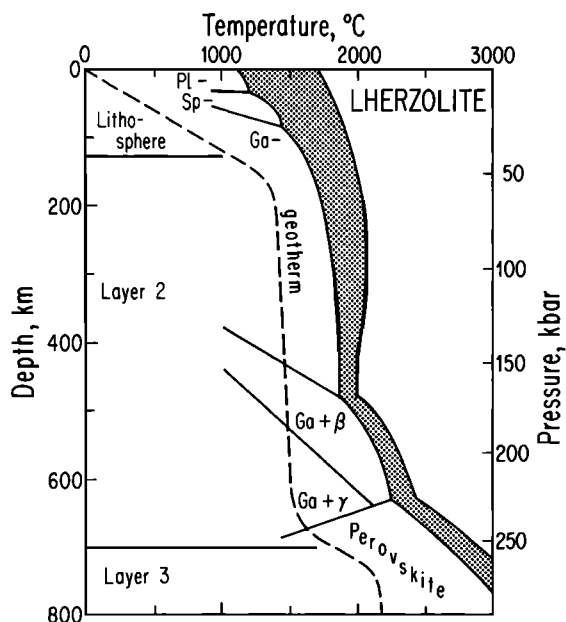


Fig. 2. Phase diagram for mantle lherzolite according to Takahashi and Kushiro [1983] and Takahashi [1986], compared with an average mantle geotherm calculated by Richter and McKenzie [1981], for a specific two-layer convecting mantle.

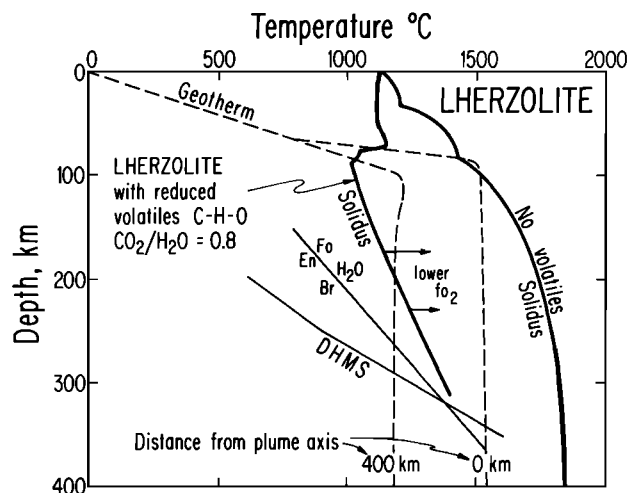


Fig. 3. Solidus curves for lherzolite with and without volatile components, and deep dehydration reactions, compared with geotherms located at center of plume axis and 400 km away, downstream. Compare Figures 2 and 5, and see text for references.

Presnall et al. [1979], considering midocean ridge basalts and the plagioclase-spinel lherzolite transition (Figure 2). For the particular values of geotherm and solidus in Figure 2, it appears that a high-temperature mantle plume might intersect the cusp on the solidus corresponding to the phase transition from spinel-lherzolite to garnet-lherzolite, at about 80-km depth. This is not a requirement of the present model.

Solidus for Peridotite with Volatiles

The solidus from Figure 2 is reproduced in Figure 3, and compared with a solidus for lherzolite in the presence of volatile components [Wyllie, 1987a]. This is based on experimental data in peridotite-CO₂-H₂O to pressures corresponding to 100 km [Eggler, 1978; Wyllie, 1978], and extrapolated to higher pressures with qualifications discussed by Wyllie [1987a]. There remains uncertainty about interpretation of conditions between 70 and 100 km [Wyllie, 1987b; Eggler, 1987], but future adjustments would simply require modification of the framework outlined for the current problem. The extrapolated solidus is assumed to remain near that for lherzolite-H₂O in Figure 3. D. H. Green (personal communication, 1987) now has experimental confirmation that with lower oxygen fugacity it will move to higher temperatures, as indicated.

