

## Primary magmas and mantle temperatures

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**Abstract:** The composition of olivine phenocrysts in Hawaiian tholeiitic picrites and in Mid-Ocean Ridge picrites vary up to  $Mg_{91.3}^{\#}$  and  $Mg_{92.1}^{\#}$  respectively. The compositions and liquidus temperatures of the magmas crystallizing the most magnesian phenocrysts can be estimated and we find that anhydrous liquidus temperatures (at 1 bar pressure) of Hawaiian tholeiitic picrites average 1365°C, for E-MOR picrites average 1355°C, and for N-MOR picrites average 1335°C. Water contents of the magmas decrease in the order Hawaiian picrites, E-MOR picrites to N-MOR picrites, and consideration of liquidus depression by these water contents leads to the conclusion that magma temperatures for all types were approximately 1325°C at ~ 1 bar. The data from parental or primary magmas suggests that the temperature contrast between 'Hot-Spot' and MOR magmas is  $\leq 20^{\circ}\text{C}$ . Application of information from partial melting studies of lherzolites and liquidus studies of the Hot-Spot and MOR picrites leads to the conclusion that both 'Hot-Spot' and MOR primary basalts are derived from mantle with potential temperature  $T_p \sim 1430^{\circ}\text{C}$ . Insofar as primitive magmas may be used to infer the potential temperature of their sources, there is no evidence for a temperature contrast of  $\Delta T_p = 100\text{-}250^{\circ}\text{C}$  between 'Hot-Spot' or 'Deep Mantle Plume' sources and ambient (MOR source) asthenospheric mantle.

Although magma temperatures are similar, the residual mantle compositions for Hawaiian picrites are refractory harzburgites, more refractory (including Cr/Cr+Al ratio) than the lherzolite to harzburgite residue from MOR picrite extraction. It is argued that the buoyancy plume and geophysically anomalous mantle beneath the Hawaiian Arch is due to compositional and not temperature contrasts in the upper mantle. The four-component mixing identified in the Hawaiian source is attributed to interaction between old subducted lithospheric slabs, buoyant or suspended in the upper mantle, and surrounding ambient mantle at  $T_p = 1430^{\circ}\text{C}$ .

**Key-words:** mantle melting, mantle potential temperature, plumes, primary magmas.

### Introduction

The variation of temperature with depth in the Earth's crust and mantle cannot be directly measured and must be inferred. One of the more direct indications of interior temperatures is provided by the eruption temperatures of volcanoes and the matching of those temperatures with knowledge of the depth of origin and magma/source relation-

ship. In the 'Plate Tectonics' model of the modern Earth, it is commonly inferred that Mid-Ocean Ridge (MOR) volcanism is an expression of convective upwelling from relatively shallow mantle depths. To a first approximation, MOR volcanism is characterized by globally similar primary or parental magma compositions, with the most magnesian quenched glass, observed in all young ocean basins, having eruption temperatures around

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1200°C–1250°C. This global, first-order similarity led to the concept of a generalized mantle potential temperature ( $T_p$ ) with  $T_p \approx 1280^\circ\text{C}$  (McKenzie & Bickle, 1988). The crust and lithosphere are interpreted as a thin boundary layer over the asthenosphere, which is inferred to have a potential temperature of 1280°C and adiabatic temperature distribution with depth. The asthenosphere composition is inferred to be a reasonably well-mixed and slightly residual lherzolite evolved from the silicate component of the primitive Earth by separation of the continental and oceanic crust and lithosphere over  $\sim 4.5$  billion years.

Although MOR basalts are interpreted as sampling a modern, well-mixed mantle source (asthenospheric mantle), a second, relatively distinctive volcanism, ‘island chain’ or ‘hot spot’ volcanism is widely interpreted as sampling a different source or sources. The ‘hot spot’ sources are relatively fixed with respect to plate-movements and are fixed or show very slow movement relative to one another. For the best-studied examples of ‘hot-spot volcanism’ such as the Hawaiian-Emperor Island chain, Iceland, Réunion, the Azores, *etc.*, the individual volcanoes sit on top of a broad topographic swell of  $\sim 1$  km elevation above the ocean floor. The presence of the broad topographic high (*e.g.* the Hawaiian Swell or Arch) in isostatic equilibrium leads to the inference of a buoyancy plume or lower density column as the cause of the hot-spots and the source of hot-spot volcanism.

The buoyancy plume is normally attributed to a temperature difference with  $\Delta T \sim 200\text{--}250^\circ\text{C}$  (rarely to  $500^\circ\text{C}$  or down to  $100^\circ\text{C}$ ) between the normal mantle and upwelling plume (McKenzie & Bickle, 1988). In much of the literature on mantle plumes, particularly on plumes from a postulated thermal boundary layer at the core-mantle boundary, there is explicit reference to temperature difference between upwelling plume and ambient mantle of at least  $200^\circ\text{C}$  (Davies, 1999). A number of authors have suggested that there is evidence from hot-spots for the presence of high-temperature picritic, *i.e.* olivine-rich magmas, in contradistinction to the olivine tholeiites considered to be characteristic of mid-ocean ridges [using the McKenzie & Bickle (1988) analysis].

The theme of this paper is to critically examine the evidence for  $\Delta T \sim 100\text{--}250^\circ\text{C}$  between plume sources, as exemplified by Hawaii in particular, and MORB sources for both N-MORB and E-MORB, by examining the evidence from primitive or primary magmas themselves.

## Some first principles

The Earth’s mantle is dominated by peridotite and peridotite is the probable source for most primary, mantle-derived magmas. The minerals of the mantle source region are all Fe/Mg solid solutions and melting with increasing temperature is thus a continuous process, with predictable Fe/Mg partitioning between crystalline phases and liquid. A similar conclusion applies to other substitutions and coupled substitutions in olivine, pyroxenes, garnet, spinel or plagioclase. In the simple case of Fe/Mg partitioning between olivine (as the major phase) and liquid, the effect of increasing pressure moves the melting loop to higher temperature but has very little effect on the partitioning. For a particular liquid, the liquidus olivine is known and conversely, for a particular mantle containing defined olivine composition, the Fe/Mg ratio (but not the FeO and MgO content) of the liquid is known (Roedder & Emslie, 1970).

Because the mantle source is multiphase, melting also has the character of eutectic melting, *i.e.* melting of the multiphase assemblage begins at temperatures below the solidus temperature of any one phase alone. The temperature for beginning of melting, the ‘pseudo-eutectic’, increases with increasing pressure but the olivine solidus temperature increases more slowly with increasing pressure than the pyroxene solidus. As a consequence, the position of the ‘eutectic-like’ melt moves towards more olivine-rich compositions with increasing pressure (Green & Ringwood, 1967; Falloon & Green, 1988).

The subduction process at convergent margins carries crust, particularly oceanic crust, and cool lithosphere back into the upper mantle and subducted crustal and lithospheric components undergo a sequence of metamorphic reactions, including dehydration reactions, towards higher pressure and higher temperature assemblages. The average crust of ocean basins is basaltic in composition and reacts to quartz eclogite or coesite eclogite in the pressure range of 1 to 5 GPa ( $\sim 30$  to 170 km). Within widely accepted models of plate tectonics, including the concept of deep mantle plumes, there is the expectation or hypothesis that melting of subducted oceanic crust will occur under conditions of eclogite stability. Thus eclogite melting within the subducted slab is one process which is invoked to explain transfer of crustal, including ocean-floor sediments, geochemical signatures from subducted slab to the overlying mantle wedge, *i.e.* to the source region for volcanics characteristic of island arcs and back-arc basins.

