

Discussion of

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

by

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1st 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 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). This is in contrast to the situation at Kilauea, Hawaii, where 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, submitted) explains both the inverse and positive Na₈-Fe₈ correlations as direct melts from a heterogeneous mantle. In addition, it avoids the conundrum of the complete absence of olivine-controlled fractionation.

January 12th, 2007 James H. Natland

The fatal assumption behind estimation of temperatures of crystallization using olivine-liquid FeO-T-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. This 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; 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), thus 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 ocludes 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# (Natland, 1989, Fig. 3), indicating crystallization from similarly magnesian parental liquids but with significant differences in Al₂O₃ 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 FeOT, and if one were to add olivine incrementally back into it to, say, a nominal primitive olivine composition of Fo₉₁, 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's no indication that it even makes sense to do this. 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 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 FeOT contents of the hybrid beyond that which the primitive mixing component could have reached by differentiation of olivine alone. This 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., 2002, 2005; Natland, this volume) that may be either more or less iron-rich than commonly construed mantle peridotite. This 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 until the full crystallization histories of the rocks are understood.

January 27th, 2007 Keith D. Putirka & J. Michael Rhodes

Falloon et al. (this volume) attempt to show that magmas from mid-ocean ridges (MORB) and ocean islands (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., 1993)—and no alternative explanation is provided.
- 2) In Falloon et al. (this volume) MORB with low-moderate FeOt are ignored. However, it is impossible to generate low-moderate FeOt MORB from high FeOt MORB by fractionation of olivine \pm plagioclase (their 1243°C and 1320°C trends cannot be derived by fractionation from their 1351°C trend). These low-moderate FeO MORB 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) over-estimate mean MORB FeOt (at 9.3% FeOt, 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% FeOt. And because olivine fractionation does not affect FeOt, there is no reason to exclude whole rocks with >9.5% MgO, which combined with glasses yield an average of 8.2% FeOt for MORB (n=192).
- 4) Falloon et al. (this volume) assume a very low Fe²⁺O for Hawaii. At Mauna Loa, sample 182-7 has 9.38% FeO—much lower than Hawaiian picrites, which average 10.4-10.6% Fe²⁺O (Herzberg and O’Hara, 2002; Putirka et al., 2007). This underestimate of Fe²⁺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 $T_{ex}>160^\circ\text{C}$ (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 10’s°C would be more than sufficient to support thermally driven active upwelling.
- 6) The inconsistency: Falloon et al. (this volume) 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 estimates at Hawaii and MORB, their *minimum* T_{ex} is 126°C (Table 5). Had Falloon et al. (this volume) used observed fO_2 and FeO for Mauna Loa (Rhodes and Vollinger, 2004), their primitive

magma would have 20.6% MgO (assuming their implied $K_D(\text{Fe-Mg})^{\text{ol-liq}}=0.318$; Table 2), and their *minimum* T_{ex} would be $\sim 70^\circ\text{C}$ hotter.

This 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°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, MORB exhibit a T_p range of 140°C , with a standard deviation of $\pm 34^\circ\text{C}$,
- 2) the MORB T_p range is 210°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 2σ .

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

Finally, although we disagree with the conclusions of Falloon et al. (this volume), unlike Natland (see his comment of January 12th), 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 Fig. 11 in Rhodes and Vollinger (2004) will make this clear.

January 28th, 2007, Dean C. Presnall

In their comment of 27th 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, submitted). Instead, the observed Na8-Fe8 systematics of MORBs match the systematics of melts at the lherzolite \pm basalt solidus of the CaO-MgO-Al₂O₃-SiO₂-Na₂O-FeO system in the narrow pressure-temperature range of 0.9-1.5 GPa, 1240-1260°C (Presnall and Gudfinnsson, 2002; submitted). This low and globally uniform potential temperature along all ridges (including Iceland), is not consistent with the existence of hot plumes (Galapagos, Iceland, Azores, Tristan, Bouvet, Afar, Easter) on or close to ridges.