The question of oxygen fugacity in the mantle and its effect on melting temperatures and magmatic products remains controversial. Wyllie [1980, 1987a] presented an interpretation of kimberlites involving the uprise of reduced gases C-H-O which dissolved in melts, became oxidized during magmatic processes, and were released at shallower levels as CO₂+H₂O. An alternative was suggested [Wyllie, 1980, p. 6905]: "A third possibility is that the components C-H-O may occur in the mantle peridotite without the

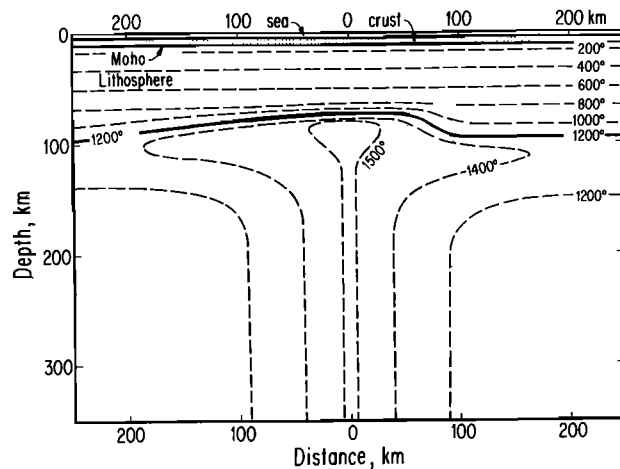


Fig. 4. Schematic isotherms for mantle plume with flow lines in Figure 5 based on plumes of Courtney and White [1986], the requirement for melting at M in Figure 3, and with asymmetry caused by motion of lithosphere plate above the plume. The change in rheology associated with the asthenosphere-lithosphere boundary layer is arbitrarily represented by the heavy line for the 1200°C isotherm.

occurrence of melting if reduced oxygen fugacity raises the solidus for the system peridotite-C-H-O, especially if the subcontinental geotherm is lower than that plotted. Partial melting would then occur only where temperatures were raised or oxygen fugacity was increased." Green et al. [1987], using the data of Taylor and Green [1987], have developed this idea in elegant detail, referring to the process as redox melting. The partial melting of peridotite in the presence of rising CH_4 , H_2 is delayed until a more oxidized level in the lithosphere is reached where the solidus becomes lowered to a temperature below the geotherm and melting begins. A more oxidized mantle was preferred by Egger and Baker [1982] and by Woermann and Rosenhauer [1985], who presented a detailed evaluation of the phase relationships in peridotite- CO_2 - H_2O with reduced oxygen fugacity. According to D. H. Egger (personal communication, 1987), "My sources now indicate that it is by no means certain that asthenospheric oxygen fugacity is low enough (well below magnetite-wustite) that volatile components in C-H-O would exist as H_2O and CH_4 rather than as H_2O and CO_2 ."

Deep Storage of Volatile Components in Solid Mantle

If there are sufficient quantities of the components C-H-O in the mantle for them to be expressed as components, rather than as adsorbed films, or dissolved in the surface of minerals, then reactions forming compounds between the volatile and solid components must be considered (for reviews see Wyllie [1978, 1987a] and Egger [1987]). Amphibole and phlogopite stability ranges are limited to the upper mantle, with amphibole normally restricted to less than 100 km, and phlogopite persisting somewhat

deeper. H_2O and CO_2 that can exist as vapors in the upper mantle become locked into minerals at greater depths. Even if the mantle is sufficiently oxidized, CO_2 cannot exist in the mantle deeper than about 75 km, because it reacts with lherzolite to form calcic dolomite (magnesite somewhat deeper). Similarly, H_2O reacts with lherzolite to form brucite or dense hydrated magnesian silicates (DHMS) at depths below the two reaction curves in Figure 3, between 250 and 350 km, depending upon the geotherm. Reduced vapors CH_4 and H_2 will persist in dense, gaseous form to greater depths.

Mantle Plume, Isotherms, and Flow Lines

In the presence of traces of volatile components the amount of melt produced from peridotite below the volatile-free solidus is very small. If it can be demonstrated that sufficient melt would be generated at somewhat lower temperatures, this would require only minor adjustments in Figures 6, 7, and 8, with no change in the petrological conclusions.

I assume that the hot spot generating the Hawaiian basalts is produced by a mantle plume, although this is not established. Other processes involving propagating fractures and shear melting have been proposed. Green [1971], for example, illustrated a process involving migrating tension fractures whereby melts were tapped not from a plume but from a chemically zoned asthenosphere. The concept of plumes may include material rising from the core-mantle boundary (Figure 1b), from a primitive layer below 670 km (Figure 1a), from a megalith at the 670-km level (Figure 1d), from a shallow enriched layer below the lithosphere (Figure 1c), or from large-volume isolated blobs within the mantle (Figure 1b).

