

## ***Eruptive Temperatures of the Paleogene Picrites of West Greenland and Baffin Bay***



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### **Abstract**

*Tholeiitic picrites from West Greenland and Padloping Island on either side of Baffin Bay have long been considered to be high-temperature eruptives from the ancestral Iceland mantle plume. Olivine-liquid FeO-MgO equilibria have been used to backtrack to calculated parental compositions, giving as much as 20.2% MgO in the melt, and to estimate eruptive temperatures for them of close to 1500°C. Crystallization histories, however, indicate that these picrites are hybrids between olivine-charged crystal sludges and differentiated magmas that followed a low-pressure olivine-plagioclase-clinopyroxene cotectic during differentiation. The glasses give eruptive  $T = 1170\text{-}1220^\circ\text{C}$ . Much of the olivine in the rocks, especially the very magnesian olivine ( $\text{Fo}_{92-93}$ ), was incorporated by magma mixing and likely scavenging of minerals from mantle wall rock. The rocks are thus hybrids and the olivine in them neither lies along a single olivine-controlled liquid line of descent, nor can it be linked to the host compositions, which are too high in iron. Therefore they are unsuitable for the backtrack procedure. Compositions of Cr-spinel support a mixing history for the picrites, but the procedure of Pousovetov & Roeder (2000) suggests temperatures of spinel crystallization no higher than about 1340°C, generally less than 1300°C, and MgO contents no higher than ~16% in the most magnesian parental melt strains.*

### **Introduction**

I have applied criteria for identification of material suitable for calculation of eruptive and potential temperatures to Paleogene picrites of West Greenland and Padloping Island on either side of Baffin Bay (Francis, 1985; Robillard *et al.*, 1992; Pedersen & Larsen, 2000; Yaxley *et al.*, 2002). The locations are at 65°N and 71°N on the eastern and western coasts of Baffin Bay, respectively (Figure 1). Robillard *et al.* (1992), Larsen & Pedersen (2002) and Yaxley *et al.* (2002), as well as earlier workers, attribute picrite eruption to the influence of a high-temperature ancestral Iceland plume. The last two of these studies, plus Herzberg & O'Hara (2002), used olivine-melt MgO-FeO relationships to conclude that the rocks are the hottest basalts of a plume head that impacted North America at some 60 Ma, the tail of which now is supposed to underlie Iceland. Altogether, no studies other than those of Baffin Bay picrites provide so adequate a data set, including compositions of bulk-rocks, glass, melt inclusions, olivine, and spinel, with which to judge the question of temperature associated with this postulated plume.