Melting of basaltic composition components is also invoked within the 'deep mantle plume' hypothesis, including concepts of accumulation and reheating of subducted crust/lithosphere slabs within a core-mantle boundary layer. Ascent of inhomogeneous plumes from such a boundary layer is inferred to lead to melting during ascent, including, at shallower depths, preferential melting of eclogite components (Davies, 1999). Finally, detachment of subducted slabs of crust and lithosphere may occur following continent/continent collision or migration of trenches/subduction zones. Such suspended slabs of inhomogeneous crust, lithosphere and enclosing ambient mantle may, over time, reach 'normal' mantle temperatures — and in doing so, undergo melting in lithologies with lower solidus temperatures than normal mantle. In a later section we return to discussion of melting in an inhomogeneous mantle source comprised of asthenospheric mantle, refractory lithosphere and subducted oceanic crust.

The arguments and deductions of this paper are based on major element and refractory minor elements, and on phase equilibria constraints. A corollary is that relative abundances of incompatible elements, particularly incompatible trace elements, *may* provide information on the melting process and conditions but, like radiogenic isotopes, are better indicators of the precursor processes and pre-melting history affecting the source region.

## Temperatures of primary magmas

### A. Hawaiian tholeiitic magmas

The main cone-building phase of Hawaiian volcanoes is dominated by olivine tholeiite magmas, *i.e.* hypersthene-normative magmas, which vary from tholeiitic picrites with > 30 % modal olivine (~ 25-20 % normative olivine) to olivine-phyric quartz tholeiites (< 5 % modal olivine, ~ 5 % normative quartz). A recent paper by Norman & Garcia (1999) presented data on picritic magmas from six volcanoes or eruptive centres. Norman & Garcia showed that for each volcano the olivine compositions plotted in a distinctive field in a CaO vs.  $Mg^{\#} = 100Mg/(Mg+Fe^{2+})$  diagram, *i.e.* the calcium content of olivine correlated with the magma composition such that Loihi represented the high Ca-end (with 0.25 to 0.40 wt% CaO in olivine). In addition the most magnesian olivines were  $\geq Mg_{98}^{\#}$  in all cases and reached  $Mg_{91.5}^{\#}$  for Mauna Loa.

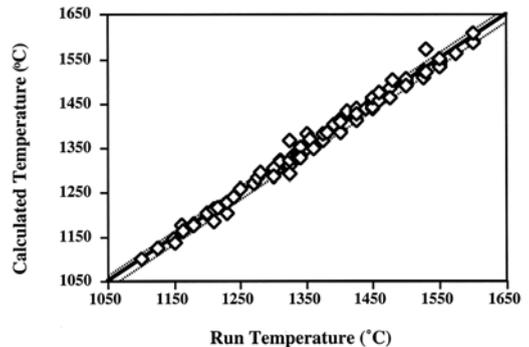


Fig. 1. Comparison of calculated liquidus temperature vs experimental temperature using the Liquidus Temperature vs. Composition relationships of Ford *et al.* (1983). Data are from experiments for liquids with  $\leq 3.0$  %  $Na_2O$  and combine Fig. 7a and 7b of Falloon & Danyushevsky (2000).

Norman & Garcia inferred that the magnesian olivines were phenocrysts, and not xenocrysts from disrupted lithospheric xenoliths, on the basis of the CaO vs.  $Mg^{\#}$  correlations [implying equilibrium between host magma and olivine and contrasting with the low CaO content (0.05 to 0.1 % CaO) in olivine of spinel lherzolite xenoliths]. The authors inferred that the primary or parental magmas were picritic with 15-16 % MgO in the mantle-derived liquids.

It is possible to use the data on Hawaiian picrites to evaluate the most primitive magmas for each volcano and to determine their temperatures of eruption. There is now a large body of experimental data on liquidus temperatures of basaltic magmas at 1 atm. and at high pressure. If we restrict ourselves to basaltic liquids with less than 2.5 %  $Na_2O$  (*i.e.* largely tholeiitic liquids) the formulation of a relationship between liquidus temperature and the major element composition of liquid [the 'Ford' geothermometer in Ford *et al.* (1983)] allows the estimation of anhydrous liquidus temperature for tholeiitic liquids, directly from their major element content (Fig. 1:  $T_{\text{Ford Thermometer}}$  vs.  $T_{\text{Experimental}}$ ). From the Fe/Mg partitioning relationship between olivine and liquid (Roedder & Emslie, 1970), it is also possible to estimate the liquidus olivine composition for a particular bulk-rock (picrite) or glass composition.

In Table 1, we have listed each of the picrite compositions for which olivine phenocryst compositions are provided by Norman & Garcia (1999). In column 2, the MgO content of the bulk rock is given and column 3 lists the composition of

Table 1. Compositional data from Hawaiian picrites, as given by Norman &amp; Garcia (1999), with their specimen numbers.

HAWAIIAN PICRITES	MgO <sub>Rock</sub> ↔ Olivine Mg <sup>#</sup> liquidus		Most magnesian olivine observed (Mg <sup>#</sup> )	Inferred primary magma MgO <sub>Primary</sub> Melt	Anhydrous liquidus Temperature of primary magma	La/Yb
<b>LOIHI</b>						
LO-02-2	24.2	93.9	88.0	13.5	1329	7.5
LO-01-04F	23.4	93.6	88.5	14.6	1347	7.8
LO-02-04C	29.6	95.3	88.5	15.0	1358	7.9
<b>KILAUEA</b>						
Glass *	14.7	89.6	90.7	16.6	1381	
KIL-1-18	13.8	88.7	90.3	16.0	1370	5.8
<b>HUALALAI</b>						
H-5	14.9	89.7	90.3	15.9	1368	4.4
H-23	25.8	94.3	89.5	16.1	1371	4.2
<b>MAUNA LOA</b>						
MLKAH-2	16.5	91.2	91.3	17.0	1386	3.9
ML-1868	21.5	92.9	91.3	18.4	1410	4.8
<b>KOHALA</b>						
KOH-1-28	20.5	92.7	90.1	15.8	1364	4.6
<b>KOOLAU</b>						
KOO-CF	15.4	90.0	89.2	14.4	1351	7.7
KOO-17a	20.9	92.5	89.2	15.3	1357	6.3
				Average	1365°C	

\*Clague *et al.* (1991).

the liquidus olivine for that bulk composition. In column 4, the most magnesian olivine phenocryst actually observed in the rock is listed. In the Loihi examples, the extremely magnesian bulk compositions have liquidus olivine composition of Mg<sup>#</sup> = 93.6 to 95.3 but the most magnesian olivine observed is Mg<sup>#</sup><sub>88</sub> (LO-02-2) or Mg<sup>#</sup><sub>88.5</sub> (LO-01-04F, LO-02-04C). Olivine of this composition is incrementally subtracted from the bulk to obtain a new composition (the inferred primary magma) which has liquidus olivine matching the observed most magnesian phenocryst. This calculation assumes an oxidation state of QFM + 0.5 log units and incrementally extracts equilibrium olivine in 0.01 % increments using the PETROLOG program (Danyushevsky, 1998). In column 5, the MgO of the inferred primary magma is listed and, using the Ford geothermometer, the liquidus temperature at 1 atm. is calculated for the three Loihi examples (1329°C, 1347°C, 1358°C). It should be

noted that if these particular picrites had actually evolved from a primary picrite with more magnesian liquidus olivine (*e.g.*, Mg<sup>#</sup><sub>90</sub>) then the calculation should have used Mg<sup>#</sup><sub>90</sub> rather than Mg<sup>#</sup><sub>88</sub> and proceeded with olivine extraction only to reach the composition crystallizing Mg<sup>#</sup><sub>90</sub>. Thus for the Loihi case, primary liquids may have been richer in MgO (and with higher liquidus temperatures ~ 1380°C). For the Kilauean picritic glass (Clague *et al.*, 1991), the liquidus olivine for the glass composition (14.7 % MgO) is Mg<sup>#</sup><sub>89.6</sub> but the most magnesian phenocryst is Mg<sup>#</sup><sub>90.7</sub>. In this case, olivine, of composition defined by the K<sub>D</sub><sup>Fe/Mg</sup><sub>Oliq</sub> relationship, is *added* until a composition with liquidus olivine Mg<sup>#</sup><sub>90.7</sub> is reached. This liquid has 16.6 % MgO and an anhydrous liquidus temperature of 1381°C. Calculations for Hualalai, Mauna Loa, Kohala and Koolau give primary liquids with 14.4-18.4 % MgO and anhydrous liquidus temperatures from 1351°C to 1410°C.