Courtney and White [1986] constructed a variety of theoretical models of hot spot mechanisms, constrained by heat flow, geoid, and bathymetric values across the Cape Verde Rise, illustrating the thermal structure in a plume rising below a static lithosphere 70-80 km thick (increasing in thickness with distance from the plume center). These thermal structures were adopted as the basis for Figure 4, with isotherm values established by the conditions for melting to occur in Figures 2 and 3 and with the symmetry of the plume modified by a moving lithosphere plate. The asthenosphere-lithosphere boundary corresponds to the 1200°C isotherm.

Frey and Roden [1987, p. 434] stated, "There is no consensus on a general petrogenetic model for Hawaiian volcanism." This leaves considerable latitude for setting up a thermal framework. There is debate about the conditions for melting at the source of the magmas parental to the Hawaiian lavas (for reviews, see Feigensen [1986], Frey and Roden [1987], Hofmann et al. [1987], Lanphere and Frey [1987], and Wright and Helz [1987]). Many geochemists maintain that garnet must be present in the source rock, but others deny this requirement. Green and Ringwood [1967] demonstrated that garnet does not appear on the liquidus of primitive olivine tholeiite, and Green et al. [1987] confirmed that magnesian picritic liquids are required to approach equilibrium with garnet peridotite. Some

geochemists have concluded independently that high-Mg picritic magmas are parental to Hawaiian tholeiites, while others argue that the high transition element abundances in Hawaiian tholeiites preclude a picritic parent.

I have adopted in Figure 4 the constraint that the Hawaiian lavas were derived from a source rock containing garnet, which requires that melting occurred at the cusp M in Figure 3 (see Figure 2 for garnet) or deeper. Therefore the geotherm corresponding to that in Figure 2, but located in the center of the plume in Figure 4, must reach M, at about 1500°C. If in fact the parental magmas were formed from spinel-peridotite at shallower depths, this requires some adjustment of the isotherms in Figure 4, with a somewhat thinner lithosphere above the plume, but I believe such adjustments can be followed through the rest of this treatment without introducing significant changes in the petrological conclusions.

The lithosphere is heated above the plume, and associated with the rise in isotherms is thinning of the lithosphere, as shown in Figure 4. This effect would be symmetrical about the plume axis for a static lithosphere, as in the diagrams of Courtney and White [1986], but if the lithosphere plate is moving, as in Hawaii, then the top of the plume becomes asymmetric, as do the isotherms. Presnall and Helsley [1982] presented a detailed analysis of diapirs and thermal plumes, in which they assumed that the velocity of the plate was greater than that of the rising material, so that the plume was bent over and incorporated into the asthenosphere on one side, or sheared off by the drifting lithosphere. This alternative merits more detailed investigation with geotherms, flow lines, and solidus curves.

For a plume diverging in all directions when it reaches the lithosphere, Figure 4 represents the distorted thermal structure (compared with the symmetrical versions presented by Courtney and White [1986]) produced by flow lines represented schematically in Figure 5. Different mantle materials, potential sources for magmas, are distinguished by different shading. Mantle plumes at the shallow levels shown in Figure 4 may have their origins in several different arrangements at depth as outlined above and illustrated in Figure 1. There is evidence [Carlson, 1987; Frey and Roden, 1987] that source materials for Hawaiian lavas include at least three components (primitive, depleted, and enriched mantle). Figure 5 shows a plume composed dominantly of primitive mantle corresponding with layer 3 in Figure 2, with an outer sheath of depleted layer 2 flowing with the plume, as indicated by the shading. The residual lithosphere is a third source material present in the melting region. Given the variety in Figure 1, it would be easy to devise other plume models capable of bringing a variety of source materials into the melting region, with geometry corresponding to layers and sheaths, or to heterogeneous blobs.

The rigid lithosphere plate moving from right to left is about 100 km thick, and as it passes over the plume it becomes heated and the 1200°C isotherm rises, thus thinning the lithosphere. Furthermore, the former lithosphere with temperature now elevated above 1200°C is capable of flow, and this material is swept to the right