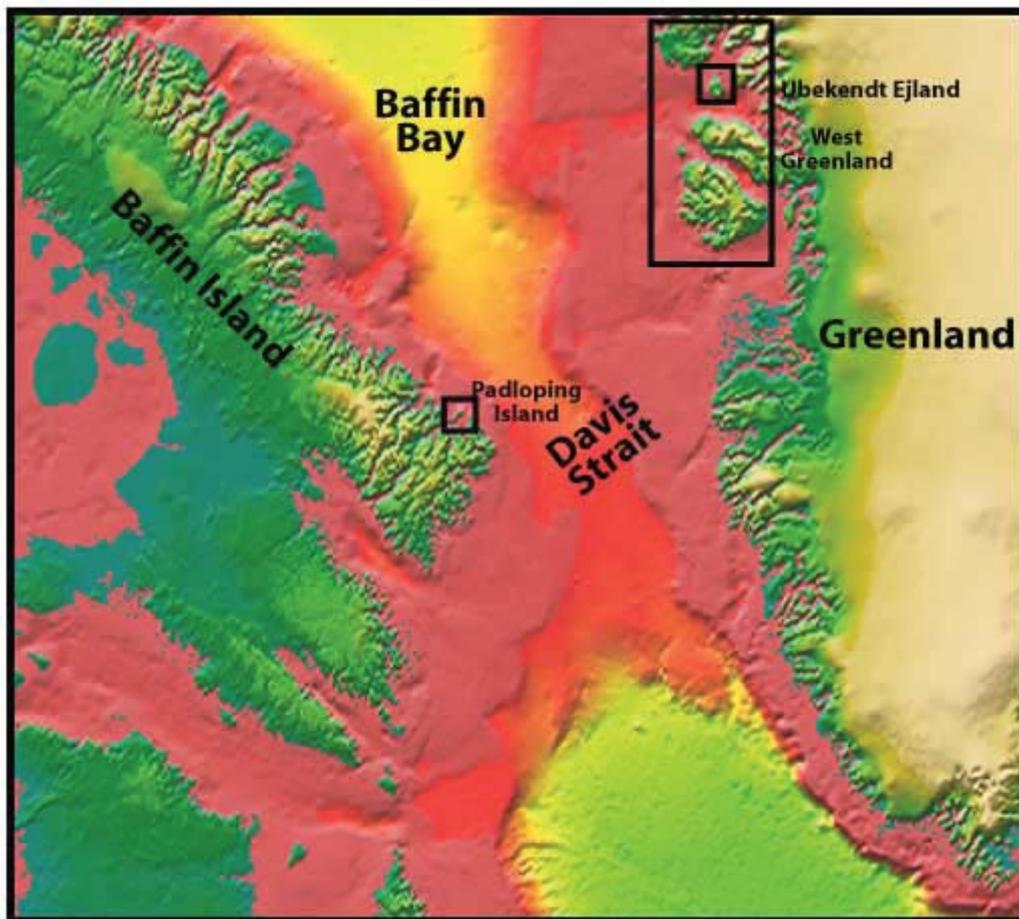


Figure 1. Location of localities around Baffin Bay mentioned in the text. West Greenland includes samples of Larsen & Pedersen (2000) and Graham *et al.* (1998). Padloping Island includes samples of Francis (1985), Robillard *et al.* (1992) and Yaxley *et al.* (2002). Ubekent Ejland is the location of xenoliths of Bernstein *et al.* (2006).

An unusual number of basalts on both sides of Baffin Bay are indeed picrites, with up to 29.8% MgO content in the bulk rock. They contain crystals of olivine as magnesian as Fo<sub>92.9</sub> that are both euhedral and normally zoned, and the olivine also occludes magmatic spinel. The minerals occur in many specimens and thus are not a special case. The abundance of so many picrites and the presence in some of them of strongly forsteritic olivine has long lent the aura of heat to petrological interpretations of this region (starting with Clarke, 1970 and Clarke & O'Hara, 1979). The question is, how hot was it?

### **Petrological Backdrop**

Following Natland (2007), basaltic glass in Baffin Bay picrite and associated basalt, analyses of which are plotted in Figure 2, strongly resembles glass found in Iceland in that it comprises tholeiitic low-Ti basalt (parental TiO<sub>2</sub> at MgO = 8% = Ti<sub>8</sub> < 1.4%), tholeiitic higher-Ti basalt and, in some formation members, nepheline-normative alkali olivine basalt (Larsen *et al.*, 2003). The stratigraphy also includes silicic lava or tuff (mainly dacite and rhyolite). In some studies, the tholeiitic higher-Ti basalt is termed E-MORB. The rocks are also similar to those of Iceland chemically, showing no "global trends" among variable Na<sub>8</sub>, Fe<sub>8</sub> and Ti<sub>8</sub> (oxide values corrected for differentiation to a nominal near-parental value of MgO = 8%; Klein & Langmuir, 1987). As at Iceland, the rocks are not, as MORB might be, related by simple variations in extent of partial