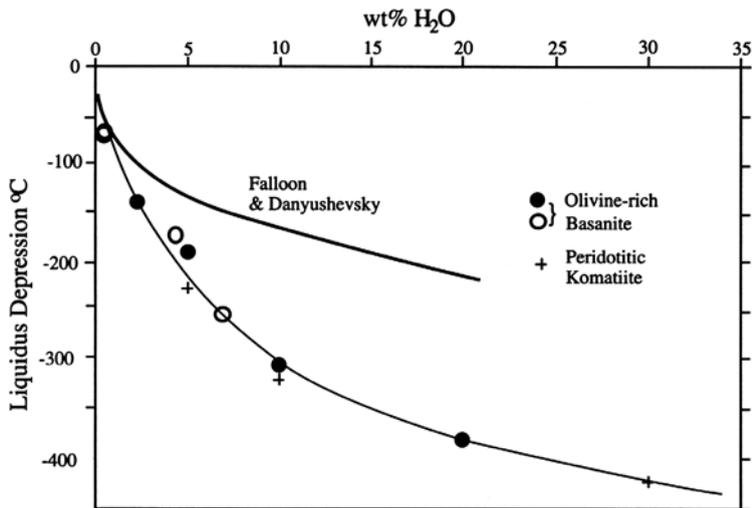


Fig. 2. Role of water-content in silicate melts in depressing liquidus temperature. Heavy curve is from Falloon & Danyushevsky (2000), compiled from literature data in which water content in glass is determined directly. Lighter curve is for olivine-rich compositions which do not readily quench to glass and in which data are obtained by bracketing liquidus temperatures for known, added water contents. Olivine-rich basanite data from Green (1973) (solid dots, based on liquids for particular water contents) and unpublished data (open circles, mid-point of temperature interval between liquid and olivine + liquid experiments for 1 %, 4.5 %, 7 % H<sub>2</sub>O). Peridotitic komatiite data from Green *et al.* (1975). Most of the data plotted in Falloon & Danyushevsky (2000) are derived from more polymerized (*i.e.* normative olivine + hypersthene, or hypersthene + quartz) liquids than the olivine-rich basanite or peridotitic komatiite and it is possible that water has a greater effect on liquidus depression in olivine-rich liquids.

The anhydrous liquidus temperatures average 1365°C but Hawaiian tholeiitic magmas have water contents of 0.3–0.8 % H<sub>2</sub>O (Garcia *et al.*, 1989; Michael, 1995) and this has a significant effect on liquidus temperatures. Apart from the picritic glass from Kilauea (Clague *et al.*, 1991), we do not have analyzed water contents of the picrites of Table 1. Correlations between H<sub>2</sub>O and other incompatible elements, including the degree of LREE enrichment, have been noted (Danyushevsky *et al.*, 2000; Michael, 1995) and to a first approximation we have taken the H<sub>2</sub>O-content of primitive Hawaiian picrites to be 0.5 wt%. This water content would give a liquidus depression of ~ 50°C (Fig. 2). Taking the average anhydrous liquidus of 1365°C (Table 1) we infer eruption temperatures of 1315°C for Hawaiian parental picrites.

To summarise, the mineralogy and bulk composition of Hawaiian tholeiitic picrites provides evidence for primary magmas of around 16 % MgO [13.5–18.4 % MgO] with eruption temperatures of ~ 1315°C. These are all magmas with characteristic LREE enrichment (column 7, Table 1). In a later section we consider the melting conditions for these magmas.

## B. Mid Ocean Ridge Basalts

On the basis of experimental studies of liquidus phases of the most magnesian MORB glasses and of the melting behaviour of a model MORB pyrolyte, we have argued that the parental or primary magmas in the mid-ocean ridge setting are picritic tholeiites with ≥ 13–14 % MgO. This argument can now be supported by data available on glass and olivine phenocryst compositions in dredged picrites/olivine tholeiites. In Table 2 we present an analysis of the temperatures of MORB magmas of N-MORB type (*i.e.* La/Yb < 1, column 7) and E-MORB type (La/Yb > 1). Because of the rapid quenching under pressure (sea-floor) of these magmas some of the glass compositions are very magnesian (column 1). The liquidus olivine for these glasses is listed in column 2, assuming an oxidation state of QFM – 0.5 log units (Ballhaus *et al.*, 1991). However, in each case, there are olivine phenocrysts of more magnesian composition, from Mg<sub>91.5</sub> to Mg<sub>92.1</sub>. For each of these glass compositions, liquidus olivine may be added incrementally (using 0.01 wt% increment of equilibrium olivine and PETROLOG program, Danyushevsky *et al.*, 2000) until the modified

Table 2. Compositional data from Mid-Ocean Ridge picrites in which Mg-rich glass and olivine phenocryst compositions are available.

MORB – GLASS*	MgO <sub>Glass</sub> ↔Olivine Mg <sup>#</sup> liquidus		Most magnesian olivine observed (Mg <sup>#</sup> )	Inferred 'primary' melt MgO <sub>Primary</sub> Melt	Anhydrous liquidus Temperature of primary melt	La/Yb
<b>N-MORB</b>						
148–896A-27R–1, 15 <sup>a</sup>	9.4	86.9	91.6	15.5	1357	0.38
148–896A-3R–1, 4 <sup>a</sup>	9.4	87.2	91.6	15.1	1350	0.30
D20-20 <sup>b</sup>	10.5	89.9	91.5	13.0	1319	0.43
DSDP 3–18 <sup>c</sup>	10.1	89.0	91.5	13.2	<u>1315</u>	0.47
				Average	1335°C	
<b>E-MORB</b>						
ARP–73–10–03a <sup>d</sup>	9.5	88.6	92.1	14.6	1340	–
CH 31-DR08–695 <sup>d</sup>	9.11	87.2	92.1	15.6	1362	–
Alvin 504–2 <sup>e</sup>	9.4	86.9	91.5	15.4	1360	–
Alvin 525–2 <sup>e</sup>	10.0	87.9	91.5	15.1	<u>1357</u>	1.94
				Average	1355°C	

\* References to sample numbers are:

a McNeill &amp; Danyushevsky (1996)

b Perfit *et al.* (1996)c Green *et al.* (1979)

d Kamenetsky (1996)

e Dmitriev *et al.* (1984)

melt composition has liquidus olivine matching the observed phenocryst of maximum forsterite content. The resultant compositions (MgO-content) are listed in column 4 (13.0–15.5 % MgO) and their anhydrous liquidus temperatures in column 5 (1319°C to 1362°C). In MORB, water contents and K<sub>2</sub>O contents are closely correlated (Danyushevsky *et al.*, 2000). The effect of water in lowering liquidus temperatures is 0–15°C in N-MORB as measured water contents (and K<sub>2</sub>O contents) are ≤ 0.1 %. E-MORB have higher K<sub>2</sub>O and H<sub>2</sub>O contents (Danyushevsky *et al.*, 2000) and the effect of dissolved water is to lower liquidus temperatures by ~ 30°C. For the data set of Table 2, the average anhydrous liquidus temperature is 1345°C and, considering a small effect of dissolved water, the inferred eruption temperature for the most primitive MORB magmas, which are picrites, is around 1325°C.