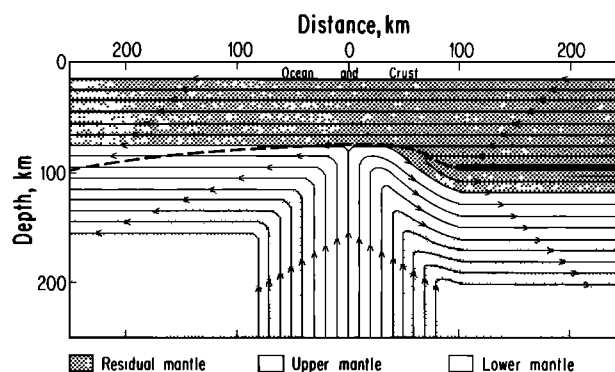


Fig. 5. Schematic flow lines for mantle plume rising beneath moving oceanic lithosphere plate. The heavy dashed line represents the asthenosphere-lithosphere boundary layer; compare Figure 4. The plume comprises a central portion rising from the lower mantle, layer 3 in Figure 2, and an outer sheath from the upper mantle, layer 2 in Figure 2.

by the diverging plume returning in the direction from whence it came, but at a deeper level. This process returns lithosphere material to the asthenosphere. The rising plume material that migrates to the left replaces the lithosphere material transported to the right, and as it cools, the 1200°C isotherm moves back to lower levels and former plume material becomes incorporated into the lithosphere. Nickel and Green [1985] referred to lithosphere thinning and thickening in similar fashion. The asymmetry of the plume and the 1200°C asthenosphere-lithosphere boundary follows from the low thermal conductivity of the rocks compared with the rate of movement of plume and plate.

The isotherms in Figure 4 provide geotherms for any given distance from the plume center. Figure 3 compares two geotherms with the solidus curves for lherzolite: the geotherm for the center of the plume, and the geotherm for a position 400 km downstream from the plume axis (where the asymmetry is greatly reduced). The steep thermal gradient above the crest of the plume implies that the lithosphere that is not swept away and incorporated into the asthenosphere is not heated very much during its passage across the plume, again, a consequence of low thermal conductivity and moving lithosphere. This picture may be modified, however, if magma rising from M causes local heating of the lower lithosphere. Comparison of the geotherm and solidus in Figure 3 shows that significant heating is required to melt the lithosphere.

Petrological Structure of Plume

Each point on Figure 4 is characterized by a specific pressure and temperature. The phase boundaries in Figure 3 are mapped in terms of pressure and temperature, and therefore they can be plotted on Figure 4, with results shown in Figure 6. For simplicity, the two dehydration curves in Figure 3 were merged into a single dehydration boundary. These phase boundaries are static, fixed in position as long as the isotherms do not change. The arrows in Figure 6

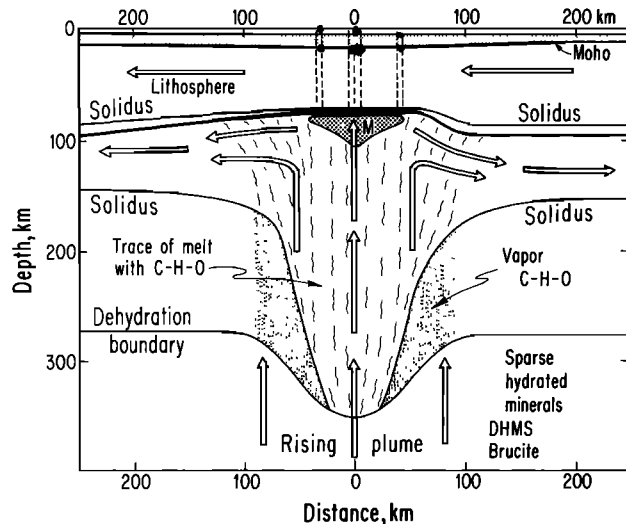


Fig. 6. The isotherms in Figure 4 and the phase boundaries in Figure 3 define the positions of solidus curves and dehydration boundaries as shown in this mantle cross section. Material following flow lines in Figure 5 crosses the phase boundaries, causing the distribution of vapor and melt in rising plume as shown. Picritic magma enters the lithosphere from the region of major plume melting at M, with magma chambers forming at various levels. Figure 7 shows more details in the region of the asthenosphere-lithosphere boundary layer (heavy line).

show movement of material through the steady state plume, following the flow lines in Figure 5. The geometry of the static petrological structure is changed by the process of moving material through this framework, with release or absorption of heat at phase transformations, and chemical differentiation caused by physical transfer of vapor or melt from its solid parent.

The changes in Figures 6 and 7 between a static framework and a dynamic model are probably smaller than those involved in the uncertainties about the variation of oxygen fugacity in the mantle and its effect on solidus temperatures. Figures 6 and 7 are therefore suitable for illustrating the types of reactions and processes to be expected in a plume reaching the lithosphere.