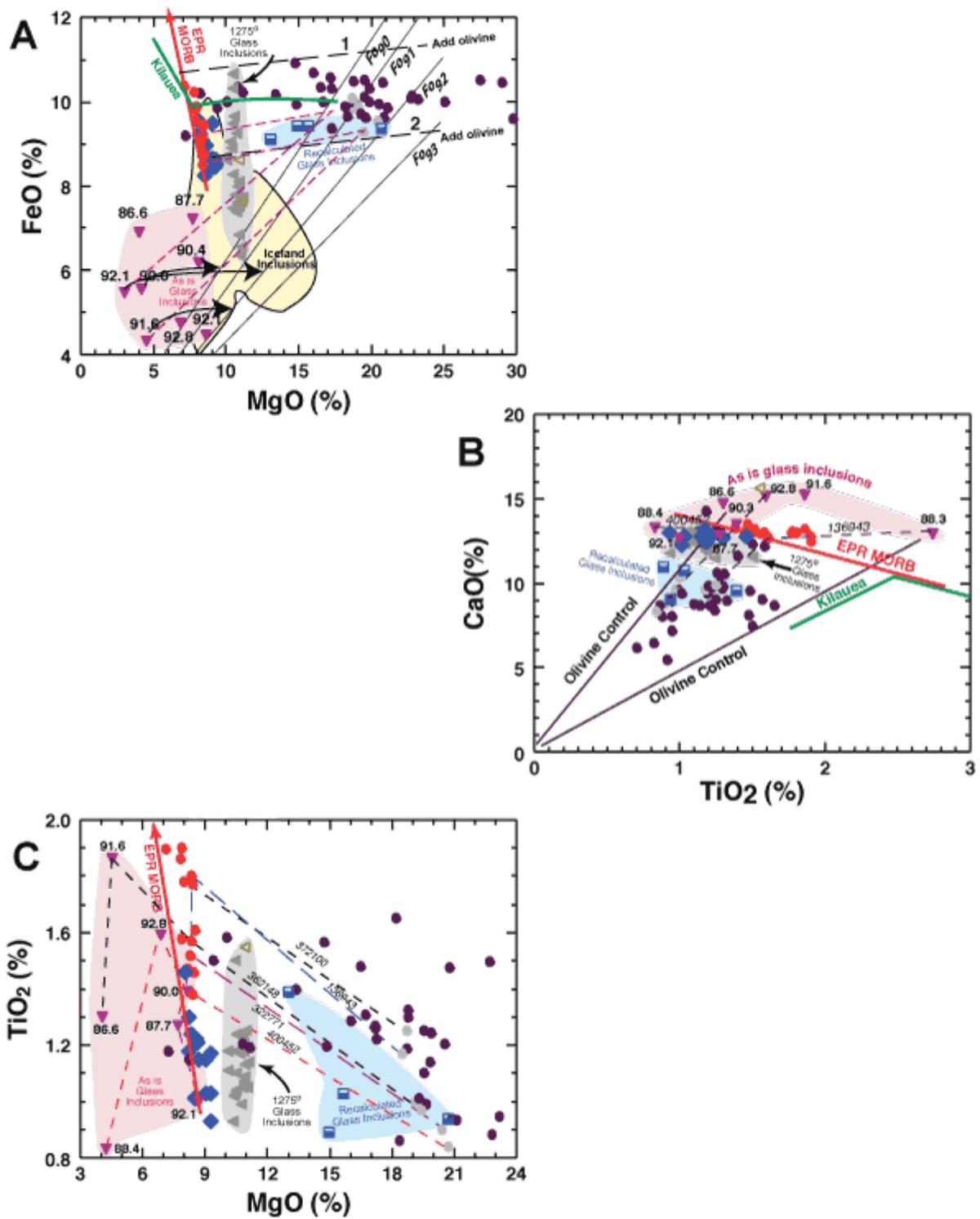
melting of a homogeneous mantle.