January 28th, 2007 Don L Anderson

The statements and conclusions in Putirka & Rhodes (comment of 27th January) require linkages

between mantle temperatures, bathymetry and composition (Langmuir et al., 1993) 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 & Rhodes attempt to show that magmas from mid-ocean ridges (MORB) and ocean islands (OIB) 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 & 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);
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°C and is usually qualified as being from “ridges that are unaffected by plumes”. This 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”. While it is true that many of the experiments upon which the plume model is based use “point sources”—e.g. 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 period of time are expected to be hotter than average (this 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 C/km a typical MORB can start to form at 120 km depth while magma with an “excess” temperature of

100°C can start to form some 10 km deeper (the lithosphere is about half the thickness of the 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 of the order of 1200°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} .” (~120°C at Hawaii); this is only 10% of the temperature excess expected from a deep mantle plume. On the other hand 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°C to 210°C ($\pm 34^\circ\text{C}$)—depending on how the data are filtered. This is only one-half of the global long-wavelength T_p range inferred from geophysics (Kaula, 1983; Anderson, 2000) but this still allows temperature excesses of ~270°C (2σ), 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; Natland, this volume; Foulger, this volume). But even the thermal and heatflow arguments do not require a deep mantle plume.

31st January, 2007 James H. Natland

An illustration about Hawaiian picrite (Figure 1) may reduce the redness of the herring surmised by Putirka and Rhodes in their comment of January 27th, 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, that 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 Site 1223, amidst 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 ~20% fresh glass, much of it enclosing olivine and Cr-spinel, they reveal in detail how Hawaiian picrite forms.

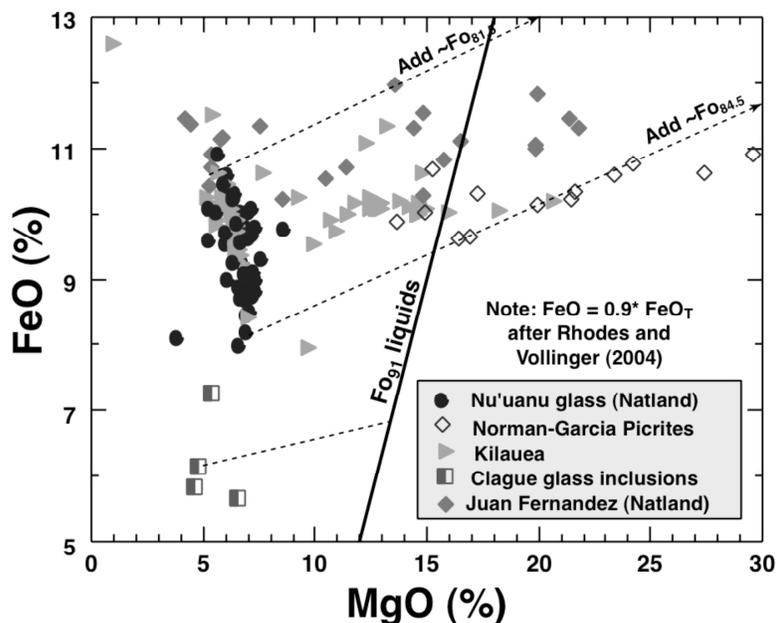


Figure 1: None of the glass compositions from Site 1223 and from dive samples from collapsed Nu'uauu blocks nearer Oahu (Clague et al., 2002) are picritic. The picritic bulk compositions thus result from olivine accumulation. All glasses are differentiated, with MgO tightly clustered from 5-7%. FeO content increases from 8-11% as MgO decreases (Figure 1). This trend indicates a cotectic, not olivine-controlled liquid line of descent. Olivine compositions are from Fo_{85.4}-75.5, with Fo correlating generally with MgNo of enclosing glass. In Figure 11 of Rhodes and Vollinger (2004), the FeO content of samples from the HSDP-2 drill site (Mauna Kea and Mauna Loa) ranges from 10-11% out to MgO = 30%. The FeO content is at the high end of the range of the Nu'uauu 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'uauu.