If there are volatile components in the deep mantle, H_2O is released from the hydrates as the plume material crosses the dehydration boundary [Woermann and Rosenhauer, 1985], joining any CH_4 that was already present and reacting with diamond to generate CH_4 [Deines, 1980]. The mantle plume with volatiles rises until it reaches the solidus boundary, where the volatiles become dissolved in a trace of melt. In the central part of the plume, for the thermal structure in Figure 4, the hydrated peridotite melts directly without an interval with a separate vapor phase. The interstitial melt generated by the volatiles is entrained in the rising plume, moving laterally into the asthenosphere to an extent controlled by the

relative rates of mantle flow, and upward percolation of melt. Significant melting of the plume occurs in the shaded area M, which is situated above the volatile-free solidus (see M in Figure 3). The line around the area M exists only for volatile-free peridotite. In the presence of volatile components there is continuity between the trace of melt present in the rock volume around M, and the greatly increased melt fraction generated within the volume M. The melt in region M incorporates the volatile-charged melt that was formed about 200 km deeper in the mantle. On either side of the melting region M, the entrained volatile-charged melt does not reach the temperature of the solidus of volatile-free peridotite before reaching the lithosphere near the $1200^\circ C$ isotherm. At this level, the change in rheology would significantly retard the upward movement of the small percentage of interstitial melt.

Fundamental questions to be addressed include the relative rates of movement of solid, vapor, and melt within the plume and in the lithosphere, and the chemical consequences of different rates of flow. Advances have been made recently in these problems [e.g., Maaløe and Scheie, 1982; McKenzie, 1984; Scott and Stevenson, 1986; Marsh, 1987; Navon and Stolper, 1987; Ribe and Smooke, 1987]. Maaløe and Scheie [1982] considered the process of partial melting of lherzolite in a rising mantle plume. They defined a partial melting boundary, the level where melting begins, and a permeability boundary at a higher level where the partly molten source became permeable. They demonstrated that once permeable, the mantle would undergo compaction and the melt would be squeezed out, accumulating

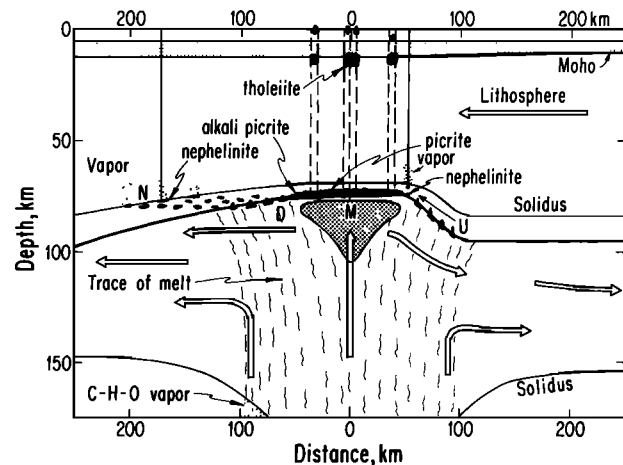


Fig. 7. Petrological structure of lithosphere and upper plume corresponding to Figure 6, with expanded vertical scale. Picritic magmas enter lithosphere from melting region M. Volatile-charged magmas enter the lithosphere on either side of M, with subsequent percolation retarded. As they approach the solidus, vapor is evolved (e.g., at N), enhancing the prospect of explosive eruption directly through the lithosphere. Paths followed by these magmas to the solidus differ according to their position on the upstream (U) or downstream (D) side of the plume, as depicted also on Figure 8.

in the form of horizontal layers. The magma may be tapped from a dispersed source consisting of a multitude of layers and interconnecting veins. With divergent flow as in the topmost part of the plume, the layers come apart and a large magma chamber may be formed [Maaløe, 1981]. Ribe and Smooke [1987] also concluded that layers of segregated melt are likely to form magma chambers above plumes at the base of the lithosphere, and that about 60–80% of the melt generated within the plume is extracted from an area comparable to that of the plume itself. Navon and Stolper [1987] modeled aspects of the chemical interaction between rising melts and the mantle in terms of ion-exchange processes with particular attention to trace elements, and including specific treatment of melt in a rising plume.