Somewhat surprisingly, no glass has  $Ti_8 < 1.2\%$ , whereas many primitive Icelandic tholeiites have  $Ti_8$  lower than 1%, reaching as low as 0.5% (Natland, 2007). Glass in tholeiitic basalt and picrite from Padloping Island has lower  $TiO_2$  and somewhat higher  $SiO_2$  and MgO than in basalt from west Greenland. Another surprise, however, is that Baffin Bay tholeiitic basalt and picrite have low  $\Delta Nb$  (Fitton *et al.*, 1997), with most low- $Ti_8$  basalts and picrites, dacite and rhyolite having negative  $\Delta Nb$  (by definition, less than Iceland), and most higher- $Ti_8$  basalt and picrite having positive  $\Delta Nb$ . Assimilation of silicic material by basalt, or basalt-rhyolite mixing, could explain the wide ranges in Nb/Y and Zr/Y and low  $\Delta Nb$  of many of the rocks (e.g., Natland, 2007). Graham *et al.* (1998) also consider that High  $^3He/^4He$  (R/Ra), (up to values of 30.6) in crushed olivine taken from West Greenland picrite is evidence for a mantle plume. However, the literature abounds in interpretations of geochemistry invoking crustal contamination or derivation of some fraction of the basalts by partial melting of ancient, previously melted continental lithosphere. Is very forsteritic olivine, unusually abundant in these rocks, and perhaps also the high  $^3He/^4He$  (R/Ra), which might be contained in fluid inclusions within it, a relic of an ancient melting event?

### **Glasses are Cotectic Liquid Compositions with Olivine Xenocrysts**

For evaluation of temperature, the most important aspect of analyzed glass on either side of Baffin Bay (Francis, 1985; Robillard *et al.*, 1992; Pedersen & Larsen, 2000) is that none of it is olivine-controlled (Figure 2).

*Figure 2: (Overleaf) Compositions of glass, bulk rocks, and melt inclusions from West Greenland and Padloping Island. A. MgO versus FeO, where  $FeO = 0.9 \times FeO_{TOTAL}$ . B.  $TiO_2$  versus CaO. C. MgO versus  $TiO_2$ . Data sources are West Greenland - Larsen & Pedersen (2002); Padloping Island - Francis (1985), Robillard *et al.* (1992), and Yaxley *et al.* (2002). Trends for East Pacific Rise (EPR) MORB glasses are regressions for analyses from the Smithsonian glass group data set (Magma Batch File, version 100199). Trends for Kilauea glasses are regressions for olivine-controlled (flat) and cotectic (steep) differentiation sequences, from Clague *et al.* (1995). Add olivine control lines 1 and 2 in A are drawn at upper and lower bounds on bulk rock compositions but with olivine-added trends using average olivine compositions in the samples; these are olivine accumulation, not liquid trends. Symbols in all diagrams are as follows. Glasses are: red circles – West Greenland; blue circles: Padloping Island. Whole rocks are: West Greenland – black and gray circles, with gray circles containing analyzed melt inclusions. Melt inclusions from West Greenland, as analyzed (as-is), are inverted maroon triangles. Next to these are forsterite contents ( $Fo$ ) of adjacent olivine, and variously dashed and colored lines connect them to their host glass and bulk-rock compositions (given by sample number in C). Glass inclusions from Padloping Island are: from experiments at 1275°C – left-pointing gray triangles and one open triangle, with higher CaO and lower  $Na_2O$  and  $SiO_2$  (see text); selected inclusions recalculated to have  $MgO > 12\%$  - half-filled blue squares. Inclusions are also enclosed in fields of different colors: West Greenland = light purple; Padloping Island – 1275°C = light gray; Padloping Island – recalculated = light blue. In A, the data field for Iceland glass inclusions is from Foulger *et al.* (2005). A also shows isopleths for MgO and FeO content in equilibrium with olivine of compositions from  $Fo_{90}$  to  $Fo_{93}$ , from Herzberg & O'Hara (2002). Arrowhead tips of the curving arrows give estimated as-is glass inclusion compositions in host olivines, interpolated to the olivine isopleth compositions.*



Glass compositions fall along a low-pressure plagioclase-olivine-clinopyroxene cotectic similar to that of depleted Pacific Mid-Ocean Ridge basalt (MORB; red arrows in Figure 2),

reaching ferrobasalt compositions, and the glass margins of the rocks contain both olivine and plagioclase microphenocrysts. The Padloping Island lava contains dunite fragments and broken crystals with kink bands. *Francis* (1985) originally and accurately described the kink-banded olivine as xenocrysts, and these were the only olivine crystals in his study with composition  $>Fo_{91}$ . Some of these very forsteritic crystals are normally zoned to about  $Fo_{88}$ , and have faceted outlines; these are xenocrysts with overgrowths. Therefore all these picrites are undeniably hybrid rocks.