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'uauu 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'uauu 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 Fo₉₁ (e.g., lowest dashed line in Figure 1), but for Nu'uauu, this is totally conjectural. Based on the equilibrium relationship with Fo₉₁ in Figure 1 (Rhodes and Vollinger, 2004), such liquid may only have had MgO ~ 13%, and a correspondingly lower eruptive temperature than any calculated at equilibrium with Fo₉₁ for Hawaii by Putirka (2005)

and Putirka et al. (in press). The MgNo of this liquid (~72) is about that of the most primitive MORB liquids.

Thus without well-documented crystallization histories, you can't pin down which melt strain in a picrite produced Fo₉₁, 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 (Fo₈₅₋₇₈); accumulation of such olivine undoubtedly occurs there and elsewhere. However, from Figure 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 Figure 1) is clearly the reason why the bulk rocks there have higher FeO contents than Kilauea tholeiites. It is not a consequence of higher potential temperature. Parenthetically, 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, Rhodes and Vollinger (2004) 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 (p. 24)." Precisely.

20th February, 2007, T.J. Falloon, L. Danushevsky and D.H. Green

In response to Presnall's comment of 1st 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 is repeated in Presnall and Gudfinnsson (submitted), but it arises from inadequate coverage of published work and from the way Presnall and Gudfinnsson (submitted) select their glass data for petrogenetic interpretation. In our chapter we plot 190 glass analyses from MOR 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; Figure 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 (submitted), 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 approximating to a linear trend in several oxide vs oxide variation diagrams, particularly MgO vs FeO and MgO vs Na₂O, they wished to avoid more magnesian glasses (> 8.5% MgO) which depart from this multiply-saturated surface. Presnall and Gudfinnsson (submitted) 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 (submitted) 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 multi-component normative projections as in our chapter, 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 (submitted) at MgO= 8% are not sufficiently magnesian to be in equilibrium with mantle olivine (>Fo89). The more magnesian parental glasses derived in our paper by olivine addition to Fo91-91.5 have *eruption* temperatures greater than the restricted $T_p=1240-1260^{\circ}\text{C}$ of Presnall and Gudfinnsson (submitted), further demonstrating the error in the low-T MORB model.

In his comment of January 28th, Presnall does not address our paper directly but disputes the argument of Purtika and Rhodes (their comment of January 27th) for variable (including >1400°C) T_p in MOR settings. We infer variable eruption temperatures, P, T and % 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 amongst them requires a much greater database of primitive melt compositions as assessed by methods used in our paper.

In his comment of January 12th, Natland expresses reservations regarding our methodology of adding olivine to olivine-saturated glasses to infer more magnesian and olivine-rich parental magmas. In our paper we use only glass (liquid) compositions which 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 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 2 and note that Natland's glass in the picrite D20-3 lies on the ol+liq trend of Figure 2 and some other glasses from Natland (1980, 1989) fall on the ol+plag trend of Figure 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 etc.. This complexity is documented by Natland (1980, 1989) and clouds

the interpretations of Presnall and Gudfinnsson (submitted) 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 which a high-temperature picritic magma 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'.