Magma Compositions

The compositions of the magmas reaching the lithosphere are labelled in Figure 7, according to their locations with respect to the solidus boundaries in Figure 8, and the experimentally determined compositions of melts under the specified conditions. Summaries of the melt compositions in these locations have been published by Jacques and Green [1980], Takahashi and Kushiro [1983], and Wyllie [1987a]. The magma generated in the major melting region M is picritic; at the edges of the region, where the percentage of melting is smaller, the magma is alkali picrite. The volatile-charged melt rising in the plume is a low-SiO₂, high-alkali magma, with composition probably similar to kimberlite at least at depths near 200 km. Migration of this relatively high temperature magma from D and U through the lithosphere to the lower-temperature solidus N (Figures 7 and 8) is sufficiently slow and in sufficiently small volumes that the melt continuously reequilibrates with the host peridotite, changing composition to nephelinite, basanite, or related magma.

Magmas in the Lithosphere

Figure 7 shows in more detail what may happen in the lower lithosphere. The picritic magmas generated by major melting in the region M enter the lithosphere, where they are probably trapped in chambers until suitable tectonic conditions permit their uprise through cracks to shallower magma chambers. Magma chambers are probably formed near the Moho, and certainly within the volcanic edifices. Note from Figures 3, 7, and 8 that the regional thermal structure imposed by the plume is not conducive to partial melting of the lithosphere. The emplacement of high-temperature picritic melt from M into the lower lithosphere, however, may introduce enough heat to cause melting in the deep lithosphere. Some geochemical interaction between plume-derived melt and lithosphere is to be expected [Chen and Frey, 1985].

The tholeiites and alkali basalts erupted at the surface must differentiate in magma chambers from the parental picritic magmas. If the volume requirements are not satisfied by the processing of the mantle plume through the region M, the scheme can be modified to yield more lava by

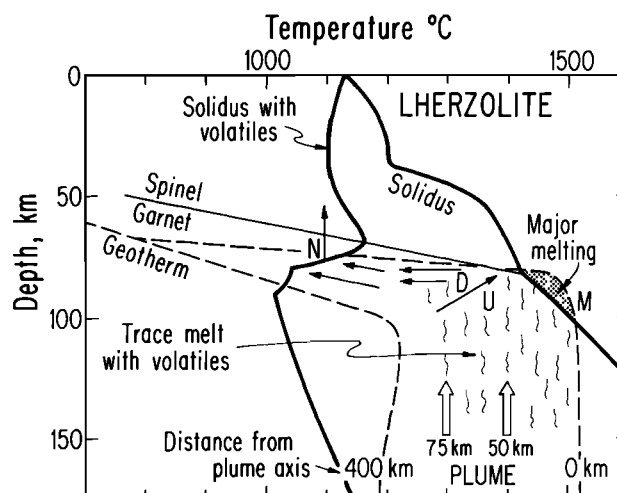


Fig. 8. Phase boundaries and plume geotherms from Figure 3 as depth-temperature framework for tracing paths of melt in Figure 7. The temperature of the volatile-charged interstitial melt rising in the plume varies as function of distance from the plume axis. Nephelinite magmas can be erupted if the magma reaches the solidus in the region of N. After the magma enters the lithosphere, magma on the upstream side of the axis (U) may rise up with decreasing temperature to the solidus at N, or it may be carried into the region of the high-temperature melt above M (Figure 7). On the downstream side of the plume axis (D), the magma is transported laterally with decreasing temperature to eruption sites well away from the axis (Figure 7).

increasing the temperature in the plume center (Figure 4). This would cause intersection of the geotherm with the solidus at a deeper level than M (Figure 3), which would thus expand the area M in Figure 7, with a much larger volume of plume being melted. More melt would also be generated by an increase in oceanic crust component in the plume [Feigenson, 1986].

The fate of the volatile-charged low-SiO₂, high-alkali melt differs according to its position within the plume. The melt in the central region is incorporated into the much larger volume of melt generated in the region M. The melt on either side of the region M reaches the lithosphere, and then follows different routes according to its location on the upstream (U) or downstream (D) side of M.

Depending upon the relative rates of mantle flow and melt percolation in the region U (Figures 7 and 8), magma on the upstream side could (1) be swept back into the asthenosphere following flow lines in Figure 5, (2) percolate slowly through the lithosphere with decreasing temperature to the solidus for peridotite-C-H-O, near position N in Figure 8, or (3) percolate upward and downstream with lithosphere flow (Figure 7) along the asthenosphere-lithosphere boundary layer and become incorporated into the hotter alkali picrite magma above the upstream margin of the melting region M.