### Comparison of 'Hot-Spot' vs. 'MORB' primary magma temperatures

From the data of Tables 1 and 2, and the preceding discussion, we infer that the temperatures of eruption of primary magmas at the Hawaiian 'Hot-Spot' (cone-building phase) are the same, within ~ 20°C, as those of primary magmas in mid-ocean-ridge settings, including both N-

MORB and E-MORB types. The slightly higher anhydrous liquidus temperatures of the 'Hot-Spot' magmas are reduced by higher volatile contents (particularly H<sub>2</sub>O) in 'Hot-Spot' magmas relative to E-MORB and particularly to N-MORB.

### Compositional differences — MORB vs. Hot-Spot magmas

If the evidence from the primitive magmas themselves argues against a significant (> 20°C approximately) temperature difference between the product of MORB petrogenesis and 'Hot-Spot' petrogenesis, then it is important to evaluate the nature and causes of compositional differences between the two settings. Relative to MORB, 'Hot-Spot' basalts have long been equated with higher incompatible element contents, with LREE enrichment and HREE depletion, and with differences in isotopic ratios, from Pb-isotopes to noble gas isotopes. Some isotopic differences imply separation between MORB sources and 'Hot-Spot' sources for periods in excess of 1 billion years (Norman & Garcia, 1999 and references therein). The contrast between MORB and 'Hot-Spot' or OIB (Ocean Island Basalt) geochemistry has been made more complex by the recognition of E-MORB (Enriched MORB) geochemistry at mid-ocean ridges, and by recognition that OIB

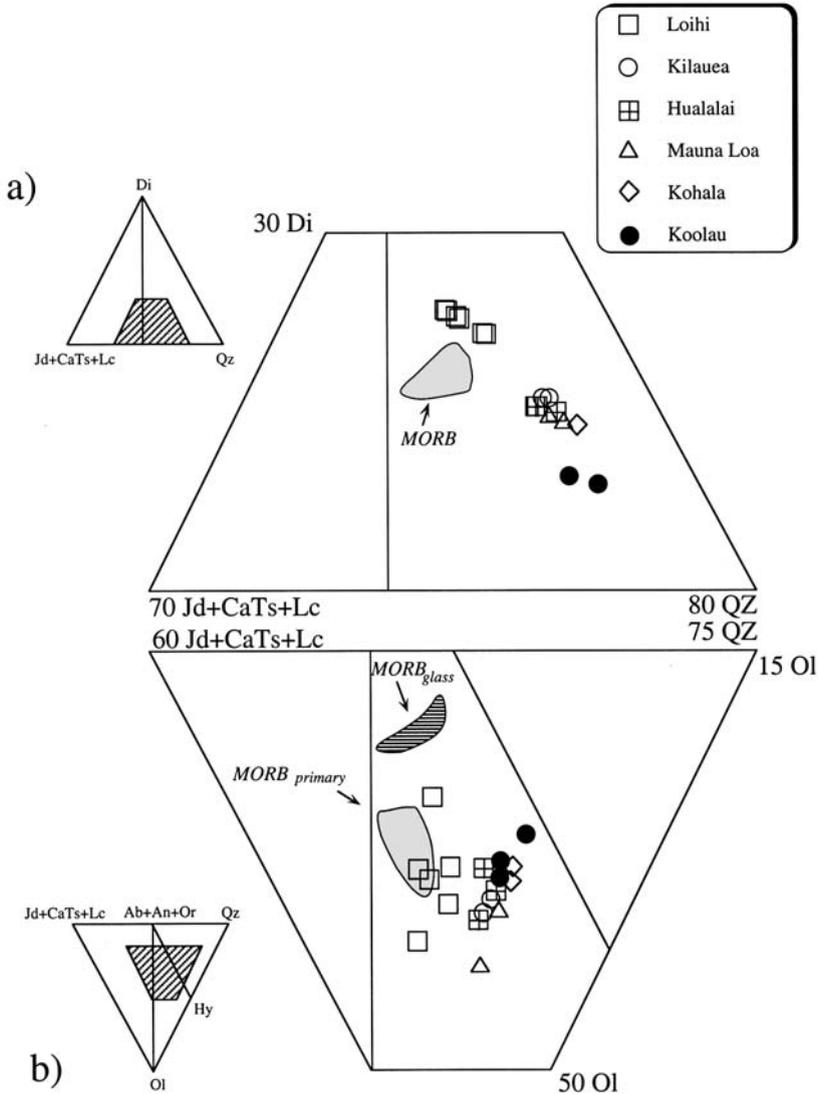


Fig. 3. Comparison of compositions of primitive MOR picrites of Table 2 and primitive ‘Hot Spot’ picrites of Table 1 using projections into the normative ‘basalt tetrahedron’ [Ol+Di+Qz+(Jd+CaTs+Lc)]. Inserts show selected areas of the tetrahedral faces. Projections are from:

(a) Olivine on to the surface Di+(Jd+CaTs+Lc)+Qz; and

(b) from Diopside on to the surface Ol+(Jd+CaTs+Lc)+Qz.

The MOR picrites form a tight cluster in both projections, noting that the olivine-addition calculations from the MORB glasses gives a field elongated along the lherzolite melting trend at 1.5 → 2 GPa and along a similar trend (1.5-2 GPa) in the projection from olivine (Fig.4).

The Hawaiian picrites show excellent clustering, particularly in the projection from olivine, and their trend and position clearly define a harzburgitic melting trend (Fig. 3a and Fig. 4).

geochemical characteristics are evident in intraplate basalts not linked to ‘Hot-Spot’ traces.

As noted previously, it is the differences in major-element compositions and in refractory-ele-

ment abundances and ratios of magmas, which are useful in terms of constraining the nature and composition of residual phases. We thus examine the major-element compositions of primitive basalts

