

# *Crystallization temperatures of tholeiite parental liquids: Implications for the existence of thermally driven mantle plumes*

**Trevor J. Falloon\***

*School of Earth Sciences and Centre for Marine Science,  
University of Tasmania, Private Bag 79, Hobart, Tasmania 7001, Australia*

**David H. Green\***

*Research School of Earth Sciences, Australian National University,  
Mills Road, Acton, Canberra 0200, ACT, Australia*

**Leonid V. Danyushevsky\***

*School of Earth Sciences and Australian Research Council Centre for Excellence  
in Ore Deposits, University of Tasmania, Private Bag 79, Hobart, Tasmania 7001, Australia*

## ABSTRACT

To compare magmatic crystallization temperatures between ocean island basalt (OIB) proposed to be plume-related and normal mid-ocean ridge basalt (MORB) parental liquids, we have examined and compared in detail three representative magmatic suites from both ocean island (Hawaii, Iceland, and Réunion) and mid-ocean ridge settings (Cocos-Nazca, East Pacific Rise, and mid-Atlantic Ridge). For each suite we have good data on both glass and olivine phenocryst compositions, including volatile ( $H_2O$ ) contents. For each suite we have calculated parental liquid compositions at 0.2 GPa by incrementally adding olivine back into the glass compositions until a liquid in equilibrium with the most-magnesian olivine phenocryst composition is obtained. The results of these calculations demonstrate that there is very little difference (a maximum of  $\sim 20$  °C) between the crystallization temperatures of the parental liquids (MORB 1243–1351 °C versus OIB 1286–1372 °C) when volatile contents are taken into account.

To constrain the depths of origin in the mantle for the parental liquid compositions, we have performed experimental peridotite-reaction experiments at 1.8 and 2.0 Gpa, using the most magnesian of the calculated parental MORB liquids (Cocos-Nazca), and compared the others with relevant experimental data utilizing projections within the normative basalt tetrahedron. The mantle depths of origin determined for both the MORB and OIB suites are similar (MORB 1–2 GPa; OIB 1–2.5 GPa) using this approach.

Calculations of mantle potential temperatures ( $T_p$ ) are sensitive to assumed source compositions and the consequent degree of partial melting. For fertile lherzolite sources,  $T_p$  for MORB sources ranges from 1318 to 1488 °C, whereas  $T_p$  for ocean island tholeiite sources (Hawaii, Iceland, and Réunion) ranges from 1502 °C (Réunion) to 1565 °C (Hawaii). The differences in  $T_p$  values between the hottest MORB and ocean island tholeiite sources are  $\sim 80$  °C, significantly less than predicted by the ther-

\*E-mails: Falloon, trevor.falloon@utas.edu.au; Green, David.H.Green@anu.edu.au; Danyushevsky, l.dan@utas.edu.au.

**thermally driven mantle plume hypothesis. These differences disappear if the hotspot magmas are derived by smaller degrees of partial melting of a refertilized refractory source. Consequently the results of this study do not support the existence of thermally driven mantle plumes originating from the core-mantle boundary as the cause of ocean island magmatism.**

**Keywords:** MORB, Hawaii, Réunion, Iceland, Kilauea, olivine, peridotite melting experiments, mantle potential temperatures, primary magmas, mantle plumes.

## INTRODUCTION

The composition and temperature at which olivine crystallizes from a mantle-derived parental liquid at a low pressure is one of the key constraints on any model of magma genesis (e.g., Green and Ringwood, 1967; Sobolev and Danyushevsky, 1994), as these parameters are necessary for estimating potential temperatures of the source mantle. Olivine is the first phase to crystallize at low pressure from any mantle-derived melt that is in chemical equilibrium with peridotite at source depths. This is due to the rapid expansion of the olivine phase volume at lower pressures, demonstrated by numerous experimental studies on model melt compositions. As both olivine crystallization temperature and the composition of liquidus olivine are sensitive to the composition of the crystallizing melt, it is possible to calculate both, given the composition of melt alone. Conversely, given an olivine composition, it is possible to test whether an observed glass or rock composition is in equilibrium with the olivine, by using empirically calibrated functions for equilibria between melt and olivine end-members. Such a calibration is referred to as an olivine geothermometer (e.g., Roeder and Esslie, 1970; Ford et al., 1983; Herzberg and O'Hara, 2002).

The application of olivine geothermometry to determine the composition of parental liquids is of particular significance to our understanding of the causes of ocean island volcanism, including inferred hot mantle plumes. The hypothesis of thermally driven mantle plumes derived from the core-mantle boundary predicts a significant temperature contrast between upwelling plume material and ambient upper mantle. Consequently decompression melts derived from the mantle plume materials should be ~200–300 °C hotter than melts derived from ambient upper mantle (McKenzie and Bickle, 1988). If the thermally driven mantle plume hypothesis is correct, then we should expect to find evidence from olivine crystallization temperatures that parental liquids to Hawaii olivine tholeiites, a typical ocean island magma, are significantly hotter than parental liquids to mid-ocean ridge basalt (MORB) olivine tholeiites. We should also find evidence for very different pressures and degrees of partial melting. The determination of olivine crystallization temperatures of parental melts is also the first step necessary for estimating mantle potential temperatures ( $T_p$ ).

In this article, we calculate the parental liquids for a representative range of tholeiite compositions from both mid-ocean

ridge (Cocos-Nazca, East Pacific Rise, and mid-Atlantic Ridge) and ocean island settings (Hawaii, Réunion, and Iceland). We find that there are no significant differences (a maximum of ~20 °C) between crystallization temperatures of MORB and ocean island basalt (OIB) parental tholeiite liquids, when the hottest parental liquids from each setting are compared. To constrain the depth of origins in the mantle for the parental liquid compositions, we have performed experimental peridotite-reaction experiments at 1.8 and 2.0 GPa (see Appendix 1), using the most magnesian of the calculated parental MORB liquids (Cocos-Nazca), and compared the others with relevant experimental data utilizing projections within the normative basalt tetrahedron. The mantle depths of origin determined for both the MORB and OIB suites are similar (MORB 1–2 GPa; OIB 1–2.5 GPa) using this approach. Finally we present model calculations for the source mantle  $T_p$  of the calculated parental liquids. These calculations are significantly dependent on the chosen models of source compositions and thus on the inferred degree of partial melting and the magnitude of latent heat of melting in the models. Below we first outline the rationale of our approach before discussing in detail our case studies from Hawaii, Réunion, Iceland, and MORB.

## RATIONALE OF APPROACH

Our approach to estimating the temperatures of tholeiite magmas is based on the use of olivine geothermometers to calculate the compositions of parental liquids. Unmodified parental liquids rarely erupt, and evidence for their existence is only preserved in the mineralogy of magnesian olivine phenocrysts enclosed in phyric magma (rock) compositions. An olivine geothermometer is therefore used to reconstruct the composition of the parental liquid, and its temperature of crystallization, by adding back olivine in incremental equilibrium steps (e.g., 0.01 wt%; see appendix in Danyushevsky et al., 2000, for a detailed explanation) into an evolved liquid (glass) composition. This use of an olivine geothermometer assumes that olivine crystallizes fractionally, i.e., olivine is chemically isolated from the melt when it is formed. Along with the assumption of fractional crystallization, the parental liquid calculation also requires the following:

1. A composition of an evolved melt within the olivine-only field. This melt should be either a natural glass composition

or an aphyric whole-rock composition that represents a liquid.

- Establishment of the composition of the most-magnesian olivine phenocryst or microphenocryst for the suite, as a target for the olivine addition calculations. In any magma suite, there will be found a range in olivine phenocryst compositions, if present (Danyushevsky et al., 2002). Most large olivine phenocrysts show normal zoning from core Mg# values ( $Mg\# = 100 \times X_{Mg} / (X_{Mg} + X_{Fe})$ ), where  $X_{Fe}$  and  $X_{Mg}$  are cation fractions of  $Fe^{2+}$  and Mg, respectively) higher than those in equilibrium with the erupted evolved liquid composition. Olivine in equilibrium with the erupted melt is usually present as both discrete microphenocrysts and rims on the more-magnesian phenocryst cores. In some suites, the magnesian olivine phenocrysts are xenocrystic and could either represent (i) disaggregated cumulate material from previously erupted magmas, which may or may not have similar magma compositions to the composition of interest or (ii) lithospheric wallrock samples detached and disaggregated as magmas or melts that have moved through the lithosphere toward crustal magma chambers. Minor element (Ca, Al, Cr, and Ni) or trace element contents of olivine may be used to evaluate phenocryst versus xenocryst relationships with respect to the host magma (Norman and Garcia, 1999). Also in most suites, it is possible to identify and analyze melt inclusions in magnesian olivine phenocrysts that demonstrate that these olivine phenocrysts are related to the evolved, erupted liquid composition, via the process of crystal fractionation. Where phenocryst and microphenocryst relationships are established, it is possible to use an olivine geothermometer to incrementally add back olivine in small equilibrium steps to obtain a parental composition that is in equilibrium with the most-magnesian phenocryst composition observed (Irvine, 1977; Albarede, 1992; Danyushevsky et al., 2000).
- An estimate of volatile content of the evolving melt (especially  $H_2O$ ) and its effect on the crystallization temperature and composition. Anhydrous calculations for magmas that have small amounts of  $H_2O$  can lead to large differences in crystallization temperatures between parental liquid compositions (e.g., Falloon and Danyushevsky, 2000). In this article, we use the model of Falloon and Danyushevsky (2000) to estimate the effect of  $H_2O$ , on olivine liquidus temperatures. The effect of  $H_2O$  on the value of the equilibrium constant for iron-magnesium exchange between olivine and liquid ( $K_D = [(X_{Fe}^{Ol}/X_{Fe}^L)] / [(X_{Mg}^{Ol}/X_{Mg}^L)]$ , where Ol is olivine, L is melt, and  $X_{Fe}$  and  $X_{Mg}$  are cation fractions of  $Fe^{2+}$  and Mg, respectively) is less well known but experimental and theoretical studies (Ulmer, 1989; Putirka, 2005; Toplis, 2005) suggest that the effect is very small for the likely  $H_2O$  contents of tholeiite parental liquids (<1 wt%) and will not introduce a significant error into calculations using olivine geothermometers under anhydrous conditions.
- An estimate of melt oxidation state. Two components are

necessary: (i) an oxygen fugacity and (ii) a model to calculate the  $Fe^{2+}/Fe^{3+}$  ratio of the melt for the given oxygen fugacity (see Danyushevsky and Sobolev, 1996, for a detailed discussion). Differences in oxygen fugacity will not have a serious effect on conclusions reached about temperature differences between parental compositions (~40 °C for four orders of magnitude difference in the oxygen fugacity).

- An estimate of olivine crystallization pressure. This estimate can be determined from studies of primary fluid inclusions in olivine phenocrysts (e.g., Anderson and Brown, 1993; Sobolev and Nikogosian, 1994).
- And finally, the choice of an appropriate olivine geothermometer. In this article, we use the geothermometer of Ford et al. (1983), as it is the most appropriate for modeling olivine crystallization.

A more detailed review of the use of olivine geothermometers in calculating parental liquid compositions is provided by Falloon et al. (2007).

## CASE STUDIES OF HOTSPOT OLIVINE THOLEIITE MAGMAS

### *Hawaii*

**Kilauea Volcano.** Kilauea is an active shield-building volcano on the island of Hawaii. The magmatic rocks of Kilauea volcano have been the subject of a large number of detailed studies. These studies provide us with a relatively comprehensive set of data on whole rocks, glass, and olivine phenocryst compositions, including volatile ( $H_2O$ ) contents. This data makes Kilauea an ideal subject for our approach of calculating parental liquids for an ocean island tholeiite series. There are two aspects of the Kilauean data set that are of particular significance for our purposes. The first is the 1959 summit eruption of Kilauea, as this event was a very intensely studied eruption both in terms of volcanology and petrology. The volcanological observations in particular recorded the height and temperatures of lava fountaining, and these data, combined with associated petrological data on glasses and volatiles, allow us to model the temperatures of eruptive magmas (see below). The second is the recovery of quenched glasses from the submarine Puna Rift (Clague et al., 1991; Clague et al., 1995). These glasses are the most magnesian (~15 wt% MgO) so far recovered for an ocean island tholeiite series. In addition, the Puna Ridge glasses contain as microphenocrysts the most-magnesian olivine compositions (Mg# 90.7) reported from Kilauea. Below we examine the 1959 eruption in detail to gain insights on likely volatile contents and pressures of olivine crystallization for Kilauea volcano. We then use the constraints derived from the 1959 eruption to examine the range of liquid compositions parental to the Puna Ridge glass compositions.

**Kilauea Summit Eruption of 1959.** The 1959 summit eruption lasted from 14 November to 20 December 1959, forming a

cinder cone (Puu Puai) and a large lava lake in Kilauea Iki pit crater. The eruption consisted of seventeen phases of high fountaining and lava output. Of particular importance is that observations were made of lava fountaining heights (maximum observed 580 m) and temperatures (using optical pyrometers; maximum observed fountain temperature 1190 °C; Ault et al., 1961). In addition, the eruptive products of individual fountaining events were sampled. The most-magnesian glass (S-5g; Table 1) for a historical Kilauea eruption was sampled on 18 November 1959 from the hottest lava fountaining and the most-olivine-rich magma of the 1959 eruption (Murata and Richter, 1966). This eruption on 18 November provides us with a unique sample for which we have a liquid composition and its eruption temperature (1190 °C). Using the olivine geothermometer of Ford et al. (1983) in combination with the model of Falloon and Danyushevsky (2000), we calculate a olivine liquidus temperature of 1190 °C, if the glass composition S-5g (also known as Iki-22 in some publications) was a liquid at 0.2 GPa containing 0.93 wt% H<sub>2</sub>O. This calculation was performed at 0.2 GPa, as this is the maximum pressure of olivine crystallization for the 1959 eruption, based on the study of primary CO<sub>2</sub> fluid inclusions by Anderson and Brown (1993). The volatile content is calculated so as to cause the appropriate liquidus depression to

match the observed lava fountaining temperature, using the model of Falloon and Danyushevsky (2000)—in the absence of 0.93 wt% H<sub>2</sub>O, the calculated liquidus temperature for S-5g is 1261 °C i.e., 71 °C above the observed eruption temperature. This model volatile content results in an H<sub>2</sub>O/K<sub>2</sub>O value of 1.94, close to the maximum observed values in olivine hosted melt inclusions from the eruption (Anderson and Brown 1993; see Fig. 1) and values in the submarine Puna Ridge glasses (Clague et al., 1991; Clague et al., 1995). During crystal fractionation, both H<sub>2</sub>O and K<sub>2</sub>O should be highly incompatible and thus should increase together in an evolving magma. The lack of correlation between H<sub>2</sub>O and K<sub>2</sub>O (Fig. 1) indicates that significant degassing of H<sub>2</sub>O has occurred (Wallace and Anderson, 1998). It is therefore appropriate to take the higher H<sub>2</sub>O/K<sub>2</sub>O values as representative of the undegassed magma. Using the calculated H<sub>2</sub>O/K<sub>2</sub>O value for sample S-5g and applying it to the lowest MgO glass from this eruptive episode (sample F-3g, Table 1), we calculate a temperature of 1064 °C at 0.2 GPa, which closely matches the recorded temperature of the coolest fountains (1060 °C; Ault et al., 1961). In summary, we believe that our modeling using the Ford et al. (1983) geothermometer in combination with the model of Falloon and Danyushevsky (2000) can accurately reproduce the eruption temperatures observed during the