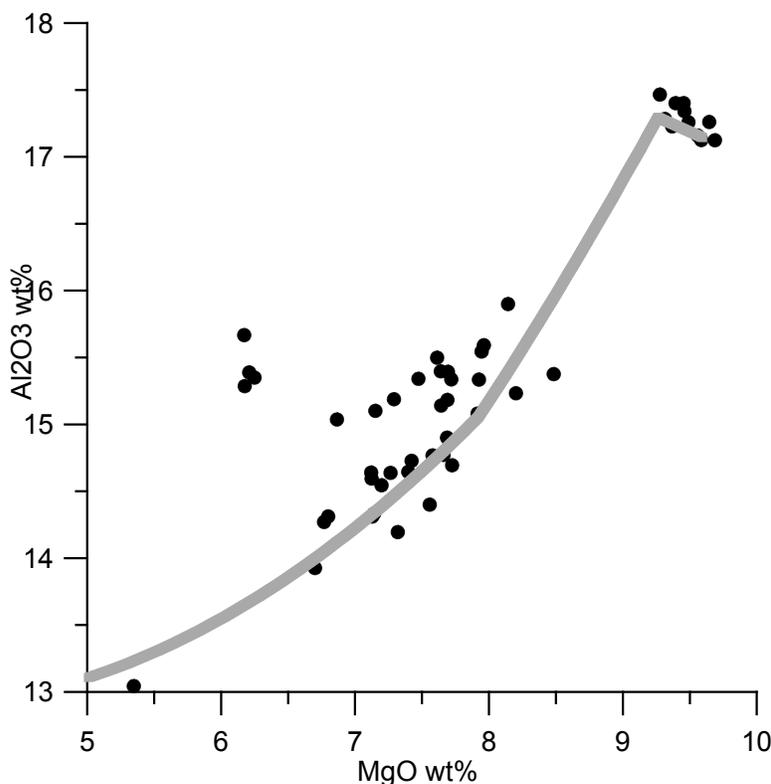


Figure 2: Composition of glasses from the Siqueiros Transform from Danyushevsky et al. (2003), and a fractionation calculated from the most magnesian glass following the model of Danyushevsky (2001). This plot clearly demonstrates that the most magnesian glasses from Siqueiros are saturated in olivine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

We respond to Natland's detailed comments on the Siqueiros locality in our electronic supplement to this comment (http://www.mantleplumes.org/P^4/P^4Chapters/Falloon_DiscElectronicSuppl.pdf). Natland (1989) documents evidence for mixing and reaction of 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+/-spinel crystallisation at the site of dissolution. This also may 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 summarises 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 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, 'fertilising' the latter and possibly fluxing melting of the new 'refertilised mantle'. Such liquids will be picritic, controlled in their major element composition by the residual phases of lherzolite/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 (27th January), in our electronic supplement (http://www.mantleplumes.org/P^4/P^4Chapters/Falloon_DiscElectronicSuppl.pdf) we have inserted material appropriately. The comments by Anderson and Natland (31st January) address the comment of Putirka and Rhodes but do not directly address our chapter. We are in general agreement with them. Importantly, Natland provides a figure and discussion illustrating the differences between olivine fractional crystallization and olivine phenocryst accumulation, refuting the approach of Putirka (2005; 2007).

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

- 1) liquidus temperature for the most magnesian glasses (Table 1),
- 2) parental magmas to these glasses and P,T conditions of compatibility with peridotite residue (Tables 2, 3, 4), and
- 3) mantle potential temperatures, with emphasis on the assumptions 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,T or non-adiabatic 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 FeO_t contents. We illustrate 3 MORBs with different FeO, different eruption and source T,P and thus inferred T_p 's. 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,P for MORB and hot spot primitive magmas.

In their point 5, Putirka and Rhodes 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, E-MORB on ridges, ‘OIB’ chemistry in rifts, seamounts and ridges) but the prediction and assertion of high magma eruption temperatures at ‘hot spots’ has remained (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 28th 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 hot spots with $\sim 60\text{-}70\%$ melting of mantle lherzolite and residual dunite of Fo93-94. These are not seen.

Contrary to their statement in their point 6, we use the observed range of FeO and eruption temperature for glasses and derive a similar range of eruption and segregation temperatures and pressures for both MOR and hot spot settings. Their point 7 and following text distorts our chapter and previous papers and we refer readers to them and to Falloon et al. (2007) for detailed discussion of the methods of Putirka (2005, 2007) for inferring potential temperature and depths of origin of magmas. Both papers calculate increasingly large errors in estimating liquidus temperatures with increasing MgO content.

We agree with the comments of Anderson (28th 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 ‘hot spots’. 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, i.e., that their magmatic expression at hot spots is characterized by much higher temperature magmas, and 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 range of magmatic temperatures in ‘hot spot’ 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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