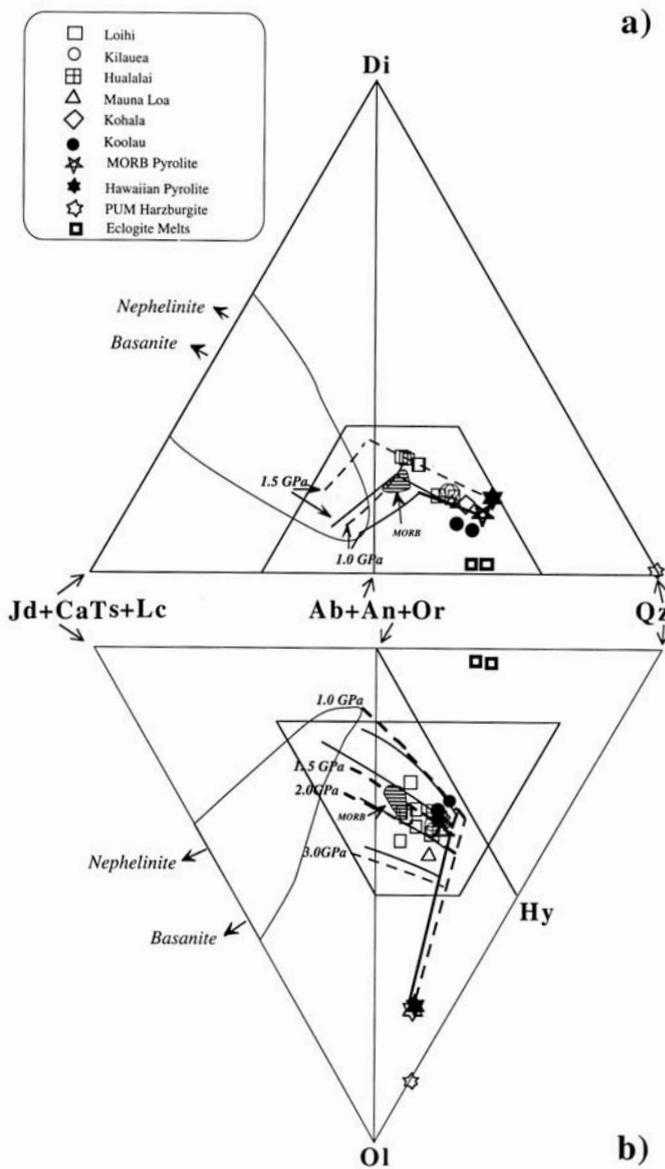


Fig. 4. Picrite compositions (Fig. 3) compared with melting trends for Hawaiian Pyrolite and MORB Pyrolite (Green & Falloon, 1998 and references therein). The MOR Picrites lie between the 1.5 to 2 GPa melting trends with lherzolite to harzburgite residue (cusp in Fig. 4a marking clinopyroxene disappearance from residue). Hawaiian Picrites (Loihi, Hawaii) lie on the harzburgite residue trend (4a) or in the harzburgite residue field (4b) for Hawaiian Pyrolite with Loihi and Hawaii possibly derived from a similar source plotting near the Hawaiian Pyrolite composition but Loihi representing a lower degree of partial melting. Koolau picrites clearly require a different source and residue which, from Fig. 4a, must have a lower Ca/Al and may have lower Ca/Na ratio than MORB Pyrolite or Hawaiian Pyrolite. In Fig. 4a and 4b we have plotted (open star) the composition of extremely refractory harzburgite from Papuan Ultramafic Belt (PUM harzburgite) as illustrative of a potentially buoyant subducted slab/wedge composition (Olivine is  $Mg_{92-3}^{\#}$ , Spinel is  $Cr_{80-90}^{\#}$ ). We have also plotted the composition of a rhyodacite melt derived from coesite eclogite at 3.5 GPa, 1250°C and 1300°C (Yaxley & Green, 1998). Finally, we have plotted the field of high  $Mg^{\#}$  basanitic and nephelinitic intraplate basalts which are host to spinel or spinel + garnet lherzolite xenoliths. These compositions are small melt fractions (< 5 % melt) from peridotite (C+H+O) in the incipient melt regime, in equilibrium with Ol+Opx+Cpx+Ga. Source mantle peridotites for Hawaiian Hot Spot magmas can be derived by refertilizing refractory mantle by addition of rhyodacite (eclogite melt) and basanite/olivine nephelinite (peridotite-(C+H+O) incipient melt), with or without asthenospheric mantle (MORB Pyrolite).

for compatibility with the results of high-pressure melting studies. For this purpose, the calculated primary magma compositions are projected into the basalt tetrahedron (Yoder & Tilley, 1962) using a normative projection (Fig. 3, Green & Falloon, 1998).

The MORB glasses of Table 2, when projected from diopside on to the base of the basalt tetrahedron [Ol+(Jd+Cats+Lc)+Qz] define a small field, within the larger field of MORB glasses with  $Mg^{\#} > 68$  defined by Falloon & Green (1988). The calculation of primary liquids based on the most magnesian olivine phenocrysts moves this field to higher normative olivine compositions, again lying within the field of primary MOR picrites defined by Falloon & Green (1987, 1988). In the projection from olivine, the glasses and estimated primary magmas are coincident and define a small field near the 15-20 % normative diopside contour. When the experimentally defined partial melting trends for MORB pyrolite are plotted on the normative tetrahedron and from mass balance calculations, it is clear that these inferred primary magma compositions are consistent with derivation from the MORB pyrolite composition by ~20-25 % melting and magma segregation at 1.5-2.0 GPa leaving a clinopyroxene-bearing harzburgite residue.

With higher incompatible element contents in 'Hot-Spot' or OIB basalts and noting their relative HREE-depletion and LREE enrichment, it is commonly inferred that these magmas derive from deeper melting processes with compositions reflecting equilibrium with garnet as a residual phase. In terms of plotting in the normative basalt tetrahedron, higher-pressure liquids in equilibrium with garnet lherzolite will lie to the silica-undersaturated side of the Ol+Di+Plag plane and further towards the olivine apex in the projection from diopside. As shown in Fig. 4, this is the field of mantle xenolith-bearing intra-plate basalt (alkali olivine basalts, olivine-rich basanites, to olivine nephelinites and olivine melilitites) and experimental studies of the liquidus phases of selected high-Mg basalts, with added  $H_2O \pm CO_2$ , have established conditions for genesis of such melts by low degrees of partial melting at pressures  $\geq 2$  GPa (e.g. Frey *et al.*, 1978).

It is therefore somewhat unexpected that the inferred primary magmas for the Hawaiian picrites plot to the silica-rich side of the Ol+Di+Plag plane of the normative tetrahedron. It is particularly notable that each volcano forms a tight cluster in the projection from olivine, but a scatter largely in the direction of olivine addition or subtraction in

the projection from diopside. This scatter is within the harzburgite-residue field (*i.e.* close to the Ol-Qz-Di face of the tetrahedron rather than in the lherzolite-residue field which lies across the Ol-Plag-Di plane. In the projection from olivine, the compositions from each volcano consistently define trends which we can identify from partial melting studies as 'harzburgite-residue melting trends'. In the projection from olivine this trend is almost perpendicular to the 'lherzolite melting trend' which controls melt composition with increasing degree of melting up to the elimination of Ca-clinopyroxene. The dispersion along olivine-control lines in the projection from diopside suggests that the olivine-addition and subtraction methods applied to the picrites is capable of improvement, e.g. for Loihi samples, the most magnesian olivine observed at  $Mg_{88.5}^{\#}$  might have been re-equilibrated and a more magnesian source  $Mg_{90-91}^{\#}$  required. The plotting position in the projection from olivine is however 'blind' to this uncertainty.

In Fig. 4, the primary picrites from Hawaii are compared with the melting trends from Hawaiian pyrolite and from MORB pyrolite. Loihi and the four volcanoes from the main island, Hawaii, plot on or very close to a harzburgite-residue melting trend with Loihi representing a smaller degree of melting and possibly lying at the clinopyroxene-out cusp. On the other hand, Koolau picrites are consistent with a harzburgite-residue melting trend but for a source lherzolite with a lower Ca/Al ratio than MORB pyrolite or Hawaiian pyrolite (evident from the plotting position in the olivine projection). In the projection from diopside, there is a suggestion that melt-segregation may have occurred at 1.0-1.5 GPa rather than 1.5-2.0 GPa, inferred from the plotting positions of the Loihi and Hawaii (main island) data.

Finally, experimental studies of the liquidus phases of primitive N-MORB glasses (particularly DSDP 3-18) and of olivine-enriched derivatives, established saturation in olivine followed by clinopyroxene to ~1.8 GPa and olivine+orthopyroxene followed by clinopyroxene at 2.0 GPa (Green *et al.*, 1979; Green & Falloon, 1998). These liquids are thus consistent with lherzolite residues. In contrast, experimental study of Kilauean tholeiitic picrite showed olivine followed by orthopyroxene at 1-1.5 GPa and orthopyroxene alone as a liquidus phase at 1.8 GPa (Green & Ringwood, 1967; Green, 1970; Eggins, 1992). Trace-element geochemists have used the relative LREE enrichment and HREE depletion of Hawaiian tholeiites to argue for gar-

net as a residual phase in melt extraction, requiring both a very small degree of partial melting and a high pressure of melt segregation. However, both the direct experimental studies and the fact that primitive Hawaiian tholeiitic picrites lie in the field for harzburgite, not lherzolite and particularly not garnet lherzolite, residues, establish that garnet is *not* a residual phase buffering the composition of primary Hawaiian picrites.