TABLE 1. GLASS COMPOSITIONS CHOSEN FOR STUDY

Number	1	2	3	4	5	6	7	8	9	10	11	12
Sample name	F-3g	S-5g	57-11c	57-13g	57-15D	57-9F	182-7	REg	ODP896A	D20-3	Vema	NO42
(wt%)												
SiO <sub>2</sub>	50.12	48.69	48.83	48.7	49.59	51.65	51.24	47.96	49.07	49.15	49.66	48.32
TiO <sub>2</sub>	3.52	2.47	1.92	1.9	2.09	2.27	2.17	2.98	0.64	0.96	0.86	0.78
Al <sub>2</sub> O <sub>3</sub>	13.36	12.39	11.38	10.99	12.03	13.43	13.4	14.2	16.08	16.99	17.77	15.81
Fe <sub>2</sub> O <sub>3</sub>	2.49	1.52	1.6	1.52	1.38	1.04	1.36	1.78	0.67	0.60	0.49	1.06
FeO	10.0	10.02	11.25	10.53	9.98	7.93	9.30	9.3	8.53	7.61	6.61	8.41
MnO	0.18	0.18	N.D.	N.D.	N.D.	N.D.	0.16	0.2	0.12	0.13	0.13	0.18
MgO	5.68	10.12	13.29	14.92	12.53	9.73	8.42	7.09	9.44	9.88	9.51	9.80
CaO	9.61	10.95	9.01	8.67	9.46	10.63	10.7	11.3	13.67	12.13	12.45	13.79
Na <sub>2</sub> O	2.57	2.03	1.65	1.63	1.77	2.03	2.16	2.62	1.62	2.41	2.32	1.63
K <sub>2</sub> O	0.71	0.48	0.3	0.32	0.33	0.36	0.31	0.76	0.02	0.02	0.06	0.04
P <sub>2</sub> O <sub>5</sub>	0.39	0.24	0.20	0.19	0.21	0.23	0.19	0.29	0.03	0.05	0.05	0.09
Cr <sub>2</sub> O <sub>3</sub>	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	0.07	N.D.	N.D.	N.D.
H <sub>2</sub> O	1.38	0.93	0.58	0.63	0.64	0.7	0.6	1.52	0.05	0.07	0.10	0.10
Mg#	50.3	64.3	67.8	71.6	69.1	68.6	61.7	57.6	66.4	69.8	71.9	67.5
Temp*	1064	1190	1269	1301	1253	1189	1157	1100	1200	1223	1207	1202
Oliv Eq†	74.8	85.0	86.6	88.7	87.4	87.1	82.9	81.1	86.2	88.3	89.3	86.9

Data sources: Glasses 1 and 2, Murata and Richter (1966); glasses 3–6, Clague et al. (1995); glass 7, Garcia et al. (1985); glass 8, Tilley et al. (1971); glass 9, glass ODP Leg 896A-27R-1, 124-130, pc.15, McNeill and Danyushevsky, 1996; glass 10, Danyushevsky et al. (2003); glass 11, Sobolev et al. (1989); glass 12, Sigurdsson et al. (2000).

Notes: Fe<sub>2</sub>O<sub>3</sub> and FeO contents of all glasses have been calculated at QFM + 0.5 log units (glasses 1–8, 12) or QFM – 0.5 log units (glasses 9–11) at 0.2 GPa using the model of Borisov and Shapkin (1990). H<sub>2</sub>O contents are either analysed in the glass (glasses 9–11) or calculated from K<sub>2</sub>O contents (glasses 1–8, 12; see text for discussion). All calculations performed with software PETROLOG (Danyushevsky, 2001). N.D.—no data.

\*Calculated olivine liquidus temperature (°C) at 0.2 GPa using the models of Ford et al. (1983) and Falloon and Danyushevsky (2000).

†Calculated equilibrium olivine composition at the 0.2-GPa liquidus temperature.

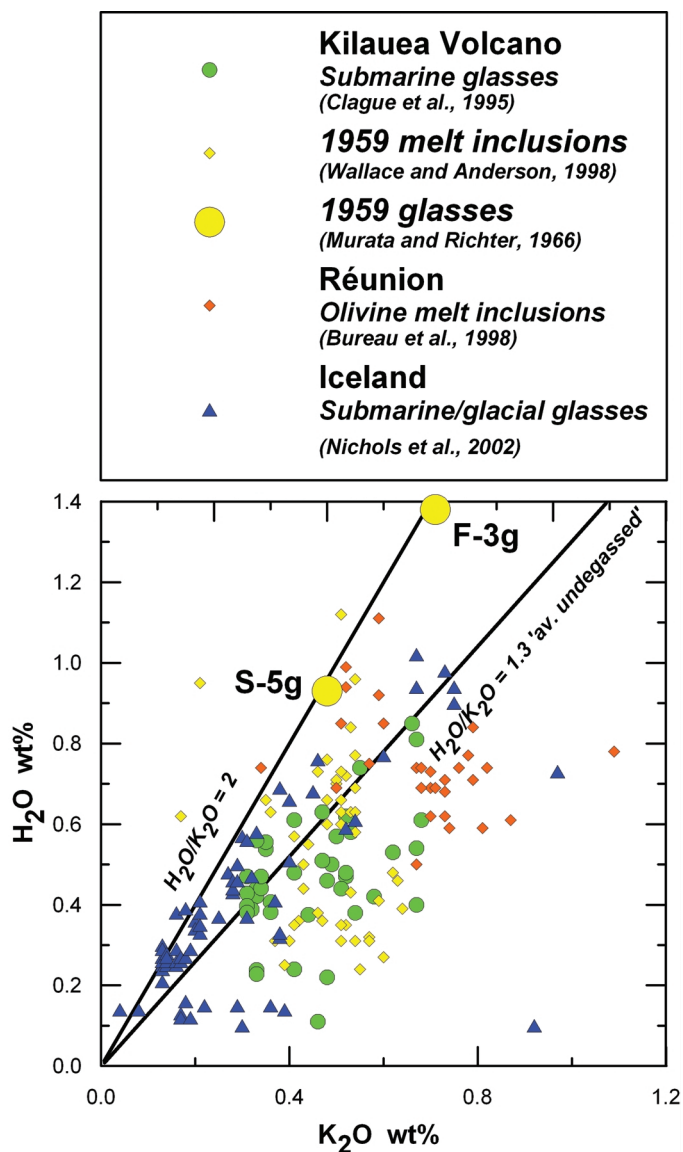


Figure 1.  $\text{H}_2\text{O}$  versus  $\text{K}_2\text{O}$  contents of glasses and melt inclusions from ocean island basalt “hotspot” tholeiite series magmas. The  $\text{H}_2\text{O}$  contents of the 1959 glasses are calculated using the observed eruption temperatures as a constraint (see text for discussion). The line “av. undegassed” is from the study of Wallace and Anderson (1998).

1959–1960 eruptions of Kilauea volcano. We therefore use a pressure of 0.2 GPa and a  $\text{H}_2\text{O}/\text{K}_2\text{O}$  value of 1.94 for the purposes of calculating parental liquid compositions to Kilauea volcano (see below).

**Puna Ridge Glasses.** In Figure 2, the  $\text{FeO}^{\text{T}}$  and  $\text{MgO}$  contents of glasses recovered from the Puna Ridge are compared with glasses and whole-rock compositions with  $>8.5$  wt%  $\text{MgO}$  from Kilauea volcano. The blue arrow in Figure 2 delineates an olivine control line defined by the whole-rock compositions of the 1959 summit eruption. This control line defines mixtures of

variably fractionated liquids and olivine crystals with an average composition of  $\sim\text{Mg}\# 87$  (Murata and Richter, 1966). The whole-rock composition (S-5) for the glass S-5g was the most-olivine-rich magma of the 1959 eruption (Fig. 2). As can be seen from Figure 2, the Puna Ridge glasses display a significant range in  $\text{FeO}^{\text{T}}$  and  $\text{MgO}$  contents compared to the whole-rock data analyzed from Kilauea volcano. Four of the Puna Ridge glasses are of particular interest (Table 1). Sample 57–11c is the most-mag-

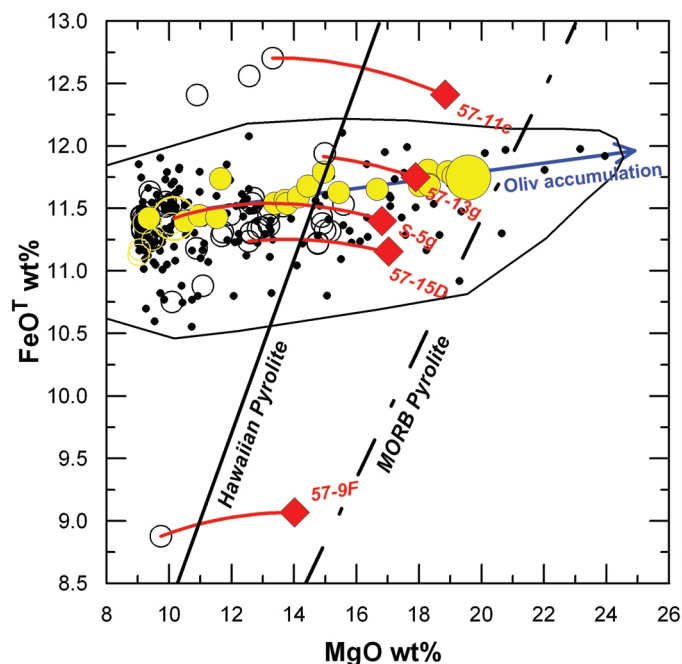


Figure 2.  $\text{FeO}^{\text{T}}$  versus  $\text{MgO}$  contents of whole rocks and glasses  $>8.5$  wt%  $\text{MgO}$  from Kilauea volcano, Hawaii. Symbols and lines: small yellow circles—whole-rock compositions from the 1959 Kilauea summit eruption; small open yellow circles—glass compositions from the 1959 Kilauea summit eruption; large open yellow circle—glass S-5g; large yellow circle—whole-rock composition S-5; blue arrow—an olivine control line representing olivine accumulation of  $\sim\text{Mg}\# 87$ ; small black circles—other Kilauean whole-rock compositions; open black circles—other Kilauean glass compositions; red diamonds—calculated 0.2-GPa parental liquid compositions (see Table 2); red lines—olivine addition paths (reverse of fractional crystallization); solid black line—locus of solidus melts for Hawaiian pyrolite composition (Green and Falloon, 1998); dash-dot black line—locus of solidus melts for mid-ocean ridge basalt (MORB) pyrolite composition (Green and Falloon, 1998). Both lines are calculated using the equations of Herzberg and O’Hara (2002). Data sources: 1959 glasses, Murata and Richter (1966), Helz (1987); 1959 whole rocks, Murata and Richter (1966), Basaltic Volcanism Study Project (1981); other Kilauea glasses, Moore (1965), Clague et al. (1991, 1995), Garcia et al. (1996, 2003); other Kilauea whole rocks, MacDonald and Eaton (1955), Tilley (1960), Muir and Tilley (1963), Moore (1965), Murata and Richter (1966), Richter and Moore (1966), Moore and Evans (1967), Aramaki and Moore (1969), Moore and Koyanagi (1969), Gunn (1971), Wright (1971), Wright and Fiske (1971), Helz (1980), Basaltic Volcanism Study Project (1981), Wilkinson and Hensel (1988), Garcia et al. (1992, 1998, 2000, 2003); Clague et al. (1995), Mangan et al. (1995), Chen et al. (1996), Norman and Garcia (1999), Quane et al. (2000).

nesian glass of a high-FeO<sup>T</sup> subset of glasses (Fig. 2), whereas sample 57-9F is the only example of a relatively low-FeO<sup>T</sup> glass. As can be seen from Figure 2, both the high- and low-FeO<sup>T</sup> glasses have no matching whole-rock compositions. That is, so far, no magma composition has been analyzed from Kilauea volcano that is consistent with olivine accumulation into a high or low-FeO<sup>T</sup> liquid composition. Glass samples 57-13g and 57-15D are important, as they both contain olivine microphenocrysts (<20 µm) with core compositions ranging to Mg# 90.7, the most magnesian observed from Kilauea volcano. As the olivines are small microphenocrysts, we can be reasonably confident that these microphenocryst have crystallized from a parental liquid related to the evolved glass composition. As well, both samples 57-13g and 57-15D have FeO<sup>T</sup> contents within the array defined by whole-rock compositions from Kilauea volcano (Fig. 2).

**Calculation of Parental Melts for Kilauea Volcano.** In Table 2 and Figure 2, we present the results of parental liquid compositions at 0.2 GPa using a target olivine composition of Mg# 90.7. We have performed calculations for S-5g and the four Puna Ridge glasses. Note that in Figure 2, the liquid lines of descent produced by adding olivine incrementally back into the glasses produce curved as opposed to straight lines. This observation emphasizes the point made by Irvine (1977) and Albaredo

(1992) that extrapolation of olivine control lines (due to olivine accumulation) will result in incorrect calculated parental liquid compositions. Calculated parental compositions range in MgO contents from 14 to 18.8 wt% with crystallization temperatures ranging from 1286 to 1372 °C. As we do not know whether the target olivine is appropriate for both the high-FeO<sup>T</sup> and low-FeO<sup>T</sup> glasses, the best-constrained parental compositions are for glasses 57-13g and 57-15D. The parental liquids for these two compositions have 17-18 wt% MgO and crystallization temperatures of 1340-1355 °C.

**Mauna Loa Volcano.** The parental composition for Mauna Loa volcano was calculated in a similar manner to the Kilauea parental compositions. We used the most-MgO-rich glass recovered from the submarine southwest rift of Mauna Loa volcano (glass 182-7, Table 1). Sample 182-7 is particularly significant, as it contains the most-magnesian olivine phenocryst composition (Mg# 91.3) obtained from Mauna Loa volcano. Using this target olivine, we calculate a parental liquid composition containing 17.53 wt% MgO at a temperature of 1354 °C (Table 2). As can be seen from Table 2, the parental liquid and temperature is essentially identical to those for samples 57-13g and 57-15D from Kilauea volcano. We believe that this result further supports the conclusion that samples 57-13g and 57-15D provide the most reliable estimates for Kilauea parental liquid compositions.