Following the rationale developed for Samoa ([Natland, 2008](#)), backtrack calculations of *Larsen & Pedersen* (2000) to estimate ultimate parental MgO content and eruptive temperature cannot be correct when applied to cotectic glass compositions. Estimated parental MgO contents are too high as are estimates of crystallization and potential temperatures because the host glasses are iron enriched by crystallization along the low-pressure cotectic. Assumption of olivine control thus gives calculated parental MgO contents that are too high for melt MgNo (=  $Mg/[Mg+Fe^{2+}]$ ) adequate to be in equilibrium with  $Fo_{91-93}$  olivine in mantle sources.

### **Melt Inclusions**

*Larsen & Pedersen* (2002) determined compositions of melt inclusions in very forsteritic olivine. These systematically have lower FeO, lower MgO and higher CaO than host glasses (Figure 2). The most forsteritic olivine, which is by no means the most abundant olivine in the rocks, contributes to olivine-addition, as indicated by tie lines between glass and corresponding whole rock compositions. The inclusions contribute to the notion that the forsteritic olivine in which they reside is foreign to their host basalt, not lying along a common liquid line of descent with other olivine, nor linked to the host glass first by olivine-controlled and then by cotectic crystallization; this is why the rocks in general, whether taken as glasses or whole rock compositions, are unsuitable for backtrack calculations.

*Yaxley et al.* (2002) subjected their specimens to reheating to 1275°C in Pt capsules in order to re-equilibrate and homogenize partially crystalline glass inclusions in olivine (gray triangles in Figure 2). Considering that this experimental temperature might still be *too low*, they further adjusted the compositions of melt inclusions to a fixed value of total iron as FeO to take into account what they took to be post-crystallization re-equilibration with the surrounding olivine. This resulted in some of the glass inclusions having estimated original MgO contents of 13-20.7% (half-filled blue squares in Figure 2).

Considered as a group, however, the cluster of high-MgO inclusions melted at 1275°C does not fall along an olivine-controlled liquid line of descent, but parallels and spans the natural cotectic trend at somewhat higher MgO contents. All inclusions originally along a single olivine-controlled liquid line of descent would cluster tightly on this diagram. *Yaxley et al.* (2002) describe the rocks subjected to experiment as being “essentially identical” to Padloping Island picrites of the earlier studies of *Francis* (1985) and *Robillard et al.* (1992), which have low-pressure cotectic glass compositions and both faceted plagioclase and olivine in glass. The re-equilibrated inclusions are thus most likely simply host glass in cross sections of olivine embayments that actually have some of the adjacent olivine melted into them at an experimental temperature *greater* than that of cotectic liquids (Figure 3). The experimental temperature is thus *too high* to represent equilibrium for all inclusions. Based on this and also petrography, the compositions of liquids that hosted the phenocrysts studied by *Yaxley et al.* (2002) are not olivine controlled.

Using an original inference of *Francis* (1985) about uniformity of bulk compositions across lava flows, however, *Yaxley et al.* (2002) did not use glasses, but instead determined whole-rock X-ray fluorescence (XRF) analyses of quenched margins as the basis for backtrack estimates. The XRF analyses of the margins have variable MgO contents, from ~14-21%, at odds with the glass analyses of earlier studies on rocks of identical petrography. As whole rock rather than

glass compositions, these certainly contain variable amounts of olivine phenocrysts, and the analyses should not be used for backtrack calculations.

To summarize, as forsteritic as some of the olivine in these rocks is, it only circumstantially occurs in mixed rocks that may have had a number of magma stems antecedent to the host rocks, and which themselves were substantially differentiated (on a low-pressure cotectic) when olivine incorporation occurred. The crystallization and mixing histories thus are very like those of Hawaiian and Juan Fernandez picrite (*Helz, 1987; Clague et al., 1995; Natland, 2003*). Dikes or other intrusions of differentiated magma intersected crystal sludges – stagnant masses of some other magma at depth; these were charged with olivine phenocrysts and xenocrysts. The intruding dikes also scavenged some olivine and spinel from wall rock. Portions of all these materials were necessary to construct each final picrite. The pre-cotectic liquid line of descent of any given olivine is unknown, there may be more than one line of descent represented in mixed lava, and the lines of descent cannot be reconstructed using the crystallization history of the picrites.