To summarize, the major-element composition of parental or primary MOR picrites (N- and E-MORB) show that the MOR picrites have lherzolite to harzburgite residues. However, the parental or primary Hawaiian picrites have harzburgite residues with Loihi magmas indicating a somewhat lower degree of melting at and slightly beyond clinopyroxene-out in the peridotite melting interval. Mauna Loa, Kilauea, Hualalai and Kohala all indicate more refractory harzburgite residues, but a similar source composition to Loihi, while the major-element composition of Koolau picrites indicates a distinctive source of relatively higher Al/Ca ratio.

### Further evidence of refractory character of Hawaiian source/residue relative to MORB source/residue

It has previously been noted that the CaO content (CaMgSiO<sub>4</sub> solid solution) of olivine phenocrysts in Hawaiian magmas increases in the order Koolau < Hawaii (Kilauea, Mauna Loa, Hualalai, Kohala) < Loihi (Norman & Garcia, 1999). This parameter reflects the normative diopside content of the magmas (Fig. 3 and 4). Norman & Garcia (1999, Fig. 3) also showed that in a plot of Cr<sup>#</sup> vs. Mg<sup>#</sup> of spinel phenocrysts and spinels included in olivine, the Cr<sup>#</sup> of Hawaiian spinels is 65-75, contrasting with MORB spinels at Cr<sup>#</sup> ~ 10-60 and approaching the spinel in high Ca boninites (Cr<sup>#</sup> = 70-80) and low-Ca boninites (Cr<sup>#</sup> = 80-90). In direct melting studies of spinel lherzolite (Jaques & Green, 1980) residual spinel increases from Cr<sup>#</sup> = 30 to Cr<sup>#</sup> = 80 with increasing degree of melting at 1 and 1.5 GPa. Thus the higher Cr<sup>#</sup> of liquidus spinels for Hawaiian picrites relative to MOR picrites is strong supporting evidence that the source peridotite for Hawaiian picrites was a more refractory composition in terms of Cr/Al ratio than the source lherzolite for MOR picrites. Alternatively, if source peridotites were similar in composition, then the Hawaiian picrites require a much higher degree of partial melting, a conclusion inconsistent with their relative incompatible element contents.

In addition to the difference in Cr/Al, Ballhaus *et al.* (1991) and Norman & Garcia (1999) demonstrate that the Hawaiian spinel compositions indicate  $f_{O_2}$  conditions near QFM + 0.5 log units, slightly more oxidized than spinels of MORB (Ballhaus *et al.*, 1991).

### Thermal vs. compositional anomaly for Hawaiian 'plume'

One of the major arguments for the existence of a deep-seated thermal plume beneath Hawaii has been the observation that the Hawaiian Island chain lies on the crest of a broad topographic swell elevated approximately 1 km above the surrounding ocean floor. It has been inferred that the topographic high is a consequence of lower mean density in the upwelling column (a buoyancy plume) and that the lower density is due to higher temperatures (White & McKenzie, 1989). A temperature difference of ~ 200°C is required to produce a density difference  $\Delta\rho = 0.02$ . A similar density difference is produced by a compositional difference in olivine from Mg<sub>80</sub><sup>#</sup> to Mg<sub>91</sub><sup>#</sup>.

If the composition of average asthenospheric mantle is close to the MORB pyrolite estimate, based on natural lherzolites with 3-4 % CaO, Al<sub>2</sub>O<sub>3</sub>, Mg<sup>#</sup> ≈ 89 then the density (at room T,P) of a garnet lherzolite assemblage is  $\rho_o = 3.35$ -3.36, Mg<sub>91</sub><sup>#</sup> = 89.5. After extraction of ~ 20 % MOR picrite to form oceanic crust and lithosphere, the residual lherzolite in the lithosphere has ~ 1.5-2 % CaO, Al<sub>2</sub>O<sub>3</sub> and the density for garnet lherzolite is  $\rho_o = 3.33$ , Mg<sub>91</sub><sup>#</sup> = 90. We also know that in the convergent margin environment, magmas such as back-arc basin tholeiites, high-Ca and low-Ca boninites are formed and the latter in particular leave extremely refractory harzburgite with magnesian olivine Mg<sub>92.5</sub><sup>#</sup>, low Al<sub>2</sub>O<sub>3</sub>, low CaO, orthopyroxene (Mg<sub>92.5</sub><sup>#</sup>) and chromian spinel (Cr<sup>#</sup> = 80-90). Such harzburgites are characteristic of ophiolitic ultramafics of Western Pacific type as illustrated in Papua New Guinea, New Caledonia, Philippines, *etc.* Garnet does not form in such refractory harzburgite at high pressure (below the enstatite to majorite reaction) and  $\rho_o = 3.30$ , Mg<sub>91</sub><sup>#</sup> = 92.5. Thus, in the subduction of oceanic crust and lithosphere, including elements of the overlying mantle wedge, bodies of refractory lherzolite to harzburgite may be returned to the mantle with both a temperature contrast (cooler) and compositional contrast ( $\Delta\rho = 0.02$  to 0.06). The buoyancy of such slabs, relative to asthenospheric mantle, will be determined by compositional factors and

their approach to temperature equilibrium with surrounding mantle. Such slabs may become detached and suspended in the upper mantle.

We suggest that subduction of cool lithosphere in the slab-wedge environment of Western Pacific type has left residues of inhomogeneous harzburgitic to lherzolitic mantle which become neutral (suspended) or buoyant within the upper mantle. Compositional buoyancy implying suspended slabs of lithospheric mantle at 150 to 500 km depth is proposed as an alternative cause of the Hawaiian Swell.

### Melting of old, neutrally buoyant subducted slabs

If detached slabs of refractory lithosphere and former oceanic crust become suspended in the upper mantle then partial melting may be initiated in several ways:

a) Conductive heating will, given time, erase the temperature contrast between slab and ambient

mantle. Compositions with lower solidus temperatures will melt preferentially and such partial melts may migrate causing compositional and temperature redistribution.

b) A redox contrast between subducted slab (IW + 3 to 4 log units) and asthenospheric or deeper mantle (IW + 1 log unit) is a potential focus for melting (Green *et al.*, 1987). If normal asthenosphere has small water and carbon contents [as inferred from measurable C&H contents of MORB] then at = IW + 1 log unit a potential mobile fluid phase is dominated by (CH<sub>4</sub> + H<sub>2</sub>O) (Green *et al.*, 1987) at temperatures near the peridotite solidus and pressures  $\geq 3$  GPa. If such a mobile fluid phase encounters a volume of mantle with higher  $f_{O_2}$ , then oxygen is extracted from the mantle volume, carbon is precipitated and water content and activity in the fluid phase is increased. Since the solidus of lherzolite, harzburgite or eclogite is sharply depressed as  $f_{H_2O}$  in a fluid increases, the redox front becomes a locus for incipient melting in the peridotite-(C+H+O) system.