TABLE 2. CALCULATED MODEL PARENTAL LIQUIDS AT 0.2 GPa

Number	Kilauea					Mauna	Réunion	MORB			Iceland
	1	2	3	4	5	Loa	7	8	9	10	11
Sample name	S-5g	57-11c	57-13g	57-15D	57-9F	182-7	REg	ODP896A	D20-3	Vema	NO42
(wt%)											
SiO <sub>2</sub>	47.24	47.53	48.01	48.48	50.44	48.74	46.23	47.70	48.29	49.36	47.00
TiO <sub>2</sub>	2.03	1.62	1.73	1.83	2.02	1.67	2.31	0.54	0.86	0.83	0.64
Al <sub>2</sub> O <sub>3</sub>	10.21	9.59	10.04	10.54	11.95	10.30	10.98	13.46	15.25	17.17	13.03
Fe <sub>2</sub> O <sub>3</sub>	1.58	1.80	1.64	1.51	1.12	1.54	1.51	0.85	0.69	0.51	1.40
FeO	9.96	10.77	10.26	9.78	8.05	9.38	9.91	8.56	7.69	6.68	8.15
MnO	0.15	N.D.	N.D.	N.D.	N.D.	0.12	0.15	0.10	0.12	0.12	0.15
MgO	16.79	18.80	17.86	17.00	14.00	17.53	16.17	15.86	13.93	10.86	16.75
CaO	9.03	7.59	7.92	8.28	9.46	8.22	8.73	11.44	10.88	12.03	11.37
Na <sub>2</sub> O	1.67	1.39	1.49	1.55	1.81	1.66	2.02	1.36	2.16	2.24	1.34
K <sub>2</sub> O	0.39	0.25	0.29	0.29	0.32	0.24	0.59	0.02	0.02	0.06	0.03
P <sub>2</sub> O <sub>5</sub>	0.20	0.17	0.18	0.18	0.20	0.15	0.22	0.03	0.04	0.05	0.07
Cr <sub>2</sub> O <sub>3</sub>	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	0.06	N.D.	N.D.	N.D.
H <sub>2</sub> O	0.76	0.49	0.57	0.56	0.62	0.46	1.18	0.04	0.06	0.10	0.08
Mg#	75.0	75.7	75.6	75.6	75.6	76.9	74.4	76.8	76.3	74.3	78.6
Temp*	1335	1372	1355	1341	1286	1354	1323	1351	1320	1243	1361
Oliv Eq†	90.70	90.70	90.70	90.70	90.70	91.30	90.65	91.60	91.50	90.50	92.40
% Oliv‡	17.6	15.7	8.6	12.4	11.0	23.1	22.7	16.3	10.3	3.4	17.5

Notes: Data sources as for Table 1. All calculations performed with software PETROLOG (Danyushevsky, 2001). N.D.—no data; MORB—mid-ocean ridge basalt.

\*Calculated olivine liquidus temperature at 0.2 GPa using the models of Ford et al. (1983) and Falloon and Danyushevsky (2000).

†Calculated equilibrium olivine composition at the 0.2-GPa liquidus temperature.

‡Amount of olivine added in incremental steps to obtain equilibrium with the target olivine compositions (see text for discussion).

### ***Réunion: The 1939 Eruption of Piton de la Fournaise***

The island of Réunion consists of two shield volcanoes, Piton des Neiges (3069 m) and Piton de la Fournaise (2631 m). The former is extinct and deeply eroded, whereas the latter is an active volcano. Picrites and olivine basalts from both volcanoes are referred to as the oceanite series. To model the composition of a parental magma for the oceanite series lavas of Réunion, we have chosen the melt composition from the oceanite magma eruption of 6 January 1939, from Piton de la Fournaise volcano. The whole-rock composition of the 1939 magma had 15.21 wt% MgO, and the melt component of the magma, represented by glass REg, has 7.18 wt% MgO (Table 1). The target olivine for our calculations is the most-magnesian olivine phenocryst composition reported from the oceanite lavas of Réunion of Mg# 90.65 (Fretzdorff and Haase, 2002). The H<sub>2</sub>O contents of olivine hosted melt inclusions in oceanite series lavas from Réunion suggest parental liquids have H<sub>2</sub>O/K<sub>2</sub>O values of ~2 (Bureau et al., 1998; Fig. 1), essentially identical to Kilauea liquids. The calculated parental liquid to the REg glass composition at 0.2 GPa, has ~16 wt% MgO and a crystallization temperature of 1323 °C (Table 2), within the range calculated for parental liquids to Kilauea.

### ***Iceland: Theistareykir Volcanic System in Northeastern Iceland***

The most-magnesian olivine phenocryst composition that has been reported from Iceland has Mg# of 92.4 (Sigurdsson et al., 2000). This olivine phenocryst comes from the Borgarhraun lava flow, which is part of the Theistareykir volcanic system and has an average whole-rock MgO content of 12 wt% (MacLennan et al., 2003). The liquid composition is represented by glass NO42, which has 9.80 wt% MgO (Sigurdsson et al., 2000). The water content in this glass has not been analyzed but can be estimated from its K<sub>2</sub>O content. Nichols et al. (2002) analyzed H<sub>2</sub>O contents of submarine and subglacial pillow basalts and concluded the Icelandic mantle source is relatively enriched in water. As can be seen from Figure 1, water contents in Icelandic lavas are very similar to Kilauea and Réunion. For the purposes of our calculation of a liquid composition parental to the glass NO42, we have used a H<sub>2</sub>O/K<sub>2</sub>O value of 2.5, which is close to the undegassed value for Icelandic magmas. The calculated parental liquid (Table 2) has 16.75 wt% MgO and a crystallization temperature of 1361 °C at 0.2 GPa.

### **COMPARISON BETWEEN “HOTSPOT” AND SPREADING-RIDGE THOLEIITES**

The thermally driven plume hypothesis predicts that there should exist a significant thermal anomaly between upwelling plume material and the temperature of ambient upper mantle. There is a general consensus that MORBs are the result of melting from upwelling ambient upper mantle at oceanic spreading

ridges. It would therefore seem a straightforward matter to compare parental liquid olivine crystallization temperatures between OIB and MORB. It is common in the literature to find the temperature of 1280 °C quoted and used as a sort of bench mark for T<sub>p</sub> of the upwelling ambient upper mantle after the study of McKenzie and Bickle (1988). However, it is important to recognize that this mantle T<sub>p</sub> of 1280 °C represents an average, not the hottest, MORB mantle T<sub>p</sub>. Using this average temperature, it would appear obvious that a thermal anomaly exists. Indeed our own calculations for Kilauea volcano (see Table 5 below) would indicate a thermal excess of ~285 °C, compared to an ambient upper mantle T<sub>p</sub> of 1280 °C, a result consistent with the thermal plume hypothesis. We, however, strongly disagree with the use of an average temperature for MORB when there is clearly a range in olivine crystallization temperatures for MORB glasses. This spread in values is demonstrated by the range in FeO<sup>T</sup> versus MgO contents of MORB glasses (Fig. 3). In Figure 3 we have plotted the FeO<sup>T</sup> and MgO contents of 682 MORB glass analyses from the study of Danyushevsky (2001). These glass compositions were all determined by electron microprobe analysis using the glass standard VG-2 (U.S. National Museum no. 111240/2; Jarosewich et al., 1980). Also plotted in Figure 3 are FeO<sup>T</sup> and MgO contents of 189 glass compositions from the Petdb database (petdb.ldeo.columbia.edu). These 189 glasses were all determined by electron microprobe analysis (we note that many reported glasses in the Petdb database are in fact glass-rich rocks and whole-rock analyses) and have MgO contents >9.5 wt%. Consequently all these glasses (liquids) are in the olivine-only phase field at low pressure. As can be seen from Figure 3, the MORB glasses from the two independent data sets display a significant range in FeO<sup>T</sup> contents, and consequently there must exist a range in temperatures of MORB parental liquids. This temperature range must occur, because if all calculation parameters are held constant, higher FeO<sup>T</sup> contents (of liquids of MORB-like chemical composition) require higher temperatures to be in equilibrium with the same olivine composition.

In this article, we wish to compare temperatures between the hottest of MORB parental liquids and the hottest of the OIB parental liquids, not averages. We believe it is appropriate to take the hottest of MORB parental liquids as representative of melts derived from adiabatic upwelling of ambient upper mantle. This decision is reasonable because there are several processes—e.g., nonadiabatic upwelling, cooling, mantle-melt reaction—that could conceivably lead to lower temperatures in MORB parental liquids. If the lowest temperatures are taken as representative of ambient upper mantle, this choice provides no explanation for why hotter MORBs exist, apart from the circular argument that they are plume related and represent a depleted, MORB-source-like component within the plume. The discussion illustrates the difficulty of testing the deep-mantle thermal plume hypothesis. In its most clearly stated form, the hypothesis asserts that spreading centers sample modern well-mixed upper mantle with normal MORB (N-MORB) trace element sig-

natures. Mantle plumes sample compositionally distinct (OIB trace element and isotopic characteristics), heterogeneous mantle sources residing in the deep mantle. To the extent that these distinctions between magma and source characteristics are abandoned, the hypothesis requires modification or rejection. Just as circularity in argument is introduced if MORB-type (geochemically) and picritic magmas indicating high temperature are said to be influenced by (or be part of) a nearby plume, so circularity of argument is introduced by blurring of geochemical distinctions. The enriched, OIB-like geochemical signatures at many spreading centers with no obvious nearby plume are often attributed to a plume-component but without any justification. For these reasons, we have sought to establish

magma temperatures and model mantle potential temperatures for the classical, geodynamically defined “fixed” hotspots and on geodynamically defined spreading centers, and the latter samples chosen for detailed scrutiny have N-MORB or depleted MORB (D-MORB) geochemistry. Anticipating our possible conclusion, if we can find no significant temperature difference between parental MORB and hotspot magmas, we seriously undermine the deep-mantle thermal plume model but are left with evidence for heterogeneous mantle, geochemical differences between MORB and OIB sources, and the hotspot source and/or cause below and detached from the moving lithospheric plate.

To compare crystallization temperatures with MORB, we have calculated parental liquid compositions at 0.2 GPa for three representative MORB glasses with differing  $\text{FeO}^T$  contents (Table 1, sample numbers 9–11). These glasses are:

1. The most-magnesian glass (Table 1, number 9) from Ocean Drilling Program (ODP) site 896A, drilled into ca. 5-Ma crust formed by the Cocos-Nazca spreading ridge (herein referred to as ODP896A; McNeill and Danyushevsky, 1996). This glass contains microphenocrysts with core compositions ranging to Mg# 91.6.
2. Glass D20–3 (Table 1, number 10) from the Siqueiros Fracture Zone, East Pacific Rise (Danyushevsky et al., 2003). The most-magnesian olivine phenocryst observed in the Siqueiros Fracture Zone has Mg# of 91.5.
3. The composition (Table 1, glass 11) of a magnesian olivine-hosted glass inclusion from the Vema Fracture Zone on the

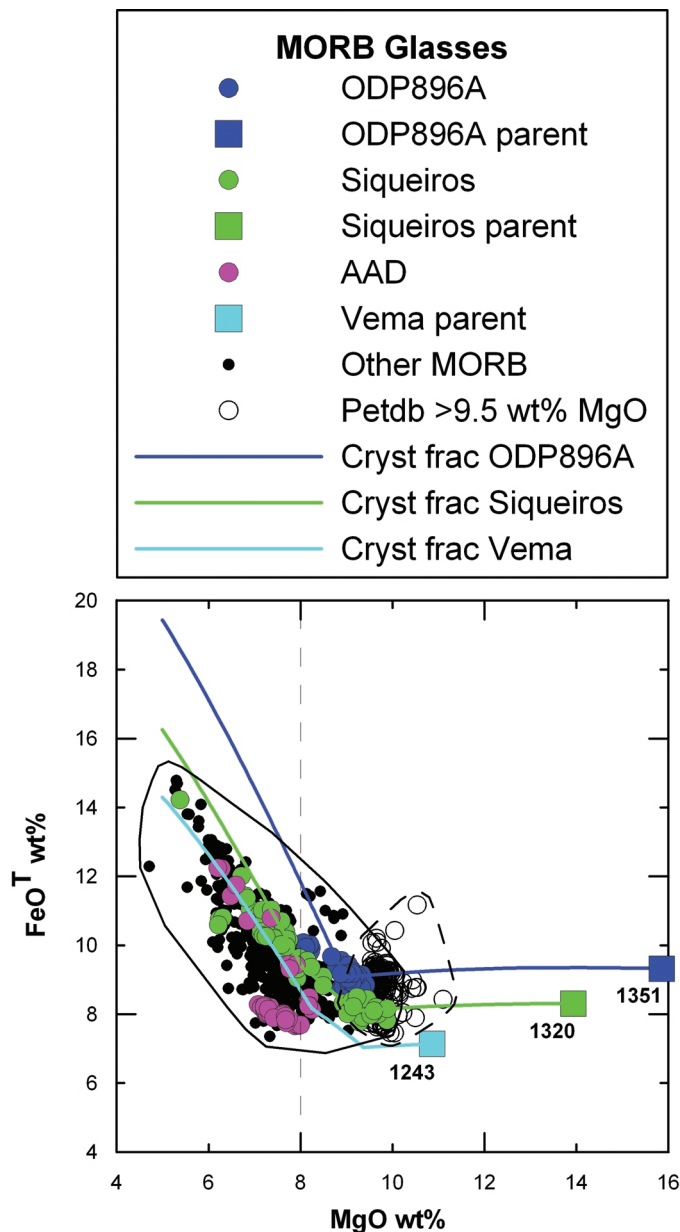


Figure 3.  $\text{FeO}^T$  versus MgO wt% of mid-ocean ridge basalt (MORB) glass compositions. Symbols, lines and fields: Blue square—parental liquid to MORB glass ODP896A (Table 2, number 8); green square—parental liquid to MORB glass D20–3 (Table 2, number 9); cyan square—parental liquid to MORB glass Vema, mid-Atlantic Ridge (Table 2, number 10); blue circles—glass compositions from hole ODP896A, Cocos-Nazca (Danyushevsky, 2001); green circles—glass compositions from Siqueiros Fracture Zone, East Pacific Rise (Danyushevsky et al., 2003); magenta circles—glass compositions from the Australian-Antarctic discordance (AAD; Danyushevsky, 2001); small black circles—MORB glasses from the study of Danyushevsky (2001); small open black circles—glasses from the Petdb database with >9.5 wt% MgO, as determined by electron microprobe analysis (petdb.ldeo.columbia.edu); solid blue, green and cyan lines—calculated liquid lines of descent for each respective parental liquid compositions undergoing crystal fractionation of olivine, clinopyroxene, and plagioclase at 0.2 GPa. Calculations were performed using the software PETROLOG and the methods of Danyushevsky (2001). MORB glass field delineated by solid black line encompasses the range of  $\text{FeO}^T$  and MgO contents for 682 glasses from Danyushevsky (2001). MORB glass field delineated by dashed black line encompasses the range of  $\text{FeO}^T$  and MgO contents of 189 glasses with >9.5 wt% MgO, as determined by electron microprobe analysis from the Petdb database (petdb.ldeo.columbia.edu). Temperatures of MORB parental liquid compositions are calculated using the olivine geothermometer of Ford et al. (1983) at 0.2 GPa (see text and Table 2 for details). Dashed line at 8 wt% MgO is the MORB reference MgO content after Klein and Langmuir (1987).



mid-Atlantic Ridge from the study of Sobolev et al. (1989). This glass inclusion composition is important, as it is representative of a relatively low-FeO<sup>T</sup> liquid compared with the other parental MORB liquid compositions. The maximum olivine phenocryst composition observed in the Vema Fracture Zone has Mg# of 90.5.

The H<sub>2</sub>O contents listed in Table 1 have all been determined on the glasses themselves. In Table 3 and Figure 3 (see also Fig. 5 below), we present the results of our parental liquid calculations for all three compositions. The range in MgO contents in parental liquids varies from from ~11 to 16 wt% MgO, and crystallization temperatures at 0.2 GPa range from 1243 to 1351 °C.