Melt entering the lithosphere on the downstream side of the plume, D, would be carried away from the plume along lithosphere flow lines

(Figure 5). It would remain above the peridotite-volatile solidus as temperature decreased along DN (Figures 7 and 8) and reach the solidus at some distance downstream from the plume axis, the distance depending on the relative values of rates of lithosphere movement and of migration of small bodies of magma through the lithosphere. Some of the alkali picrite from the downstream margin of the region M may be similarly transported, through a shorter horizontal distance, providing an additional opportunity for its differentiation.

The formation of cracks is required for escape of the small volumes of nepheline-normative magma. Crack propagation may be accomplished by buoyancy-driven magma fracture and by vapors [Artyushkov and Sobolev, 1984; Spera, 1984], and many small intrusions may suffer thermal death deep in the lithosphere [Spera, 1984]. When magma approaches the solidus N, vapor is released, and this may enhance crack propagation and permit eruption of nephelinites directly from the asthenosphere-lithosphere boundary to the surface. Note the two locations indicated in Figure 7, one close to or overlapping with the major melting region on the upstream side of the plume, and the other separated from it by a significant distance (and time) on the downstream side. Vapor released at depth N could have high $\text{CO}_2/\text{H}_2\text{O}$, in contrast to vapors deeper than the change in solidus slope near 80 km (Fig. 8), which must have high $\text{H}_2\text{O}/\text{CO}_2$ (if sufficiently oxidized).

Magmatic History of Lithosphere Crossing a Plume

Consider the sequence of events for a lithosphere plate drifting over a hot spot generated by a mantle plume. A specific surface area receives magmas in succession following the processes depicted in Figure 7 from right to left.

The first event could be submarine eruption of nephelinites or basanites, followed rapidly by, or almost contemporaneously with, the formation of alkali picrites at depth and their fractionation to alkali basalts at shallower levels. This is followed with no time break by major eruptions of tholeiites, formed by fractionation in the uppermost mantle and crust of the voluminous tholeiitic picrites generated from the plume material passing through the region M, together with any melt generated in the overlying lithosphere. The width of M, and the time interval for the eruption of tholeiites, is extensive compared with the initial alkalic magma phase, and its duration depends on the width of the mantle plume. The volume of tholeiite produced also increases with width of the mantle plume, and with its axial temperature. The third phase is the eruption of alkali basalts differentiated from the relatively small volumes of deep-seated alkali picrites generated at the downstream edge of the melting region, M; these eruptions would be subaerial, on the volcanic pile. This stage could be somewhat extended in time, and intermittent, if some of the alkali picrite becomes trapped in the lithosphere and is transported in the direction downstream before escaping to shallower levels. The final stage

occurs much later, when the small pockets of nephelinite magma are erupted. These eruptions could be either submarine or subaerial, depending on the distance DN in Figure 7 compared with the spacing between volcanic islands. Crack propagation may be favored in lithosphere that has previously provided conduits to the surface. Thereafter, magmatic activity ceases until the lithosphere crosses some other thermal disturbance in the mantle capable of raising the geotherm to a level where it crosses the solidus (Figure 2).

If the Hawaiian magmas are generated by a mantle plume, attention should be directed toward the processes occurring on either side of the plume track. Symmetry requires that on either side of the main tholeiite eruptions, transverse to the profiles illustrated in Figures 6 and 7, alkali picrite and volatile-charged melt should be entering the lithosphere. It is not surprising that these have no surface manifestation on volcanic islands. The alkali basalts and the nephelinitic lavas on the downstream side of Figure 7 are exposed because they were erupted through the older islands of voluminous tholeiite. Dredging on either side of the Hawaiian chain might yield some alkali basalts, but the volume of nephelinitic lavas produced in similar areas by processes corresponding to the paths DN in Figures 7 and 8 are likely to be very small.

This sequence of events appears to satisfy the broad geological and petrological history of the Hawaiian Islands [Decker et al., 1987].

Magma Sources, Trace Elements, Isotopes, and Metasomatism

There is an enormous literature on trace element and isotope constraints related to source materials and mantle reservoirs for Hawaiian and other oceanic island lavas [Decker et al., 1987; Frey and Roden, 1987; Carlson, 1987]. Presnall and Helsley [1982], Sen [1983], Chen and Frey [1985], Anderson [1985], and Feigenson [1986] are among those who have related the geochemical evidence specifically to a mantle plume.