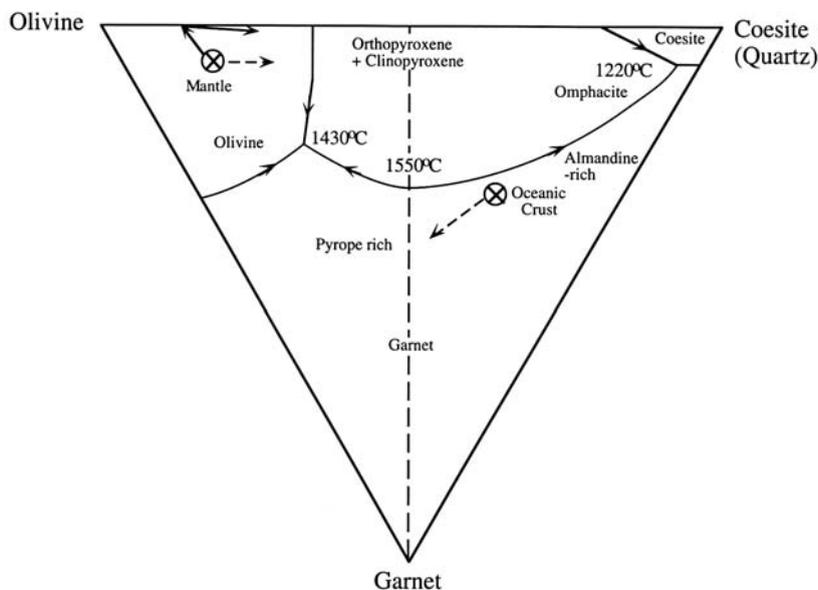


Fig. 5. Projection from Clinopyroxene (Jadeite+Diopside) on to plane Olivine + Garnet + Coesite in the high pressure normative tetrahedron (Yaxley & Green, 1998). The figure illustrates extraction of picritic melts lying near the lherzolite minimum melt for the composition marked 'Mantle'. Residues approach the Olivine + Orthopyroxene join. On the coesite-normative side of the figure the composition of average Oceanic Crust is plotted. In an inhomogeneous mantle of recycled lithosphere (mantle + residue) = Oceanic Crust which is being re-heated or is part of a Mantle Plume, melting first begins at the (coesite+omphacite+garnet) minimum ( $\sim 1220^\circ\text{C}$  at 35 GPa) and the melt is rhyodacitic in composition. Extraction of such melts drives the residue composition on to the garnet + pyroxene thermal divide and migration of melt into, and reaction with or crystallization within, surrounding peridotite moves the bulk compositions of the 'Mantle' and 'Residue' in the direction illustrated.

c) As the refractory slab becomes warmer by conductive heating, its density will decrease and it may become positively buoyant. If the volume is large, then a diapir or plume may begin to ascend — partial melting will occur by adiabatic decompression and will occur selectively according to the chemical inhomogeneity of the plume or diapir.

The key melting relationships of a mixed mantle region at depths of  $\sim 120$  km (3.5 GPa) are illustrated in Fig. 5 in which compositions are presented in terms of high-pressure normative components reflecting high-pressure mineralogy, *i.e.* olivine, orthopyroxene, clinopyroxene (including jadeite solid solution), garnet and coesite. The pyroxene-garnet plane divides silica-normative compositions (basaltic, andesitic, dacitic compositions) from olivine-normative compositions (olivine-rich picrites, olivine nephelinites and peridotites). Referring to Fig. 5, at 3.5 GPa an average oceanic crust composition has subsolidus mineralogy of garnet + clinopyroxene (omphacite) + coesite. At 1250°C, this composition consists of dacitic to rhyodacitic melt with residual garnet + omphacite. The melt is strongly reactive towards enclosing peridotite and will react and freeze at or near the eclogite/peridotite contact. The enclosing peridotite is then locally enriched in orthopyroxene, garnet and clinopyroxene — refertilized lherzolite with distinctive low Ca/Al and Ti/Al and relatively high Na/Ca. (Compositional vectors indicating extraction of rhyodacitic melt from oceanic crust and addition to pyrolytic mantle or to residual harzburgitic mantle — after basalt/picrite extraction — are shown in Fig. 5 and may be inferred from the plotted positions of eclogite melts, pyrolite and harzburgite on Fig. 4.)

With increasing temperature, the melt composition in the eclogite component is constrained to lie on the garnet + clinopyroxene  $\pm$  orthopyroxene cotectic (Fig. 5) but cannot mix across the pyroxene + garnet plane. This plane acts as a thermal divide with residual garnet + clinopyroxene  $\pm$  orthopyroxene from the original eclogite becoming progressively more Mg-rich and jadeite-poor respectively as the dacitic to andesitic melt leaves the eclogite to react with and freeze within the enclosing peridotite. This type of enrichment or refertilisation of refractory harzburgite or lherzolite is suggested as the mechanism for generating the distinctive Koolau source, identified previously.

If the temperature reaches  $\sim 1420^\circ\text{C}$  then melting occurs at the olivine + pyroxene + garnet minimum in the peridotite. Melts are picritic

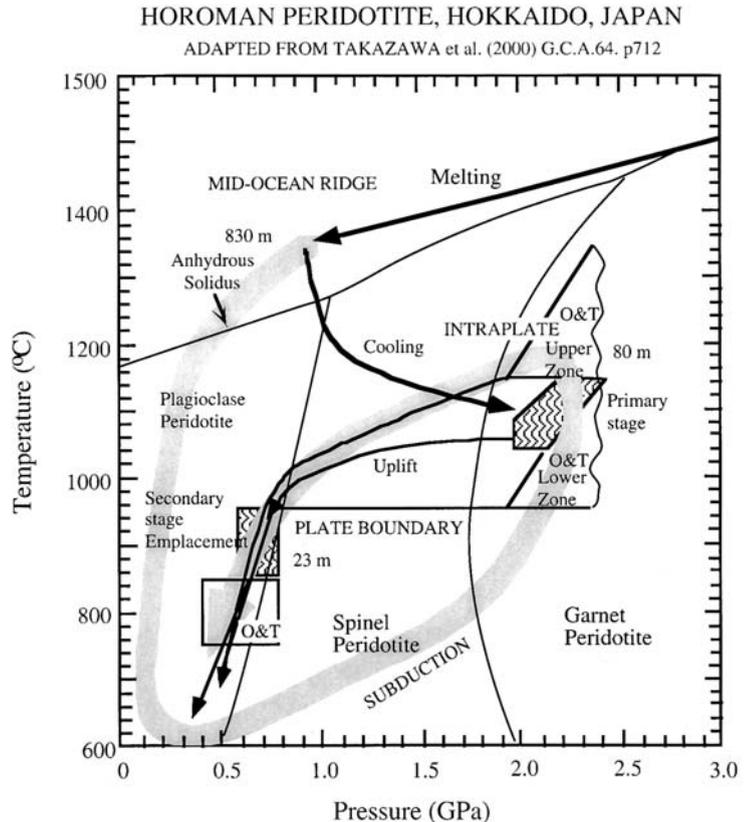
nephelinites, controlled by the coexistence of olivine + orthopyroxene + clinopyroxene + garnet. This type of refertilisation, by migration of incipient melt of picritic nephelinite character, may occur in an intraplate setting and initiate the buoyancy by partial melting and detachment of a diapir or plume from the larger source (slab) or its margins.

In an evolving regime of temperature increase at high pressure or in adiabatic decompression at  $T_p \sim 1450^\circ\text{C}$ , the differential melting behaviour between coesite eclogite and enclosing harzburgite or lherzolite, will lead towards homogenization of mineral compositions in compatible and incompatible elements (but not in strongly refractory components such as  $\text{Cr}_2\text{O}_3$ , NiO). However, the original heterogeneity of the subducted oceanic crust, refractory lithosphere and enclosing mantle will remain evident in modal variations, *e.g.* an original coesite eclogite with  $\text{Mg}^\# \sim 65$  may finally become a garnet ( $\text{Mg}_{83}$ ) + diopside ( $\text{Mg}_{91}$ ) lens or layer, contrasting with surrounding garnet ( $\text{Mg}_{83}$ ) lherzolite or harzburgite in which garnet is Cr-rich and diopside rich in Na + Cr.