A significant result of our calculations (Table 2) is that the olivine crystallization temperature of the representative ODP896A parental liquid is significantly hotter than the commonly quoted ambient upper-mantle T<sub>p</sub> value of 1280 °C and is higher than the values inferred for MORB by Presnall et al. (2002), based on experimental studies on simple systems (average mantle potential temperature required ~1260 °C). We have chosen this particular high-FeO<sup>T</sup> MORB glass from ODP896A simply because we have detailed information on both its mineralogy and geochemistry. Such detailed information, required for our calculations, is currently not available from the vast majority of studies that report MORB glass geochemistry. The ODP896A glass has normal N-MORB geochemistry (L.V. Danyushevsky, unpubl. data, 2006), and we believe it is representative of other high-FeO<sup>T</sup> MORB glasses

for which we do not have detailed information on mineralogy. Over half the glasses in the Petdb database ( $n = 80$ ) with MgO contents >9.5 wt% have FeO<sup>T</sup> > 9 wt%, similar to the ODP896A glass. The average FeO<sup>T</sup> of these high FeO<sup>T</sup> (>9 wt%) glasses is  $9.3 \pm 0.3$ , thus the ODP896A glass (FeO<sup>T</sup> = 9.1) has slightly below-average FeO<sup>T</sup> compared to these glasses. These high-FeO<sup>T</sup> glasses are sampled from a range of “normal” crustal thickness spreading ridges (Carslberg Ridge, Chile Ridge, Easter Microplate Rifts, East Pacific Rise, Galapagos Spreading Center, Juan De Fuca Ridge, mid-Atlantic Ridge, and Red Sea) and are associated with low-FeO<sup>T</sup> glasses both in time and space. Although primitive high-MgO and high-FeO<sup>T</sup> glasses, such as ODP896A, are rarely sampled at spreading centers, glass compositions derived from such parents are relatively common. Dmitriev et al. (1985), using an eight-component discriminant function, determined that at least 37% of MORB glasses from the mid-Atlantic Ridge were derived from such high-FeO<sup>T</sup> parental compositions.

If we use our approach and calculate parental compositions for all the high-MgO glasses (>9.5 wt%;  $n = 189$ ) from the Petdb database, using K<sub>2</sub>O abundances as an estimate for H<sub>2</sub>O contents and an olivine target composition of Mg# 91.5, then calculated parental liquids at 0.2 GPa display a significant range in both MgO contents (13–18 wt%) and temperatures (1300–1430 °C); however, the averages ( $15.1 \pm 0.9$  wt% MgO;  $1343 \pm 19$  °C) are very close to the calculated values for the ODP896A parental composition for which we have detailed petrological information. To determine the true range in composition and crystallization temperatures of MORB parental liquids requires more detailed petrological and mineralogical studies from a range of different spreading ridges. The uncertainty in calculations for which only electron probe analyses of glass compositions are available is that we do not know the actual H<sub>2</sub>O contents, nor do we have any idea of the maximum Mg# of olivine phenocrysts. It may well be that we are underestimating the temperatures of some of the high-FeO<sup>T</sup> glasses, as some picrite suites with similar MORB geochemistry have olivines up to Mg# 93–94 (e.g., Baffin and Gorgona Island picrites). It is a circular argument that assigns these high temperature picrite suites to the influence of mantle plumes: they may well be derived from normal ambient upper mantle.

In summary, therefore, we believe that when comparisons are made between olivine crystallization temperatures of parental liquids for OIB and MORB magmas, it is the temperatures inferred for parental liquids for high-FeO<sup>T</sup> MORB glass suites that should be used to test the thermal plume hypothesis. The result of our calculations suggest there are no significant differences in the olivine crystallization temperatures when OIB and MORB parental liquids are compared (the differences between MORB and the best-constrained calculations for Kilauea volcano are essentially zero, and the Icelandic parent is only 10 °C hotter). If other picrite suites with MORB geochemistry are used in this comparison, then OIB parental liquids would be significantly cooler than MORB parental liquids.

**TABLE 3. COMPARISON OF MODEL KILAUEA PARENTAL LIQUIDS WITH EXPERIMENTAL MELT COMPOSITIONS FROM HW-40 AT 2 GPa UNDER ANHYDROUS CONDITIONS**

Glass name	T-2018	57-15D	57-13g	T-650
(wt%)				
SiO <sub>2</sub>	48.15	48.48	48.01	50.19
TiO <sub>2</sub>	2.35	1.83	1.73	1.74
Al <sub>2</sub> O <sub>3</sub>	10.76	10.54	10.04	8.61
Fe <sub>2</sub> O <sub>3</sub>	N.A.	1.51	1.64	N.A.
FeO	10.16	9.78	10.26	9.34
MgO	16.27	17.00	17.86	19.96
CaO	9.38	8.28	7.92	7.40
Na <sub>2</sub> O	1.87	1.55	1.49	1.40
K <sub>2</sub> O	0.39	0.29	0.29	0.32
P <sub>2</sub> O <sub>5</sub>	N.D.	0.18	0.18	N.D.
H <sub>2</sub> O	N.D.	0.56	0.57	N.D.
Temp*	1500	1506	1523	1550
Mg#	74.1	75.6	75.6	79.2

Data sources: experimental glasses T-2018 and T-650, Falloon et al. (1988).

Note: N.D.—no data; N.A.—not applicable.

\*The anhydrous olivine liquidus temperatures calculated at 2 GPa using the model of Ford et al. (1983) and assuming all Fe as Fe<sup>2+</sup>.

## MANTLE POTENTIAL TEMPERATURES

As mentioned previously, the calculation of parental liquid compositions is only the first step in the process of calculating a mantle  $T_p$ . To calculate mantle  $T_p$  values, it is necessary to make a number of assumptions concerning source compositions, mantle melting, and the melt segregation process. It is beyond the scope of this article to present any comprehensive argument or defense of any particular mantle  $T_p$ . What we do in this section is present some simple mantle  $T_p$  calculations based on the following assumptions:

1. For comparative purposes, we first assume that the mantle sources for both OIB and MORB primary melts are of pyrolite (lherzolite) composition, with bulk CaO and  $Al_2O_3$  contents within the range of ~3–4 wt%. We believe the assumption of a peridotitic (pyrolite) source is justified by the magnesian contents of both the calculated parental liquids and co-existing olivines (Table 3). The production of primary melts of MgO contents of between 10 and 19 wt% MgO, and olivines of Mg# 90.5–92.4 requires an olivine-rich peridotite source. Many studies of Hawaiian volcanism have argued the need for recycled crustal components in the mantle source in the form of eclogite and/or pyroxenite (Takahashi and Nakajima, 2002; Sobolev et al., 2005). However, numerous experimental melting studies have demonstrated that eclogite and pyroxenite source components are incapable of producing any of the calculated parental liquids unless there has either been extensive reaction with peridotite before melt segregation (Green and Falloon, 2005) or, alternatively, mixing with a dominant peridotite melt component before eruption (Sobolev et al., 2005). In Figure 2, we have plotted the locus of  $FeO^T$  and MgO contents of initial melts (i.e., at the solidus) for both Hawaiian and MORB pyrolite compositions (Green and Falloon, 1998), using the equations of Herzberg and O'Hara (2002). Melts formed above the solidus by higher degrees of melting lie to the right of these lines. As can be seen from Figure 2, it is impossible for a MORB pyrolite composition source to produce melt compositions with appropriate  $FeO^T$  and MgO contents that match the calculated parental liquid compositions to Kilauea volcano. This observation was used as an argument by Herzberg and O'Hara (2002) for more-magnesian parental liquids for Hawaii. For a mantle composition, such as MORB pyrolite (7.55 wt%  $FeO^T$ ), to be a source for Kilauea parental liquids, MgO contents, in the case of the 57–13 g parental liquid, would need to be >20 wt% MgO with olivine of >Mg# 91.9 at 0.2 GPa. So far there is no evidence for such magnesian olivine compositions from Hawaiian volcanoes. However, if the mantle source has  $FeO^T$  and MgO contents similar to Hawaiian pyrolite composition (9.5 wt%  $FeO^T$ ; Green and Falloon, 1998), then the calculated parental liquid compositions are possible primary melts of this mantle composition.
2. We assume that the melting process can be closely modeled by simple batch melting, and that the compositions of unmodified primary melts segregate from their mantle sources as a fully integrated and equilibrated melt compositions. Thus, based on this assumption, we should be able to find at some  $P$  and  $T$  a close match in composition between our calculated parental liquids and experimental melts from our assumed mantle source if the parental liquids are indeed primary mantle derived melts. Justification for this assumption is presented in Green and Falloon (1998), and in the case of Kilauea, is supported by the petrogenetic model presented by Eggins (1992a,b). Our assumption of batch melting means that we can simply calculate the degree of partial melting of our mantle source composition by using the abundances of the relatively incompatible minor elements  $TiO_2$ ,  $Na_2O$ , and  $K_2O$  in our calculated parental liquids.
3. We ignore the possible effect of the relatively minor amounts of  $H_2O$  and  $CO_2$  that are contained in our calculated parental compositions, on the positions of mantle melting cotectics within the normative tetrahedron. We are not focusing on the C-H-O-controlled solidus and incipient melting regime but on the onset of the major melting regime (Green and Falloon, 1998, 2005). The main effect of  $H_2O$  is to produce a higher degree of partial melt at a given temperature compared to volatile-free conditions. However, when melt compositions are compared on an anhy-

The Hawaiian pyrolite composition is believed to be a result of a complex history of previous melt extraction followed by mantle refertilization by small-degree melts from recycled subducted oceanic crust interacting with refractory lithosphere (harzburgite) and normal mantle (Yaxley and Green, 1998; Green and Falloon, 2005). We have previously suggested that, rather than assuming that the source composition (Hawaiian pyrolite) should resemble the most-fertile mantle-derived lherzolites with 3–4 wt% CaO and  $Al_2O_3$  (Green and Falloon, 1998), and thus yielding parental tholeiitic picrites by 30–40% melting (see Table 5), a more appropriate source is a refertilized harzburgite with ~1.5–2.0 wt% CaO and  $Al_2O_3$  (Green et al., 2001). Such a harzburgitic source would be compositionally buoyant if it is at a temperature similar to the enclosing MORB pyrolite, and on upwelling would yield ~10% tholeiitic picrite with harzburgite residue. We therefore make alternative assumptions in our mantle  $T_p$  calculations that either a very fertile lherzolite (Hawaiian pyrolite) or a refertilized harzburgite is a suitable mantle source composition for OIB primary melts. The two assumptions demonstrate the importance of the melt fraction and latent heat of melting estimates to calculation of mantle potential temperature. For MORB, we assume a composition similar to peridotite MM-3 (Baker and Stolper, 1994; see Fig. 5 below) and MORB pyrolite (Green et al., 1979) as suitable mantle source compositions—both of these are lherzolite compositions with 3–4 wt% CaO and  $Al_2O_3$ .

drous basis, no significant differences are apparent either in the compositions of experimental melts or in their normative positions (for H<sub>2</sub>O approximately <1% in the melt).

- We assume that the mantle source upwells adiabatically and that when it crosses, its solidus undergoes partial melting. This assumption is very important, as mantle T<sub>p</sub> calculations presented will be in error if upwelling is not adiabatic (c.f., Ganguly, 2005). For these models of tholeiitic picrite genesis, we assume that the “solidus” is actually the entry into the major melting regime of the peridotite-C,H,O system (see assumption 3 above; Green and Falloon, 1998, 2005; Fig. 3).
- To calculate the amount of heat loss caused by melting of the mantle source, we have used the values for the latent heating of melting (100 cal/g) and heat capacity (0.3 cal/g/deg) recommended by Langmuir et al. (1992). For the two alternative source compositions (lherzolite and harzburgite) for Hawaiian or other hotspot magmas, the heat loss calculated is very different, such that ΔT<sub>p</sub> for different melt fractions is ~3.3 °C per 1% melt (see Table 5).
- We assume a mantle solid adiabat of 10 °C/GPa (Birch, 1952) and liquid adiabat of 30 °C/GPa (McKenzie and Bickle, 1988).

### Depths of Mantle Equilibration

To determine mantle depths of mantle equilibration of our calculated parental liquids, we have projected their compositions from both olivine and diopside end-members into the normative tetrahedron and compared them with experimentally determined mantle melting cotectics (Figs. 4 and 5; Tables 3 and 4). In Table 5, we also present a pressure of mantle equilibration using the experimentally calibrated empirical model of Albarède (1992). Both methods give reasonably consistent results.

**Kilauea Volcano.** In Figure 4, the glasses and parental compositions from Kilauea are shown projected from olivine (Fig. 4A) or diopside (Fig. 4B) and compared with experimentally determined cotectics on the Hawaiian pyrolite composition. Green and Falloon (2005) have previously shown that compositions from a range of Hawaiian volcanoes define arrays of compositions in the olivine projection consistent with equilibrium with harzburgite residues but with differing Ca/Al ratios. In Figure 4A, it can be seen that the mantle source for the 1959 Kilauea summit eruption must have a higher Ca/Al value compared to the source for the Puna Ridge glasses and other historical eruptions of Kilauea (e.g., data of Norman and Garcia, 1999; Norman et al., 2002). The Hawaiian pyrolite was based on the 1959 eruption compositions, and thus the Ca/Al ratio of Hawaiian pyrolite matches the S-5g glass in Figure 4A but is too high for the Puna Ridge glasses.

In Figure 4B, the calculated parental compositions show a range of mantle equilibration pressures from ~1 GPa (57-9F) to 2.5 GPa (57-11c). The two best-constrained parental compositions (57-13g and 57-15D) are both consistent with a pressure

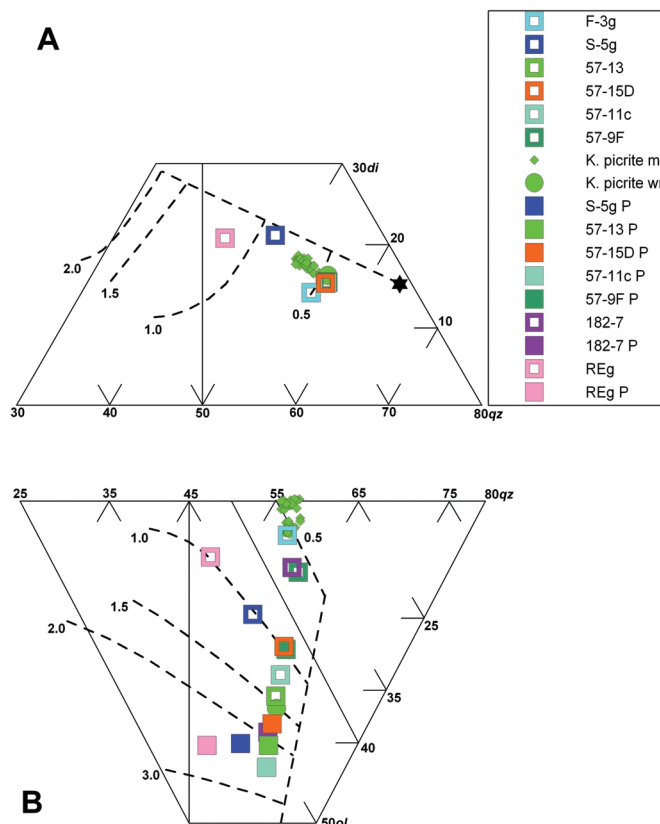


Figure 4. Projection of partial melting trends for Hawaiian pyrolite composition (dashed curves, 0.5–3.0 GPa), i.e., liquid compositions lying on lherzolite (olivine [ol] + orthopyroxene [opx] + clinopyroxene [cpx] + spinel) and harzburgite (ol + opx + chromian-spinel) within the normative basalt tetrahedron (Falloon and Green, 1988; Falloon et al., 1988; T.J. Falloon, unpubl. data, 2006). Projections are from (A) olivine and (B) diopside. Glasses and calculated parental liquid compositions (P) are from Tables 2 and 3. Kilauea picrite whole-rock (K. picrite wr) and melt inclusions (K. picrite ml) are from Norman and Garcia (1999) and Norman et al. (2002). Sample numbers in the key are from Tables 1 and 2.