Figure 5 represents one possible model among the several available (Figure 1). Figure 5 distinguishes between three mantle source rocks commonly considered to represent distinct geochemical reservoirs, which are brought into the melting region. The lower lithosphere is considered to be composed of residual peridotite, depleted by removal of MORB; however, metasomatic enrichment can be accomplished by the local intrusion of volatile-rich melts or vapors evolved at depth from these melts, as depicted near U and N in Figure 7. The central plume material, assumed here to be derived from the deep mantle layer 3 in Figure 2, represents a primitive, undepleted mantle reservoir source, in contrast with the marginal plume material and layer 2 in Figure 2, which is assumed to represent depleted mantle.

The framework outlined in Figures 6 and 7 suggests that different deep source materials are involved in the formation of the tholeiites and the nephelinites, with a prospect that more than one deep source may provide material for the alkalic magmas near the margins of the melting

zone. The lithosphere reservoir may contaminate nephelinites and alkali basalts on the upstream side of the plume, U, but this is less likely on the downstream side, D (see Figure 5). The tholeiites from the deep mantle reservoir may be contaminated with the lithosphere source to the extent that the hot picrite magmas at the lithosphere-asthenosphere boundary cause local melting of the lithosphere [Chen and Frey, 1985].

Appeal to mantle metasomatism to account for enrichment or depletion of mantle sources in selected trace elements is common [Frey and Roden, 1987; Wright and Helz, 1987]. Metasomatic changes caused by intrusion of melts are quantitatively different from those caused by passage of a dense vapor phase. A melt may react with its wall rock, effecting metasomatic exchanges, but eventually the melt solidifies and a significant mass of new material is added to the host rock. The passage of vapors or solutions, however, causes reactions with wall rocks that may entail either leaching or precipitation, and it takes much more vapor or solution to cause significant metasomatic changes. Little is known about the solubilities of mantle components in vapor at high pressures. Schneider and Eggler [1986] reviewed existing data, presented new measurements, and calculated that high fluid/rock ratios would be required to effect significant changes in the chemistry of rocks, concluding that vapors were not effective metasomatic agents and therefore that most mantle metasomatism was accomplished by magmas. The role of vapors as mantle metasomatic agents may be even more limited according to recent research by Watson and Brennan [1987] on the wetting characteristics of CO_2 - H_2O fluids. Their results suggest that such fluids in the upper mantle may exist only as isolated pores primarily at grain corners, incapable of migration except by hydrofracture.

In order to unravel the mystery of mantle metasomatism, many more data are required on (1) the compositions of mantle vapors in terms of CO_2 , H_2O , CH_4 , H_2 , (2) the solubilities of mantle components in these vapors as a function of pressure and temperature, and (3) the physical properties of the rock-vapor systems, as well as those of rock-melt systems. Wilshire [1984] has described the metasomatism of individual mantle samples by the intrusion of magmas, and by the infiltration of solutions emanating from the magmas during solidification [Wilshire, 1984]. Therefore let us assume that both vapors and melts can accomplish metasomatism (either by infiltration or by hydrofracture or both), and explore the consequences in Figures 6 and 7. The migration and distribution of incompatible trace elements are controlled largely by the behavior of the volatile components, and their solution in and release from melt, as illustrated in Figures 6 and 7.

The high concentrations of incompatible elements in the traces of melt in the rising plume will be swamped by the major melting in the center of region M, becoming more abundant toward the margins of the region, where alkalic picrites are generated by a smaller degree of partial melting. The melts on both upstream and downstream sides of the plume will carry the high concentrations of incompatible elements, along

with the geochemical signatures of depleted sources. The release of vapors into the lower lithosphere depicted in Figure 7 may transfer concentrations of incompatible elements into metasomatized lithosphere, contributing to local geochemical anomalies where the deep lithosphere is subsequently involved in melting. According to the experimental results of Schneider and Eggler [1986], the extent of metasomatism by vapors may vary sensitively with depth in this region. Schneider and Eggler [1986] demonstrated that addition of even small quantities of CO_2 to a hydrous solution causes a significant decrease in the solubilities of peridotite components. They suggested that rising aqueous solutions would leach the mantle with minor precipitation until they reached a level of about 70 km, where a "region of precipitation" is associated with the formation of amphibole and consequent change in vapor composition toward CO_2 .

There appears to be sufficient flexibility in the physical framework presented in Figures 6 and 7 for adjustments to be made to satisfy many of the geochemical constraints, but this detailed exercise remains to be done. The framework appears to be robust enough to accommodate modifications, and it is precisely through such modifications arising from the interplay of geochemistry, fluid dynamics, and experimental petrology that we may advance our understanding of what goes on beneath Hawaii.

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