If, during adiabatic upwelling of a plume or diapir of inhomogeneous, recycled mantle of harzburgite to garnet pyroxenite composition, the garnet lherzolite solidus is exceeded and melting increases with decompression/upwelling, then the continuing presence of melt trapped within the plume or diapir will tend to homogenize phase compositions, but not residual phase proportions. With increasing melting at decreasing pressure, melts are constrained to lie on the garnet + clinopyroxene + orthopyroxene + olivine cotectic for low degrees of melting at high pressure, adjusting to the (Cr+Al) spinel + clinopyroxene + orthopyroxene + olivine (lherzolite) cotectic at higher degrees of melting at lower pressure, and then to the spinel (Cr-rich) + orthopyroxene + olivine (harzburgite residue) for even higher degrees of melting. The positions of these cotectics in the normative basalt tetrahedron have been defined by experimental studies and reflect both bulk composition and P,T control (Fig. 4).

In relation to the preceding discussion, it is relevant to note recent work on the Horoman Peridotite in Hokkaido, Japan. Takazawa *et al.* (2000) have inferred a complex P, T, time history which is summarized in Fig. 6. This history is of primary crust/lithosphere formation followed by subduction and modification by melting [including both selective melting of eclogite (slab and wedge environment) and incipient melting in garnet-lherzolite regime (intraplate setting)]. This

Fig. 6. The possible Pressure-Temperature-time history of Horoman Peridotite as inferred by Takazawa *et al.* (2000). Figure modified from Takazawa *et al.* (2000) Fig.17, as an illustration of a peridotite with a long history from primary upwelling, through subduction and refertilization to uplift and reactions in a low pressure and high temperature emplacement.



evolution from refractory mantle to re-enriched, inhomogeneous mantle is followed by upwelling and incipient melting in the plagioclase lherzolite stability field. The sequence of events proposed for Horoman Peridotite has similarities with the model proposed here for the Hawaiian source (reference also Green *et al.*, 1987, Fig. 6; Green & Falloon, 1998, Fig. 7.12).

## Conclusions

Examination of bulk compositions, glass compositions and olivine phenocrysts from Hawaiian picrites and from Mid-Ocean Ridge picrites (N-MORB and E-MORB) demonstrates that the liquidus temperatures of primary or parental magmas are approximately 1320°C in both settings. There is no evidence from the magmas themselves for a high temperature ( $\Delta T = 100\text{--}200^\circ\text{C}$ ) deep-seated mantle plume beneath Hawaii as picritic magmas, not olivine tholeiite magmas, are parental or primary magmas of both the Hawaiian 'Hot-Spot' and Mid-Ocean Ridges.

The primary picrites of both settings have major-element compositions consistent with equilibrium partial melting of lherzolite and magma segregation at 1.5-2.0 GPa, and approximately 1430-1450°C. The modern mantle has a potential temperature near 1430°C and this is a controlling factor in Mid-Ocean Ridge, 'Hot-Spot', Back-Arc and Island Arc magmatism.

Although we argue for similar mantle temperatures controlling Hawaiian ('Hot-Spot') volcanism and Mid-Ocean Ridge volcanism and for similar depths of magma segregation, there is clear evidence for compositional difference between the primary magmas and thus, between their sources. Hawaiian picrites have higher volatile ( $\text{CO}_2 + \text{H}_2\text{O}$ ) contents, higher incompatible element contents (LILE), higher LREE and lower HREE, than MOR picrites. Conversely, they have higher Cr/Al ratios in liquidus spinels and the most magnesian liquidus olivines are similar in both cases. We have demonstrated that Hawaiian picrites leave a more refractory harzburgite residue in contrast to a lherzolitic to harzburgitic residue for MOR picrites.

The major element, trace element and isotopic fingerprints of Hawaiian tholeiitic magmas appear to require a four-component mixing:

a) Residual, refractory harzburgite ( $Mg\#_{01} \sim 91-92$ ,  $Cr\#_p \sim 70$ );

b) Local refertilization of residual harzburgite in a subduction environment by partial melting of subducted oceanic crust giving pyroxene-rich haloes and garnet pyroxenite residues from initial coesite eclogite.

c) Migration of an incipient melt, characteristic of the peridotite–(C+H+O) system at  $P > 2.5$  GPa,  $T \sim 1250-1450^\circ\text{C}$  within the garnet stability field and at P,T conditions between the peridotite–(C+H+O) solidus and the (C+H)-free peridotite solidus.

d) Addition of ( $\text{CH}_4+\text{H}_2\text{O}$ ) fluid phase at  $f\text{O}_2 \approx \text{IW} + 1$  log unit, the fluid phase being derived from deeper levels on the mantle and probably carrying a noble-gas signature from deeper mantle.

The time intervals and tectonic setting for processes a) to d) can be explored by trace element and isotopic abundances. The initial stage (a) may be formation of lithosphere in a mid-ocean ridge environment or may reflect several melt extraction events culminating in boninite extraction in a convergent margin/mantle wedge setting. Stage (b) is envisaged as a consequence of prograde metamorphism and tectonic interleaving of subducted oceanic crust, underlying oceanic lithosphere and overlying mantle wedge lithosphere. Stage (c) is envisaged as an intraplate setting in which detached and neutrally buoyant ‘old-subducted slabs’ approach ambient temperature over time, and begin to melt at the peridotite–(C+H+O) solidus. Stage (d) is continuous over time if mantle degassing is pervasive, although a ( $\text{CH}_4+\text{H}_2\text{O}$ ) fluid is mobile at  $f\text{O}_2 \leq \text{IW} + 1$  but initiates melting at the peridotite–(C+H+O) solidus if it encounters oxidized mantle, such as old subducted slab. In this case carbon is deposited and  $\text{H}_2\text{O}$ -rich fluid initiates melting (Green *et al.*, 1987; Green & Falloon, 1998).

The model proposed suggests that the Hawaiian Swell or Arch is a response to density-contrast in or below the asthenosphere. The density contrast is interpreted as a consequence of a relict, more refractory but inhomogeneous harzburgitic volume of mantle (noting that  $\Delta\rho_0 = 0.01$  is equivalent to  $\Delta T \sim 100^\circ\text{C}$ ). Green *et al.* (1987) and Green & Falloon (1998) inferred that such compositional ‘plumes’ or anomalous mantle could form from old subducted slabs. Alternatively, delamination of refractory, deep continental lithosphere may provide the compositional anomaly.

Trace element and isotope geochemists have consistently advocated the mixing of several components in the source for ‘Hot-Spot’ basalts, representing reservoirs evolved independently over times up to 1-2 billion years (*cf.* Norman & Garcia, 1999 and references therein). In terms of the ‘deep mantle plume’ model, this concept of geochemically mixed source has been incorporated by accumulating subducted oceanic crust and lithosphere at the core/mantle boundary, reheating this material as a thermal boundary layer and ascent as thermal plumes from the core/mantle boundary (Davies, 1999).

In this paper we detect no temperature difference between magmatism at the type ‘Hot-Spot’ or ‘Plume tail’ expression, *i.e.* Hawaii, and Mid-Ocean Ridge settings. We argue that the cause of island chain volcanism is a compositional buoyancy anomaly at depths below or in the asthenosphere. Upwelling mantle plumes or diapirs feeding individual Hawaiian volcanoes are sourced in the interface between old subducted slabs or delaminated continental lithosphere, and ambient upper mantle (MORB source), probably at depths of 200-300 km, with incipient melting [peridotite–(C+H+O)] giving place to major melting during adiabatic upwelling from these depths. The ‘Hawaiian Plume’ is inferred to be a thermal and partial melting anomaly only at relatively shallow depths, particularly in penetrating into the oceanic lithosphere, but to have its origins in a long-lived compositional anomaly, at or near ambient mantle temperatures, within the asthenosphere of the Upper Mantle.

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Received 15 August 2000

Modified version received 16 January 2001

Accepted 31 January 2001