of ~2 GPa and closely match experimental compositions from Hawaiian pyrolite at this pressure (Table 3). This pressure of mantle equilibration matches the expected pressure at the base of the oceanic lithosphere (~60 km) beneath Hawaii, based on geophysical studies (Eaton and Murata, 1960). Thus the parental liquid compositions calculated for 57-13g and 57-15D are consistent with the model in which a compositionally buoyant diapir (i.e., plume) is emplaced into the base of the oceanic lithosphere, and melt fractions are integrated and equilibrated with a harzburgite residue before segregation via melt channeling through the oceanic lithosphere to crustal magma reservoirs at ~0.2 GPa (~6 km) depth beneath Kilauea volcano. In such a model, the garnet signature observed in Hawaiian magmas is either the result of the integration of very small melt fractions derived from equilibrium with garnet over a range of pressures

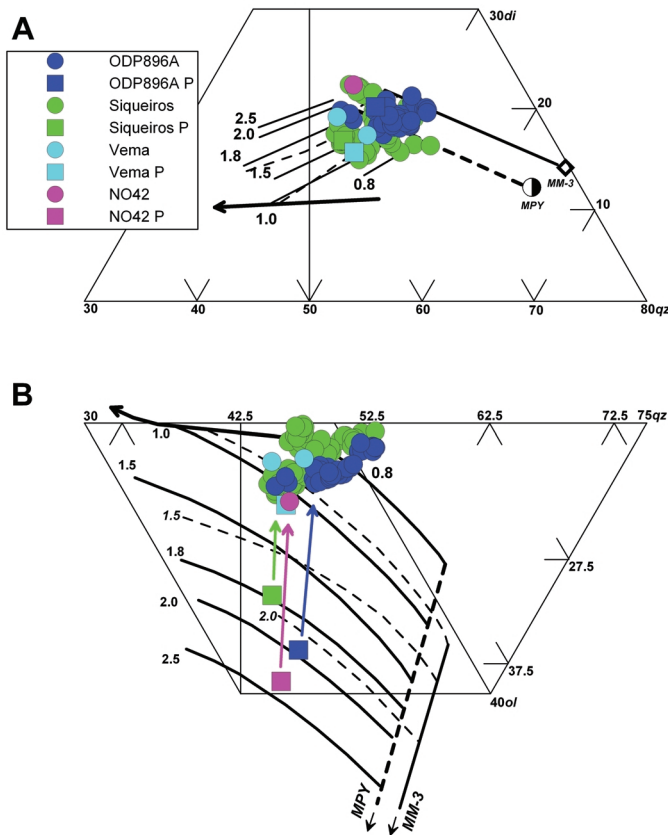


Figure 5. Projection of partial melting trends for MORB-pyrolite (MPY; thick solid lines, 0.8–2.5 GPa) and MM-3 lherzolite composition (thin dashed lines, 1.0–2.0 GPa) within the normative basalt tetrahedron (Falloon and Green, 1988; Falloon et al., 1999; Falloon et al., 2001; T.J. Falloon, unpubl data, 2006; Appendix 1). Solid line with arrow is the locus of liquids in equilibrium with plagioclase lherzolite (olivine + orthopyroxene + clinopyroxene + plagioclase ± chromian-spinel) at 1 GPa—such liquids are low in normative diopside and olivine. Projections are from (A) olivine and (B) diopside. Glasses and calculated parental liquid compositions (P) are from Tables 2 and 3. Data for ODP896A, Siqueiros, and Vema glasses are from L.V. Danyushevsky (unpubl data, 2006). Colored arrows represent olivine control lines from the respective parental liquid compositions.

(Eggin, 1992a,b) or is a source feature derived by reaction between melts caused by the incipient melting of eclogite and refractory harzburgite (Yaxley and Green, 1998).

However, the low-Fe (57-9F) and high-Fe (57-11c) glasses are inconsistent with this model. The calculated parental compositions to these glasses indicate pressures of ~1 GPa for 57-9F and ~2.5 GPa for 57-11c. It is very difficult to relate the differences in inferred pressures of mantle equilibration to any known mantle melting or reaction process, as both these glasses are colinear with glasses 57-13g and 57-15D in the projection from olivine in Figure 4A. That is, all the Puna Ridge parental glass compositions define a single olivine control line in both Figure 4A and B, whereas Figure 2 indicates that it is impossi-

TABLE 4. COMPARISON OF MODEL ODP896A PARENTAL LIQUID WITH EXPERIMENTAL MELTS FROM MM-3 AND MPY-87 AT 2 GPa UNDER ANHYDROUS CONDITIONS

Melt name	T-4107	ODP896A	T-4190
(wt%)			
SiO <sub>2</sub>	47.75	47.7	48.30
TiO <sub>2</sub>	0.61	0.54	0.74
Al <sub>2</sub> O <sub>3</sub>	13.96	13.46	13.40
Fe <sub>2</sub> O <sub>3</sub>	N.A.	0.85	N.A.
FeO	8.49	8.56	8.01
MnO	N.D.	0.10	0.15
MgO	15.37	15.86	15.95
CaO	11.83	11.44	11.48
Na <sub>2</sub> O	1.40	1.36	1.57
K <sub>2</sub> O	N.A.	0.02	N.A.
P <sub>2</sub> O <sub>5</sub>	N.A.	0.03	N.A.
Cr <sub>2</sub> O <sub>3</sub>	0.34	0.06	0.36
H <sub>2</sub> O	N.D.	0.04	N.D.
Temp*	1460	1474	1475
Mg#	76.3	76.8	78.0

Data sources: experimental glasses T-4107 and T-4190 (see appendix Table A3).

Note: N.D.—no data; N.A.—not applicable.

\*The anhydrous olivine liquidus temperatures calculated at 2 GPa using the model of Ford et al. (1983) and assuming all Fe as Fe<sup>2+</sup>.

ble for them to be related by olivine fractionation because of their strong differences in FeO<sup>T</sup> content. Some process other than mantle melting or reaction must be invoked to explain the compositions of glasses 57-11c and 57-9F. As both these glass compositions have no known matching whole-rock compositions, we consider them to be anomalous and therefore the mantle T<sub>P</sub> calculated for their parental liquids should be treated with caution.

**Mauna Loa.** The parental composition to the glass 182-7 is also consistent with a mantle equilibration pressure of 2 GPa. This observation further supports the view that 2 GPa represents the pressure at the base of the oceanic lithosphere and that major element compositions of parental liquids reflect equilibration with a harzburgite mantle source at this pressure.

**Réunion and Iceland.** Both parental compositions give depths of mantle equilibration of ~2.5 GPa (Figs. 4B and 5B).

**MORB.** In Figure 5, we compare the parental compositions of our three representative MORB glass compositions with experimentally determined cotectics from melting experiments on peridotite composition, MM-3 and MORB pyrolite. The parental compositions define a range of pressures (~1–2 GPa; Table 5) correlated with their FeO<sup>T</sup> contents. In particular, the calculated parental composition to the glass ODP896A shows a very close match to experimental reaction compositions using the parental composition for the same glass calculated by McNeill and Danyushevsky (1996) and peridotite MM-3 (Table 4, Appendix 1). This result gives support to the assumption that

TABLE 5. PETROGENETIC SUMMARY AND MODEL CALCULATIONS OF MANTLE POTENTIAL TEMPERATURES

Parent	$T$ (0.2 GPa)*	$T$ (final)†	$\Delta T$ (adiabat)§	$P$ (final)¶	$P$ (final)‡	MgO (wt%)**	H <sub>2</sub> O (wt%)††	Oliv Mg# (0.2 GPa)§§	Oliv Mg# (final)¶¶	Mantle residue‡‡	F (%)***	$T_p$ †††
<b>MORB</b>												
ODP896A	1351	1441	36	2.00	2.01	15.86	0.04	91.6	90.5	Sp LHz	0.20	1488
Siqueiros	1320	1400	32	1.80	1.65	13.93	0.06	91.5	90.4	Sp LHz	0.13	1425
Vema	1243	1286	16	1.10	1.16	10.86	0.10	90.5	89.8	Sp LHz	0.13	1318
<b>Iceland</b>												
NO42	1361	1468	41	2.40	2.34	16.75	0.08	92.4	91.3	Hz	0.17	1501
<b>Hawaii</b>												
S-5g	1335	1436	41	2.20	2.27	16.79	0.76	90.7	89.4	Hz	0.33 <i>0.12</i>	1524 <i>1454</i>
57-13g	1355	1451	39	2.10	2.16	17.86	0.57	90.7	89.5	Hz	0.40 <i>0.14</i>	1563 <i>1477</i>
57-15D	1341	1422	33	1.80	1.93	17.00	0.56	90.7	89.7	Hz	0.38 <i>0.14</i>	1531 <i>1451</i>
57-11c	1372	1488	47	2.50	2.44	18.8	0.49	90.7	89.3	Hz	0.44 <i>0.16</i>	1610 <i>1516</i>
57-9F	1286	1326	16	1.00	1.23	14.00	0.62	90.7	90.1	Hz	0.34 <i>0.12</i>	1429 <i>1356</i>
<b>Mauna Loa</b>												
182-7	1354	1445	37	2.00	1.92	17.53	0.46	91.3	90.3	Hz	0.42 <i>0.15</i>	1565 <i>1475</i>
<b>Réunion</b>												
REg	1323	1440	48	2.50	2.52	16.17	1.18	90.65	89.1	Hz	0.26 <i>0.09</i>	1502 <i>1445</i>

Notes: Values in italics for F and for  $T_p$  refer to calculations based on a refractory Hawaiian Pyrolite composition (harzburgite + 10% picrite). Hz—harzburgite ± Cr-spinel; MORB—mid-ocean ridge basalt; Sp LHz—spinel lherzolite.

\*Calculated olivine crystallization temperature of the parental liquids at 0.2 GPa (see Table 2).

†Calculated olivine crystallization at the pressure of mantle equilibration; see the column  $P$  (final)‡.

§Difference between the temperature of the parental liquid if it rises to the surface along a liquid adiabat and the olivine crystallization temperature at 0.2 GPa, i.e., a measure of possible superheat of the magma.

¶Pressure of mantle equilibration estimated from the position of the parental liquid compositions within the normative tetrahedron (see text for discussion)

‡Pressure of mantle equilibration estimated by the empirical calibration of Alabarede (1992).

\*\*MgO wt% of the calculated parental liquids (Table 2).

††H<sub>2</sub>O wt% of the calculated parental liquids (Table 2).

§§Calculated equilibrium liquidus olivine at 0.2 GPa (Table 2).

¶¶Calculated equilibrium liquidus olivine at the pressure of mantle equilibration; see column  $P$  (final)¶.

‡‡Residue mineralogy at the pressure and temperature of mantle equilibration.

\*\*\*Degree of partial melting estimated for the parental compositions.

†††Calculated mantle potential temperature derived from  $T_{\text{final}} + \Delta T_{\text{fusion}} - \Delta T_{\text{solid adiabat}}$ , where  $\Delta T_{\text{fusion}}$  is the latent heat of melting based on the model source and derived F%, and  $\Delta T_{\text{solid adiabat}}$  is the adiabatic cooling from  $P_{\text{final}}$  to surface along the olivine adiabat.

mantle melting can be modeled by simple batch melting, and that major elements of primary melts reflect the pressure of last equilibration with the mantle and do not represent a mixture of independent melt fractions finally assembled in a crustal magma chamber.

**Summary.** The results of this study, based on reasonable assumptions concerning source compositions and melting processes, indicate very little differences in pressures of mantle equilibration between MORB and OIB parental liquids (1–2 GPa versus 1–2.5 GPa). We have used the mantle equilibration

pressures and the models of Ford et al. (1983) and Falloon and Danyushevsky (2000) to calculate temperatures of equilibration with olivine at the pressures indicated from the normative tetrahedron analysis. These temperatures are listed in Table 5 ( $T_{\text{(final)}}$ ). As was the case with temperatures at 0.2 GPa, there is very little difference in temperatures between these parental compositions derived from similar pressures of mantle equilibration. The best-constrained compositions from Kilauea and Mauna Loa give temperatures and pressures of mantle equilibration essentially identical to the hottest MORB composition

ODP896A (1422–1451 °C versus 1441 °C). These results strongly suggest that there is unlikely to be any significant differences in source mantle  $T_p$  between OIB and MORB.

### Calculation of Mantle $T_p$

Based on the assumptions listed previously (see above), we present in Table 5 our results of mantle  $T_p$  calculations for the calculated parental compositions. The highest calculated MORB source  $T_p$  is 1488 °C, for Réunion 1502 °C, for Iceland 1501 °C, and the best-constrained estimate for Hawaii is 1565 °C. These calculations assume a source pyrolite (3–4 wt% CaO and  $Al_2O_3$ ) and require a high degree of melting of such a fertile source to yield harzburgite residue for OIB parental magmas. These simple but nonunique results demonstrate that there does not exist a significantly large difference in mantle  $T_p$  between OIB and MORB mantle. If the degree of melting for Hawaii is the lower of the values presented in Table 5, caused by a more refractory source than Hawaiian pyrolite, then the  $T_p$  differences between OIB and MORB source mantle would be essentially zero.

## CONCLUSIONS

In this study, we have modeled the reverse of olivine fractionation to calculate parental liquids for a range of OIB and MORB glasses. Our results suggest that there is very little difference in either the temperature of crystallization or the pressure and temperature of mantle equilibration between parental liquids in OIB and MORB settings. This result strongly suggests that it is unlikely that there is a significant difference in values of source mantle  $T_p$ . This conclusion is supported by simple nonunique calculations of mantle  $T_p$  based on a number of reasonable assumptions.

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## APPENDIX 1. EXPERIMENTAL PERIDOTITE-REACTION EXPERIMENTS ON A MODEL PARENTAL LIQUID COMPOSITION AT 1.8 AND 2.0 GPA

### Introduction

In this appendix, we present experimental results (see Tables A1–A3) of peridotite-reaction experiments between peridotite MM-3 (Baker and Stolper, 1994) and the parental liquid to glass ODP896A (glass ODP Leg 896A-27R-1, 124–130, pc.15, calculated by McNeill and Danyushevsky, 1996, at 0.01 MPa) at 1.8 and 2.0 GPa. At 2.0 GPa, we also present the results

of reaction experiments using the magnesian MORB composition ARP74 10-16 (Bougault et al., 1979).

### Experimental and Analytical Techniques

The starting compositions (ODP896A, ARP74, and MM-3; Table A1) were prepared from a mixture of analytical grade oxides and carbonates (Ca, Na) ground under acetone in an agate mortar. This mixture was pelletized and sintered overnight (~16–20 h) at 950 °C. An appropriate amount of synthetic fayalite was then added to the sintered mix, and the mixture was again ground under acetone, before storage in glass vials in an oven at 110 °C. Experiments were performed using standard piston-cylinder techniques in the High Pressure Laboratory, formerly housed in the School of Earth Sciences, University of Tasmania. All experiments used NaCl/Pyrex assemblies with graphite heaters, fired pyrophyllite and alumina spacers, mullite and alumina surrounds, graphite capsules, and a  $W_{97}Re_3/W_{75}Re_{25}$  thermocouple. The thermocouple enters the assembly through a composite two- and four-bore alumina sheath. The thermocouple was separated and protected from the graphite capsule by a 1-mm alumina disc. No pressure correction was applied to the thermocouple calibration. The thermocouple junction is formed by crossing the thermocouple wires utilizing the four-bore alumina sheath, which forms the top 5 mm of the alumina thermocouple sheath. All experimental components and starting materials were stored in an oven at 110 °C. Experiments were performed using the hot piston-out technique, and pressures are accurate to within  $\pm 0.1$  GPa. Temperature was controlled to within  $\pm 1$  °C of the set point using a Eurotherm(TM) type 818 controller.

At the end of each experiment, the entire experimental charge was mounted and sectioned longitudinally before polishing. Experimental run products were analyzed either by wavelength dispersive microanalysis using Cameca SX-50(TM) microprobes housed in the Central Science Laboratory, University of Tasmania (operating conditions: 15 KV, 20  $\eta$ A) or energy-dispersive microanalysis using a Cameca MICROBEAM™ microprobe housed in the Research School of Earth Sciences, Australian National University (operating conditions: 15 KV, 5  $\eta$ A). All glass analyses have been normalized to the composition of international glass standard VG-2 (Jarosewich et al., 1980), which was analyzed together with the glasses under the same analytical conditions.

Our experimental results are presented in Tables A2 and A3. In Table A3, electron microprobe analyses of selected run products and mass balance calculations are presented. Table A3 demonstrates that our run products are essentially homogeneous and produce good mass balance with low sums of residuals squared. Our run times of 24 h are sufficient to have produced experimental run products closely matching equilibrium assemblages for the respective bulk compositions used.

**TABLE A1. STARTING COMPOSITIONS USED (WT%)**

Composition	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	Cr <sub>2</sub> O <sub>3</sub>
MM-3	45.50	0.11	3.98	7.18	0.13	38.30	3.57	0.31	0.68
ODP896A	47.90	0.57	13.87	9.29	0.10	15.15	11.60	1.40	0.07
ARP74	50.82	0.83	15.20	8.12	N.A.	10.25	12.21	2.09	N.A.

Note: N.A.—not applicable.

**TABLE A2. EXPERIMENTAL RUN DATA ON PERIDOTITE REACTION EXPERIMENTS**

Number	Run number	Run temperature* (°C)	Time (hours, minutes)	Basalt (wt%)†	ML/ML‡	Basalt composition	Phase assemblage
<b>1.8 GPa</b>							
1	T-4154	1425	24	0.26	0.68	ODP896A	Ol+Opx+Cpx+L
2	T-4153	1450	24	0.31	0.61	ODP896A	Ol+Opx+L
<b>2.0 GPa</b>							
3	T-4107	1450	24	0.44	0.81	ODP896A	Ol+Opx+Cpx+L
4	T-4190	1450	24	0.37	0.68	ARP74	Ol+Opx+L
5	T-4191	1475	24	0.30	0.56	ARP74	Ol+Opx+L

Notes: Cpx—clinopyroxene; L—liquid; Ol—olivine; Opx—orthopyroxene.

\*Recorded temperature of the experiment.

†The wt. fraction of basalt composition added as a layer in the peridotite reaction experiments

‡The wt. fraction ratio between the initial amount of added basalt component to the final amount of melt component in the experiment obtained by mass balance (Table A3). Values <1 indicate that the peridotite component has melted and contributed to the melt phase during reaction. Thus the final equilibrium assemblage is the result of a melting and reaction process not a crystallization and reaction process (values >1).

**TABLE A3. COMPOSITIONS OF EXPERIMENTAL RUN PRODUCTS**

Run number	Phase	MB											
		MB BC*	MM-3†	Type	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Cr <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O
<b>1.8 GPa</b>													
T-4154	Olivine	0.424(4)	0.55(1)	N.A.	40.81	N.D.	N.D.	N.D.	8.52	0.21	50.11	0.34	N.D.
	Orthopyroxene	0.17(1)	0.22(3)	N.A.	54.51	0.07	4.86	1.25	5.07	0.1	31.29	2.76	0.09
	Clinopyroxene	0.02(1)	0.06(3)	N.A.	52.37	0.09	5.66	1.47	4.55	0.06	23.37	12.08	0.35
	Glass	0.38(1)	0.17(2)	N.A.	48.09	0.65	14.16	0.38	8.12	0.12	14.8	12.09	1.49
	Residual¶	0.0217	0.1699										
T-4153	Olivine	0.395(6)	0.56(1)	N.A.	41.43	N.D.	N.D.	N.D.	7.94	0.12	50.16	0.33	N.D.
	Orthopyroxene	0.09(1)	0.18(2)	N.A.	54.51	0.07	4.86	1.25	5.07	0.1	31.29	2.76	0.09
	Glass	0.51(1)	0.25(1)	N.A.	48.6	0.56	12.83	0.54	8.21	0.11	16.31	11.51	1.23
	Residual	0.0552	0.1662										
<b>2.0 GPa</b>													
T-4107	Olivine	0.306(5)	0.56(2)	N.A.	40.9	N.D.	N.D.	N.D.	9.27	N.D.	49.68	0.15	N.D.
	Orthopyroxene	0.11(2)	0.20(5)	N.A.	54.93	N.D.	4.63	0.69	5.51	N.D.	31.78	2.45	N.D.
	Clinopyroxene	0.04(2)	0.05(5)	N.A.	51.92	N.D.	5.47	1.04	4.97	N.D.	24.88	10.02	0.17
	Glass	0.54(1)	0.20(3)	N.A.	47.75	0.61	13.96	0.34	8.49	N.D.	15.37	11.83	1.4
	Residual	0.0341	0.3622										
T-4190	Olivine	0.263(4)	0.55(1)	A(3)	41.3(3)	N.D.	N.D.	0.25(2)	8.4(1)	0.10(4)	50.2(3)	0.31(0)	N.D.
	Orthopyroxene	0.185(7)	0.21(2)	A(3)	54.87(9)	0.09(2)	4.3(1)	1.1(1)	4.8(1)	0.13(4)	32.0(1)	2.5(1)	0.12(1)
	Glass	0.548(5)	0.23(2)	A(5)	48.3(1)	0.74(4)	13.4(1)	0.36(2)	8.0(1)	0.15(4)	15.95(7)	11.48(9)	1.57(3)
	Residual	0.0256	0.2603										
T-4191	Olivine	0.324(6)	0.56(1)	A(3)	41.3(3)	N.D.	N.D.	0.36(2)	8.0(1)	0.10(2)	50.3(7)	0.5(3)	N.D.
	Orthopyroxene	0.14(10)	0.17(2)	A(3)	55.6(2)	0.07(4)	3.4(2)	0.88(7)	4.83(7)	0.09(4)	33.0(1)	2.02(2)	0.09(2)
	Glass	0.535(8)	0.26(1)	A(5)	48.8(2)	0.66(3)	12.66(7)	0.43(3)	7.8(2)	0.15(2)	17.20(5)	10.8(1)	1.37(2)
	Residual	0.0636	0.1447										

Notes: Mass balance in weight fraction was performed using least squares linear regression and the software PETMIX. "Residual" refers to the square of the sum of the residuals. Numbers in parentheses next to each analysis or mass balance are 1σ in terms of the last units cited; e.g., 0.540(9) refers to 0.540 ± 0.009. A(n)—analysis is an average with the number of analyses used to calculate the average given in the parentheses. N.D.—not determined; N.A.—not applicable.

\*Mass balance using the bulk composition of the experiment.

†Mass balance using the peridotite composition MM-3 only.

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## DISCUSSION

### 1 January 2007, Dean C. Presnall

The only unequivocal magma composition is one that is produced from a direct analysis of a glass. The validity of calculated compositions, such as those produced by Falloon et al. (this volume) is always dependent on a series of assumptions. Despite thousands of microprobe analyses of mid-ocean ridge basalt (MORB) glasses, no compositions that show a trend of olivine-controlled crystallization have ever been reported, either from “normal” ridge segments or Iceland (over 400 glass analyses just from Iceland). In contrast, at Kilauea, Hawaii, direct analyses have shown the existence of a clear trend of olivine-controlled crystallization (Clague et al., 1991, 1995). MORB generation (including Iceland) at low and relatively uniform pressures (0.9–1.5 GPa) and potential temperatures (~1240–1260° C) (Presnall et al., 1979, 2002; Presnall and Gudfinnsson, 2007, written communication) explains both the inverse and positive Na<sub>8</sub>-Fe<sub>8</sub> correlations as direct melts from a heterogeneous mantle. In addition, it avoids the conundrum of the complete absence of olivine-controlled fractionation.

### 12 January 2007, James H. Natland

The fatal assumption behind estimation of temperatures of crystallization using olivine-liquid FeO<sup>T</sup>-MgO relationships is that the most forsteritic olivine in a rock, or surmised to be representative of a liquid in equilibrium with the mantle, is probably not related to the host liquid composition along a single closed-system liquid line of descent. That assumption is the entire basis for the procedure of adding incrementally more forsteritic olivine into liquid compositions until an “equilibrium” liquid is reached, but it is wrong. This is because most picrites are hybrid rocks, the results of magma mixing, which can be established by careful studies of crystallization histories. The mixing is of two types: (1) mixing between primitive magma strains near or

somewhere above their melt sources in the mantle and (2) mixing between primitive and differentiated magma strains.

An example of mixing of primitive magma strains is sample D20-3 in Table 1 of Falloon et al. (this volume), from Siqueiros fracture zone on the East Pacific Rise. The glass analysis is from Danyushevsky et al. (2003), but an interesting aspect of the history of this sample is that it was obtained from almost exactly the location of an earlier dredged picrite (Batiza et al., 1977) with a glass composition (Natland, 1989, Table 1, analysis SD7-C) that is identical to within parameters considered by Melson et al. (1976) to represent material from the same eruption. The glass contains olivine dendrites and plagioclase spherulites (Natland, 1980), and thus it is on a two-phase cotectic. Even though the rock is a picrite, the glass is not an olivine-controlled liquid. Nor indeed is even the most-magnesian MORB glass (Presnall et al., 2002). Olivine in SD7-C, at least, commonly occludes Cr-spinel within skeletal embayments that also contain glass. Some of the spinel is zoned. But most importantly it has a substantial range in Cr# but little in Mg# (Fig. 3 in Natland, 1989), indicating crystallization from similarly magnesian parental liquids but with significant differences in Al<sub>2</sub>O<sub>3</sub> content (Poustovetov and Roeder, 2000). Such differences cannot be the result of olivine crystallization and must reflect mixing at about the same temperature of melt strains derived from a heterogeneous source.

Pertinent here is that two other chemically distinctive picrites were obtained in the same dredge haul (one of these during the later expedition). The one studied by Natland (1989) in detail (SD-7A) has spinel with a similar range in Cr# as SD7-C but at systematically lower Mg#; associated olivine is also a bit more iron rich, and there are rare plagioclase phenocrysts. The glass has higher FeO<sup>T</sup>, and if one were to add olivine incrementally back into it to, say, a nominal primitive olivine composition of Fo<sub>91</sub>, then the estimated parental crystallization temperature (and glass MgO content) would be higher than that

of SD-7C. But from the minerals actually in the rock, there is no indication that it even makes sense to do this addition. The cooler picrite is the more iron-rich. The two samples did not derive from the same mantle composition, even though they erupted side by side and maybe at almost the same time. They simply sampled the array of primitive liquid compositions across the melting domain in two different ways.

A second example is provided by Icelandic picrites. Many, including samples from the Borgarhraun flow studied by Falloon et al. (this volume), contain phenocrysts of plagioclase and clinopyroxene (e.g., Slater et al., 2001; Maclennan et al., 2003). Primitive basalt from site 896 on the Costa Rica Rift also contains plagioclase phenocrysts (Shipboard Scientific Party, 1993). These basalt samples are proof positive that the liquids did not follow olivine-controlled liquid lines of descent. Besides the phenocrysts, melt inclusions in Icelandic picrites provide substantial evidence for mixing of primitive melt strains as well as evidence for pyroxenite in the melt source (Sigurdsson et al., 2000; Slater et al., 2001; Foulger et al., 2005).

Next, Hawaii, particularly Kilauea and Puna Ridge, provides examples of mixing between primitive and differentiated liquids. Such mixing has been amply demonstrated by petrological studies of the Kilauea rift system (e.g., Wright and Fiske, 1971). It has the effect of elevating FeO<sup>T</sup> contents of the hybrid beyond that which the primitive mixing component could have reached by differentiation of olivine alone. Such mixing results in artificially high temperature estimates. Clague et al. (1995) discussed this mixing in the paper that also presented the composition of the most primitive Hawaiian glass (a grain with no phenocrysts in a cored, thinly bedded sand). This type of mixing is also evident in the mineralogy of similar but more iron-rich, tholeiitic picrites from the Juan Fernandez Islands (Natland, 2003). Another possibility is that some Hawaiian and Icelandic melt strains derive from eclogitic components in mantle sources (e.g., Sobolev et al., 2000, 2005; Natland, this volume) that may be either more or less iron-rich than commonly construed mantle peridotite. This variation in source petrology will contribute to heterogeneity of melt strains contributing to erupted picrites and add further complexity to estimation of crystallization and potential temperatures. Therefore, no one should try to estimate these temperatures until the full crystallization histories of the rocks are understood.

### 27 January 2007, Keith D. Putirka and J. Michael Rhodes

Falloon et al. (this volume) attempt to show that magmas from mid-ocean ridges (MORB) and ocean islands (ocean island basalt [OIB]), or “plumes,” have similar mantle potential temperatures ( $T_p$ ). Their strategy contains five errors and an inconsistency. Errors 1 and 5 derive from a misunderstanding of the thermal implications of the plume model, errors 2 and 3 lead to overestimates of ambient mantle  $T_p$ , error 4 leads to an underestimate of hot spot  $T_p$ , and all errors minimize mantle excess temperatures ( $T_{ex} = T_p^{OIB} - T_p^{MORB}$ ):

1. Their physical model is wrong. Plumes represent point sources, so only the highest temperatures at any OIB are relevant to  $T_p^{OIB}$ , which repeats an error in Green et al. (2001). Falloon et al. (this volume) also presume that only the highest MORB temperatures represent ambient mantle, contradicting our understanding of the linkages between mantle temperatures, bathymetry, and geochemistry (Langmuir et al., 1992)—and no alternative explanation is provided.
2. In Falloon et al. (this volume) MORB with low-moderate FeO<sup>T</sup> are ignored. However, it is impossible to generate low—moderate FeO<sup>T</sup> MORB from high FeO<sup>T</sup> MORB by fractionation of olivine ± plagioclase (their 1243°C and 1320°C trends cannot be derived by fractionation from their 1351°C trend). These low—moderate FeO<sup>T</sup> MORBs must reflect differences in ambient  $T_p^{MORB}$  and cannot be ignored when estimating ambient  $T_p$ ; source heterogeneity does not alleviate the problem, because olivine thermometry is independent of source composition (Putirka, 2005).
3. Falloon et al. (this volume) overestimate mean MORB FeO<sup>T</sup> (at 9.3% FeO<sup>T</sup>, for MORB glasses with >9.5% MgO [ $n = 80$ ]). But there are at least 137 glasses in PetDB with MgO > 9.5%, and they average 8.5% FeO<sup>T</sup>. And because olivine fractionation does not affect FeO<sup>T</sup>, there is no reason to exclude whole rocks with >9.5% MgO, which combined with glasses yield an average of 8.2% FeO<sup>T</sup> for MORB ( $n = 192$ ).
4. Falloon et al. (this volume) assume a very low Fe<sup>2+</sup>O for Hawaii. At Mauna Loa, sample 182-7 has 9.38% FeO—much lower than Hawaiian picrites, which average 10.4–10.6% Fe<sup>2+</sup>O (Herzberg and O'Hara, 2002; Putirka et al., 2007). This underestimate of Fe<sup>2+</sup>O at Hawaii artificially reduces  $T_p^{Hawaii}$  by >70°C.
5. The authors state that plume  $T_{ex}$  must be 200–300°C. Some models of excess bathymetry suggest that  $T_{ex} > 160°C$  (Sleep, 1990; Schilling, 1991), but the thermal plume model only requires that over a given depth range, the Rayleigh number is above critical. At Iceland, where acoustic anomalies extend to 670 km (Foulger et al., 2005),  $T_{ex}$  of a few tens of degrees Centigrade would be more than sufficient to support thermally driven active upwelling.
6. Falloon et al. (this volume) are inconsistent in that they use the highest FeO (and  $T$ ) at MORB but not at OIB.
7. Concluding statements aside, Falloon et al. (this volume) present a convincing case that some ocean islands have very high  $T_{ex}$ . If we eliminate only error 6, they derive a minimum  $T_{ex}$  of 122°C at Hawaii; taking the average of their  $T_{ex}$  estimates at Hawaii and MORB, their minimum  $T_{ex}$  is 126°C (Falloon et al., this volume, Table 5). Had Falloon et al. used observed  $fO_2$  and FeO for Mauna Loa (Rhodes and Vollinger, 2004), their primitive magma would have 20.6% MgO (assuming their implied  $K_D(Fe-Mg)^{ol-liq} = 0.318$ ; Table 2 in their article), and their minimum  $T_{ex}$  would be ~70°C hotter.

Their article is the third attempt by this group to argue that at Hawaii  $T_{ex} = 0^\circ\text{C}$ . Green et al. (2001) suggested that because Siqueiros (along the East Pacific Rise) and Hawaii yield olivine phenocrysts with similarly high forsterite (Fo) contents, that  $T_p$  must be similar. Putirka (2005), however, showed that because Hawaiian lavas contain more FeO, Hawaiian  $T_{ex}$  is at least  $220^\circ\text{C}$ . More recently, Falloon et al. (2007) suggested that the  $T_p$  estimates of Putirka (2005) at Hawaii were too high due to model error. But Putirka et al. (2007) show that Hawaii, Iceland, and Samoa have high  $T_{ex}$  regardless of which thermometer is used (and that the Ford et al., 1983, thermometer has systematic error not present in Putirka et al., 2007, and Beattie, 1993). Finally, Putirka et al. (2007) demonstrate that:

1. Excluding Iceland, MORBs exhibit a  $T_p$  range of  $140^\circ\text{C}$  with a standard deviation of  $\pm 34^\circ\text{C}$ ;
2. The MORB  $T_p$  range is  $210^\circ\text{C}$  when Iceland is included—consistent with Langmuir et al. (1992); and
3. Hawaii, Iceland, and Samoa have  $T_p$  values that do not overlap with MORB within  $2\sigma$ .

It is thus safe to conclude not just that the mantle thermally convects, but that convection currents drive intraplate volcanism.

Finally, although we disagree with the conclusions of Falloon et al. (this volume), unlike Natland (see his comment of 12 January), we do not find fault with their general approach. We agree with Natland that the strong linear Hawaiian trends are mixing trends and not liquid lines of descent (Rhodes and Vollinger, 2004). Nonetheless, Natland's suggestion that parental magmas are lower in FeO, and bear little relationship to the mixing trends, is a red herring. Perusal of Figure 11 in Rhodes and Vollinger (2004) will make this point clear.

### 28 January 2007, Dean C. Presnall

In their comment of 27 January, Putirka and Rhodes accept the arguments of Langmuir et al. (1992) that mantle temperature, bathymetry, and geochemistry are linked. However, a global examination of MORB glass analyses in the Smithsonian database shows no such linkages (Presnall and Gudfinnsson, 2007). Instead, the observed Na8-Fe8 systematics of MORBs match the systematics of melts at the lherzolite  $\pm$  basalt solidus of the CaO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-Na<sub>2</sub>O-FeO system in the narrow pressure-temperature range of 0.9–1.5 GPa and 1240–1260°C (Presnall et al., 2002; Presnall and Gudfinnsson, 2007). This low and globally uniform potential temperature along all ridges (including Iceland) is not consistent with the existence of hot plumes (Galápagos, Iceland, Azores, Tristan, Bouvet, Afar, Easter) on or close to ridges.

### 28 January 2007, Don L. Anderson

The statements and conclusions in Putirka and Rhodes (comment of 27 January) require linkages between mantle tempera-

tures, bathymetry, and composition (Langmuir et al., 1992) and seismic velocity that may not exist (Presnall and Gudfinnsson, 2007; Anderson, this volume). It is traditional to attribute melting, bathymetric, petrologic, and tomographic anomalies to variations in absolute temperature of a common parent rock. This simplistic approach underlies all plume speculations and much of the current discussion.

In their comment, Putirka and Rhodes attempt to show that magmas from mid-ocean ridges and ocean islands have mantle sources with distinctly different potential temperatures ( $T_p$ ) and similar compositions. They do not define “plume,” “normal mantle,” or “potential temperature,” but definitions can be extracted from their discussion. The inferred temperature differences are within plate tectonic and normal convection expectations. They may appear large in the context of the isothermal-mantle assumption that underlies the plume hypothesis.

Putirka and Rhodes make a series of critical but hidden assumptions;

1. Adiabatically corrected magma temperatures give the potential temperature at the source and also define the potential temperature and temperature gradient of the underlying mantle;
2. Variations in the temperatures of magmas are due to lateral variations in mantle temperature (and are not due to different depths or lithologies in and below the surface boundary layer); and
3. MORB from mature spreading ridges bracket the allowable temperature range of “normal mantle.”

It is evident that they consider temperature excesses of greater than a few tens of degrees to be sufficient to define a plume. Low acoustic velocities also uniquely define, to them, hot buoyant upwellings. The MORB range is considered to be about  $200^\circ\text{C}$  and is usually qualified as being from “ridges that are unaffected by plumes.” This qualification excludes new ridges, shallow ridges, and ridge segments near arbitrarily defined “hotspots.” It also ignores the variations in older MORB. They argue that only the hottest temperatures are diagnostic of plume temperatures, because plumes are from point sources. Although it is true that many of the experiments upon which the plume model is based use “point sources”—injection of a hot fluid through a narrow tube—it is not evident that there are such point sources in the mantle. “Normal mantle” in these studies is defined as isothermal and homogeneous, and the thermal effects of plates and continents are ignored.

The actual range in mantle temperatures is likely to be much greater than is observed along mature and “normal”-depth ridges. Regions of the mantle that have been covered by large plates for a long time are expected to be hotter than average (this effect has been called “continental insulation,” but it also applies to large oceanic plates). Thus, the mantle under Hawaii, Iceland, Samoa, and Afar is expected to be hotter than along the East Pacific Rise.

But lateral potential-temperature variations are not the only cause of magma temperature variations. There is no reason why the geotherm cannot cross the solidus within a conduction layer, particularly if the outer layers of the Earth are chemically buoyant. With a conduction gradient of  $10^\circ\text{C}/\text{km}$ , a typical MORB can start to form at a depth of 120 km while magma with an excess temperature of  $100^\circ\text{C}$  can start to form some 10 km deeper (the lithosphere is about half the thickness of the thermal boundary layer [TBL]). If both of these sources are then brought adiabatically to the surface, the latter will melt more and will appear to have a higher potential temperature. In neither case is it necessary that the source or underlying mantle be on an adiabat or be homogeneous.

The expected excess temperatures of mantle plumes are on the order of  $1200^\circ\text{C}$ , the temperature rise across a deep TBL. The usual explanation of the much smaller observed temperature excesses is that only the top of the deep TBL is involved. But the usual type of plume discussed in the geodynamics literature is a cavity plume, which involves the lowest-viscosity, lowest-density, and highest-temperature part.

The authors assume that excess bathymetry and “acoustic anomalies” are due entirely to excess temperature. They argue that some islands have “very high  $T_{ex}$ ” ( $\sim 120^\circ\text{C}$  at Hawaii); this value is only 10% of the temperature excess expected from a deep mantle plume. However, a modest temperature excess is quite consistent with normal variations in the shallow mantle from plate tectonic processes and with variations in the surface boundary layer.

Putirka et al. (2007) argue that current spreading ridges exhibit a petrological  $T_p$  range of  $140\text{--}210^\circ\text{C}$  ( $\pm 34^\circ\text{C}$ )—depending on how the data are filtered. This range is only half the global long-wavelength  $T_p$  range inferred from geophysics (Kaula, 1983; Anderson, 2000), but it still allows temperature excesses of  $\sim 270^\circ\text{C}$  ( $2\sigma$ ), sufficient to take the mantle from a state of incipient melting to extensive melting, even if the large range of lithologically plausible melting temperatures is ignored. It is probable however, that melting and tomographic anomalies mainly depend on lithology, and only secondarily on absolute temperature (Anderson, this volume; Foulger, this volume; Natland, this volume). But even the thermal and heatflow arguments do not require a deep mantle plume.

### 31 January 2007, James H. Natland

An illustration of Hawaiian picrite (Fig. D-1) may reduce the redness of the herring surmised by Putirka and Rhodes in their comment of 27 January, from my earlier comment. Two sorts of data may be obtained from picrites: bulk compositions and mineral compositions. Without the latter, inferences about incremental addition of olivine to liquids or olivine accumulation are subject to ambiguity. Bulk compositions do not reveal any of the potential combination of differentiated or primitive magma strains that may contribute to the whole. Ideally, one would like to have compositions of all liquids that contributed to the bulk rock.

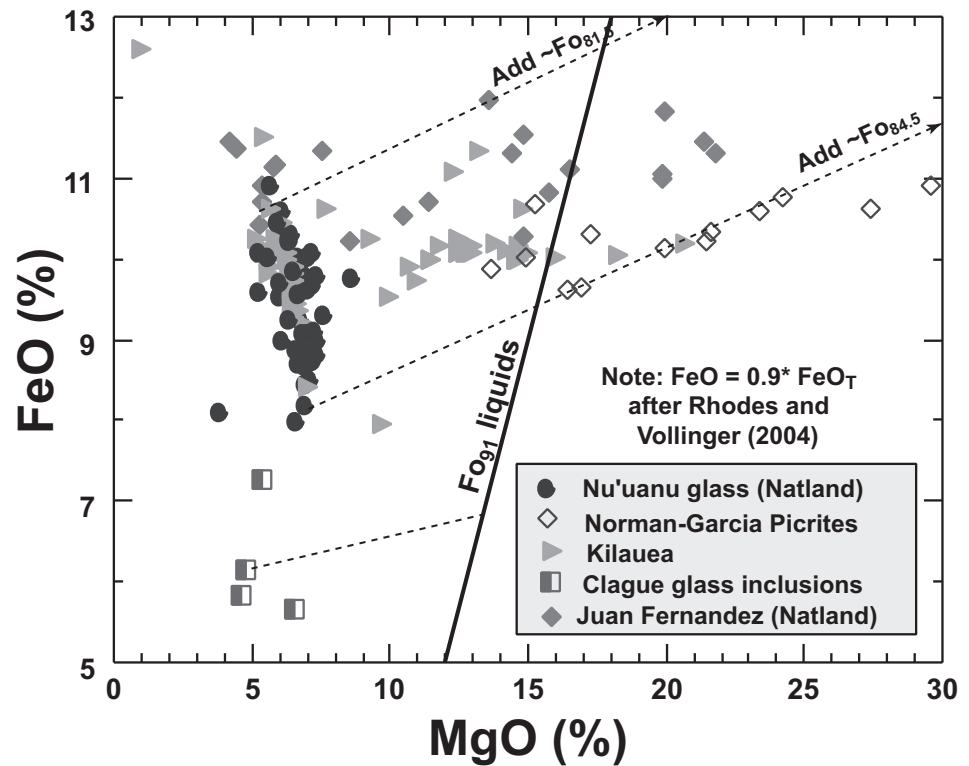
The Nu’uanu landslide was a major sector collapse of Oahu’s Ko’olau volcano, which was accompanied by one or more enormous submarine vitric pyroclastic eruptions. The eruptives cascaded out and around large downslumped blocks of the volcano (Clague et al., 2002) and over the seafloor for a distance of about 200 km, where beds of the glass-rich material cap the crest of the Hawaiian arch (Stephen et al., 2003). There, at Ocean Drilling Program (ODP) site 1223, amid several thin turbidites and mudstones (Garcia et al., 2006), two beds of glassy vitric tuff, each well indurated and several meters thick, were cored. The two beds consist mainly of angular shards of basaltic glass variably altered at their rims and, given the overall state of alteration of the rock, unusually fresh olivine. In bulk composition they are olivine-rich tholeiites and picrites (10.9–15.8% MgO; Shipboard Scientific Party, 2003). With  $\sim 20\%$  fresh glass, much of it enclosing olivine and Cr-spinel, they reveal in detail how Hawaiian picrite forms.

Most olivine-rich tholeiites and picrites from Kilauea and other Hawaiian volcanoes (Clague et al., 1995; Norman and Garcia, 1999) also have FeO contents (10–11%) within the range of Mauna Kea and Mauna Loa picrites. But some Kilauea glass has FeO contents as low as 8%, matching low Nu’uanu values, and some glass inclusions in olivine (Clague et al., 1995) have even lower FeO contents (5.6–7.2%). Basalt glass with low FeO content (5.6–9%) clearly exists at Kilauea volcano and within individual thin sections of Nu’uanu picritic vitric tuff. It is not a red herring. These liquids probably had more-magnesian parental precursors that may even have crystallized olivine as forsteritic as  $\text{Fo}_{91}$  (e.g., lowest dashed line in Fig. D-1), but for Nu’uanu, this suggestion is totally conjectural. Based on the equilibrium relationship with  $\text{Fo}_{91}$  in Figure D-1 (Rhodes and Vollinger, 2004), such liquid may only have had MgO of  $\sim 13\%$ , and a correspondingly lower eruptive temperature than any calculated at equilibrium with  $\text{Fo}_{91}$  for Hawaii by Putirka (2005) and Putirka et al. (2007). The Mg# of this liquid ( $\sim 72$ ) is about that of the most primitive MORB liquids.

Thus without well-documented crystallization histories, you cannot pin down which melt strain in a picrite produced  $\text{Fo}_{91}$ , which is almost never present in the rock anyway. Many olivine phenocrysts from Kilauea and other Hawaiian picrites (Wilkinson and Hensel, 1988; Clague et al., 1995; Norman and Garcia, 1999) are much more iron-rich ( $\text{Fo}_{85\text{--}78}$ ); accumulation of such olivine undoubtedly occurs there and elsewhere. However, from Figure D-1, if you follow the procedure advocated by Putirka and Rhodes, then tiny Juan Fernandez has to be significantly hotter than huge Hawaii. Instead, accumulation of iron-rich olivine (Natland, 2003; represented by the upper dashed lines in Fig. D-1) is clearly the reason why the bulk rocks there have higher FeO contents than found in Kilauea tholeiites. It is not a consequence of higher potential temperature. (Based on my own mineral data, the same consideration applies to Samoa.)

Mixing between primitive and differentiated compositions is the most serious bugbear in the calculation of primitive parental liquid compositions and potential temperatures. Ironically,

Figure D-1. Plot of FeO versus MgO for several suites of glass samples. None of the glass compositions from site 1223 and from dive samples from collapsed Nu'uuanu blocks nearer Oahu (Clague et al., 2002) are picritic. The picritic bulk compositions thus result from olivine accumulation. All glasses are differentiated, with MgO tightly clustered from 5 to 7%. FeO content increases from 8 to 11% as MgO decreases. This trend indicates a cotectic, not olivine-controlled, liquid line of descent. Olivine compositions are Fo<sub>85.4-75.5</sub>, with Fo correlating generally with Mg# of enclosing glass. In Figure 11 of Rhodes and Vollinger (2004), the FeO content of samples from the Hawai'i Scientific Drilling Project (HSDP)-2 drill site (Mauna Kea and Mauna Loa) varies from 10–11%, and the MgO value extends to 30%. The FeO content is at the high end of the range of the Nu'uuanu glasses. Inasmuch as original FeO content of liquids that produced the olivine and the olivine compositions themselves are unknown, all HSDP picrites thus could be produced by olivine addition to substantially differentiated cotectic liquids. Subtrends with higher FeO could simply indicate accumulation of more iron-rich olivine, such as at that at Nu'uuanu.



Rhodes and Vollinger (2004, p. 24) made exactly the same argument, observing "if our interpretation is correct, and of general application, the common practice of back-calculating to putative primary magma compositions through addition of incremental amounts of equilibrium olivine, may well lead to incorrect results and spurious correlations with MgO." Precisely.

**20 February 2007, T.J. Falloon,  
L. Danyushevsky, and D.H. Green**

In response to Presnall's comment of 1 January, we reiterate that our purpose is to identify the highest temperature liquids among MORB or OIB. We used glass compositions, analysed by electron microprobe, and calculated parental compositions by incremental addition of equilibrium olivine where the glass contains both its liquidus olivine and more magnesian olivine, using the latter as the target or limit justified by the petrography. We make no other assumptions.

Presnall comments "Despite thousands of microprobe analyses of MORB glasses, no compositions that show a trend of olivine-controlled crystallization have ever been reported." This statement is not true. It arises from inadequate coverage of published work and from the way Presnall and Gudfinnsson

(2007, written communication) select their glass data for petrogenetic interpretation. In our chapter we plot 190 glass analyses from mid-ocean ridge settings with >9.5% MgO, from the PetDB database. Among these, there are clearly glasses that lie in the ol+liq field at eruption pressures and temperatures (e.g., high-magnesian glasses from Siqueiros; Fig. D-2). In addition we have clear mineral-composition evidence from selected sites (Siqueiros, ODP hole 896A), which demonstrate the sequence of appearance of phases as ol (+rare spinel), ol+plag, ol+plag+cpx. Experimental demonstration of this sequence goes back at least to the paper of Green et al. (1979). We are not the first to identify this range of MORB glasses extending into the ol+liq field. In the petrogenetic analysis of global MORB glass chemical compositions of Presnall and Gudfinnsson (2007; written communication), the authors state that "the LKP (Langmuir et al., 1992) procedure of retaining only analyses with MgO values between 5 and 8.5% has been followed." The authors further restrict their dataset to the Smithsonian database. Langmuir et al. (1992) wished to select glasses lying on the ol+plag+cpx fractionation surface at low pressure and which approximate to a linear trend in several oxide versus oxide variation diagrams, particularly MgO versus FeO and MgO versus Na<sub>2</sub>O. They wished to avoid more-magnesian glasses (>8.5% MgO), which

depart from this multiply-saturated surface. Presnall and Gudfinnsson (2007, written communication) exclude higher temperature, MgO-rich glasses from their consideration and then state that there is no evidence for them, in spite of other authors' publications of such glasses.

In advocating a relatively low-temperature and low-pressure origin for MORB, Presnall and Gudfinnsson (2007, written communication) argue that parental MORB lie on the ol, opx, cpx, plag and/or sp saturation surface at 0.9–1.5 GPa, based on consideration of compositions interpolated at MgO = 8.0%. Neither these nor higher-temperature glasses with >9.5% MgO lie precisely on the 1-GPa ol, opx, cpx, plag, sp multiply-saturated surface when projected into multicomponent normative projections as in our article, Green et al. (2001), or Green and Falloon (2005)—they lie at higher normative diopside and lower normative olivine than the multiphase cotectic. In addition the glasses Presnall and Gudfinnsson (2007, written communication) at MgO = 8% are not sufficiently magnesian to be in equilibrium with mantle olivine (>Fo<sub>80</sub>). The more magnesian parental glasses derived in our article by olivine addition to Fo<sub>91–91.5</sub> have eruption temperatures greater than the restricted  $T_p = 1240–1260^\circ\text{C}$  of Presnall and Gudfinnsson (2007, written communication), further demonstrating the error in the low-temperature MORB model.

In his comment of 28 January, Presnall does not address our article directly but disputes the argument of Purтика and Rhodes (their comment of 27 January) for variable (including >1400°C)  $T_p$  in mid-ocean ridge settings. We infer variable eruption temperatures,  $P$ ,  $T$ , and percentage melting along ridges, but we consider that attribution of this variability to differences in  $T_p$ , differences in the departure from adiabatic upwelling, or source compositions is premature. We expect that all these factors have roles, and choice among them requires a much greater database of primitive melt compositions as assessed by methods used in our article.

In his comment of 12 January, Natland expresses reservations regarding our methodology of adding olivine to olivine-saturated glasses to infer more-magnesian and olivine-rich parental magmas. In our article we use only glass (liquid) compositions that contain olivines matching liquidus olivine compositions (derived from experimentally calibrated Fe/Mg partitioning; Ford et al., 1983) and contain, in addition, more-magnesian olivine phenocrysts and microphenocrysts. Our interpretation is consistent with the petrographic observations and is the simplest interpretation. We agree with Natland that most picrites are mixtures of crystals, particularly olivine, and the interpretation of liquid composition requires petrographic information on mineral compositions and their relation to crystal fractionation and possibly magma mixing (Danyushevsky et al., 2002). However, we are not using compositions of picrite rocks (i.e., rocks with >10–15% modal olivine). We are using glasses and only the highest-temperature and most-magnesian glasses lying in the ol+liq phase field at their liquidus temperatures. We demonstrate this in Figure D-2, and note that Natland's glass in

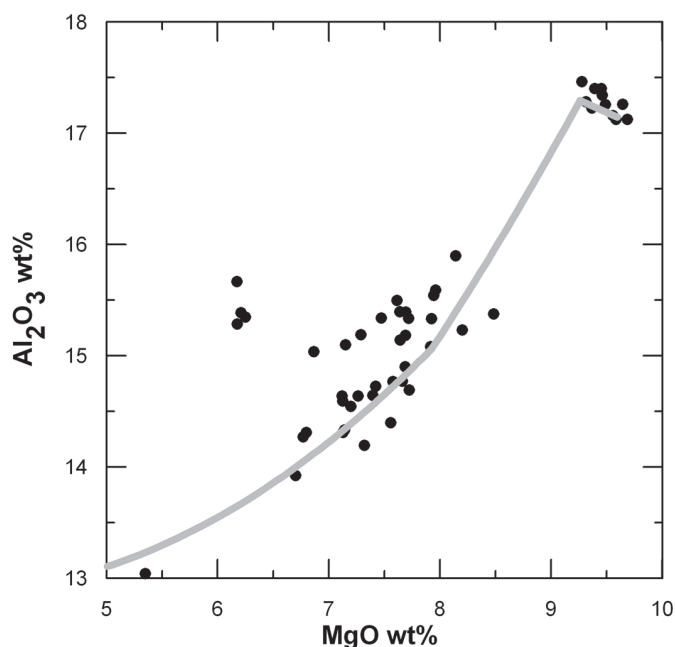


Figure D-2. Composition of glasses from the Siqueiros transform from Danyushevsky et al. (2003), and a fractionation trend (gray line) calculated from the most magnesian glass following the model (PETROLOG) of Danyushevsky (2001). This plot clearly demonstrates that the most magnesian glasses from Siqueiros are saturated in olivine only.

the picrite D20-3 lies on the ol+liq trend of Figure D-2, and some other glasses from Natland (1980, 1989) fall on the ol+plag trend of Figure D-1 between 8 and 9% MgO.

We also agree with Natland that most MORB glasses are evolved and the processes of evolution include crystal fractionation, “AFC” (assimilation, fractionation, crystallization) processes, reactive porous flow, magma mixing, and the like. This complexity is documented by Natland (1980, 1989) and clouds the interpretations of Presnall and Gudfinnsson (2007, written communication) and of Langmuir et al. (1992) using arrays of evolved glasses with 5.–8.5% MgO. However, by choosing the glasses as discussed, we identify the highest temperature melts (i.e., the most primitive observed liquids). There is no conflict between the recognition of the complexity of processes that a high-temperature picritic melt may undergo as it cools, reacts, crystallizes, and quenches to crystals + glass, and the identification of the highest temperature liquid (i.e., the liquid approaching or reaching an end-member “primitive magma”).

We respond to Natland's detailed comments on the Siqueiros locality in a data repository item supplement to this comment.<sup>1</sup> Natland (1989) documents evidence for mixing and reaction of

<sup>1</sup>GSA Data Repository Item 2007193, Expanded discussion and reply on Falloon et al. (2007) article, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA, or editing@geosociety.org, at [www.geosociety.org/pubs/ft2007.htm](http://www.geosociety.org/pubs/ft2007.htm).

high-temperature melts with crystalline inclusions, notably An-rich plagioclase. For a liquid already crystallizing olivine (+spinel), the energy required to dissolve phases with which the liquid is undersaturated is provided by exothermic olivine  $\pm$  spinel crystallization at the site of dissolution. This process may also lead to entrapment of ephemeral and local melt inclusions and zoned spinels, reflecting the competition between diffusive homogenization of melt and growth of olivine and spinel (Danyushevsky et al 2003). Natland summarizes evidence for more than one parental magma type in the Siqueiros sampling area. We also argue for significant source heterogeneity and complex evolution of primitive magmas at crustal and subcrustal depths.

In responding to Natland's comments on Iceland and Hawaii, we recognise complexity in the evolution of primitive melts, but our methodology, using both petrography and PETROLOG (Danyushevsky, 2001; Danyushevsky et al., 2003) modeling, identifies the highest-temperature melts in each sampled setting and explores the implications of such melts for constraining the actual melting and melt segregation process ( $P$ ,  $T$ , and residual phases) from a lherzolite or harzburgite source. Natland comments that "it is possible that Hawaiian and Icelandic melt strains derive from eclogitic components in mantle sources." Pyroxenitic or eclogite heterogeneity in mantle sources of MORB, and particularly OIB, may permit melting at temperatures below the peridotite solidus, but such melts can rarely reach the surface. Normally they react out of existence in the enclosing peridotite, "fertilizing" the latter and possibly acting as a flux by lowering the solidus temperature and causing melting of the new refertilized mantle. Such liquids will be picritic, controlled in their major-element composition by the residual phases of lherzolite and/or harzburgite but reflecting in their trace-element (incompatible elements) and isotopic composition their precursor (eclogite+residual peridotite) histories (Yaxley and Green, 1998; Yaxley, 2000).

Concerning the comments of Putirka and Rhodes (27 January), in the GSA data repository item (see footnote 1 in this discussion), we have inserted material appropriately. The comments by Anderson and Natland (31 January) address the comment of Putirka and Rhodes but do not directly address our article. We are in general agreement with them. Most importantly, Natland provides a figure and discussion illustrating the differences between olivine fractional crystallization and olivine phenocryst accumulation, refuting the approach of Putirka (2005) and Putirka et al. (2007).

Contrary to the statements of Putirka and Rhodes, we do not use a particular model with respect to hotspots and we are careful to document the three sequential steps followed to infer:

1. Liquidus temperature for the most magnesian glasses (Table 1 in Falloon et al., this volume);
2. Parental magmas to these glasses and  $P$  and  $T$  conditions of compatibility with peridotite residue (Tables 2, 3, and 4); and
3. Mantle potential temperatures, with emphasis on the as-

sumptions with respect to melt fraction/source composition and latent heat of melting and their effect on  $T_p$  estimation (Table 5).

We do not use mean temperatures or compositions, as these are meaningless when each magma batch, upwelling column or diapir, ridge segment, seamount, or volcano may have a different compositional, source  $P$  and  $T$ , or nonadiabatic ascent path to eruption. We reject the numerical and computational approach, which compiles and manipulates thousands of analyses of rocks and glasses to seek arithmetic algorithms to define geodynamic models, ignoring petrological information and physical reality. A more detailed discussion of Putirka (2005) and Putirka et al. (2007) is presented by Falloon et al. (2007).

Putirka and Rhodes incorrectly state that we ignore MORB with low to moderate  $\text{FeO}^T$  contents. We illustrate three MORBs with different  $\text{FeO}$ , different eruption and source  $T$  and  $P$  and thus different inferred  $T_p$  values. We do not advocate a unique ambient mantle  $T_p$  or plume  $T_p$ , but demonstrate the lack of any evidence for differences between the range of eruption temperatures or source  $T$  and  $P$  for MORB and hotspot primitive magmas.

In their point 5, Putirka and Rhodes (27 January) suggest that thermally buoyant plumes require minimal temperature differences from normal mantle. Other presentations of the deep-mantle plume hypothesis have weakened the original constraints of a fixed point source (plumes are deflected, flow along ridge axes, are intermittent, etc.) and unique source compositions (depleted mantle in plumes; enriched MORB on ridges; OIB chemistry in rifts, seamounts, and ridges) but the prediction and assertion of high magma eruption temperatures at hotspots has remained a cornerstone of the hypothesis (e.g., Campbell, 2006). Although in their comment Putirka and Rhodes appear to relax this difference to within the spread of MORB sources (see also Anderson's comment of 28 January), elsewhere (Putirka, 2005; Putirka et al., 2007) they support  $T_p$  for MORB of  $>1400^\circ\text{C}$  and excess  $T_p$  for "plumes" of  $\sim 200^\circ\text{C}$ . Such values for  $T_p$  should produce peridotitic komatiite magmas at hotspots with  $\sim 60\text{--}70\%$  melting of mantle lherzolite and residual dunite of  $\text{Fo}_{93\text{--}94}$ . These magmas are not seen.

Contrary to their statement in their point 6 (Putirka and Rhodes, 27 January), we use the observed range of  $\text{FeO}$  and eruption temperature for glasses and derive a similar range of eruption and segregation temperatures and pressures for both mid-ocean ridge and hotspot settings. Their point 7 and following text distorts our article and previous papers, and we refer readers to them and to Falloon et al. (2007) for detailed discussions of the methods of Putirka (2005) and Putirka et al. (2007) for inferring potential temperature and depths of origin of magmas. Both Putirka papers calculate increasingly large errors in estimating liquidus temperatures with increasing  $\text{MgO}$  content.

We agree with the comments of Anderson (28 January) concerning the Putirka and Rhodes discussion, and more importantly, we agree with his emphasis on compositional rather than thermal variability as causes for relative density, buoyancy, and



seismic properties within the upper mantle. In Green et al. (2001) and Green and Falloon (2005) we specifically advocate a key role for refractory, buoyant, old subducted slabs as causes for topographic and magmatic aspects of “hotspots.” We agree that the fuller exploration of the geodynamic consequences of mantle compositional heterogeneity is timely.

We have to our satisfaction tested and denied a key prediction (oft-quoted as a confirmation) of the deep-mantle thermal plume hypothesis—that their magmatic expression at hotspots is characterized by much-higher-temperature magmas and by greater depths and extents of melting than apparent in the ‘normal’ mantle upwelling and melting at mid-ocean ridges. We find on the contrary that the ranges of magmatic temperatures in “hotspot” and mid-ocean ridge settings are similar, and observed differences in magmatic products are a consequence of compositional heterogeneity in the upper mantle.

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