### ***Further Difficulties with Reconstructed Melt Inclusions***

The reconstruction procedure of *Yaxley et al. (2002)* forces the compositions of the melt inclusions usually to higher but sometimes lower FeO than the simple experimentally re-equilibrated melt inclusions happened to have. The experimental melt inclusions have a restricted range of MgO contents, from 10.1-11.4%, as befits melts quenched at a single temperature of 1275°C, and the group of them is not olivine-controlled (Figure 2). This is in accord with quenching of experimental charges equilibrated at a single temperature. Additional reconstruction calculations stretch out the range of MgO contents, to 9.7-20.2%, and in particular force many inclusions to have high estimated original MgO contents (Figure 2A). This also provides a very crude and partial, but I would say only apparent, degree of olivine-control to the range of melt-inclusion compositions (light blue fields in Figure 2), seemingly justifying the conclusion that some parental liquid stems must have had very high MgO contents.

However, three assumptions are involved. The first is that re-equilibration at 1275°C was at too low a temperature simply because the FeO content *seems too low* when compared with whole-rock compositions; the second is that low-FeO inclusions must automatically have experienced Fe loss to surrounding olivine; the third is that an ancient continental lithospheric mantle source of unknown diversity, but which clearly provided a continental lithospheric geochemical signal to the picrites (*Yaxley et al., 2002*), provided no low-FeO magma stems to the hybrid state of these rocks, as they more obviously did to the West Greenland samples studied by *Larsen & Pedersen (2002)*. All three assumptions can be questioned, but all point in the same direction. The surest way ultimately to infer a high-temperature picritic composition is to add iron that was not measured back in to the original inclusion compositions, and then “correct” the MgO contents accordingly.

The suitability of melt inclusions for backtrack calculations in general can be questioned for two reasons. First, some inclusions may have been captured from boundary layers adjacent to growing olivine crystals and from which some elements excluded from the growing olivine might have become concentrated (*Natland, 2003*). The width of a boundary layer depends on cooling rate (*Gaetani & Watson, 2000*) and heightened concentrations of a variety of trace and minor elements might have occurred. Interpretations of, for example, the extent and type of enrichment of mantle sources (*e.g., Yaxley et al., 2002*) thus might be compromised.

Second, backtrack calculations may be compromised by high diffusion rates of FeO and MgO across olivine and into inclusions after mixing into a foreign magma (*Gaetani & Watson, 2000*). In numerical simulations, inclusion compositions in olivine tholeiite from the Reykjanes peninsula, Iceland, were modified in a matter of months. That is they were modified at rates

comparable to residence times in magma chambers and conduits, to become similar to those of host liquids, with somewhat lower MgO and distinctly higher FeO contents, depending on how differentiated the new host liquid happened to be. Thus mixing of primitive magmas into differentiated cotectic liquids tended to make inclusion compositions *more* iron rich than they were initially, adding to the potential to calculate parental crystallization temperatures that are too high. Furthermore, the final liquid compositions are more sensitive to changing temperature than olivine compositions; indeed, the effect on olivine compositions was minimal. Mixing throws melt inclusions out of equilibrium with host olivine, an effect that is particularly marked in the data of *Larsen & Pedersen (2002)*. Attempts by *Gaetani & Watson (2000)* to simulate addition of olivine back into melt inclusion compositions and then to follow the cooling history of the olivine failed to produce liquid lines of descent agreeing with the olivine addition path. They state “*This mismatch demonstrates the inherent potential for uncertainty in melt inclusion correction calculations due to diffusive exchange between an inclusion and its host olivine during the later stages of cooling*” (p. 35).

If these problems are ignored, the melt inclusions in olivine of composition  $Fo_{90-92}$  of *Larsen & Pedersen (2002)* with iron as measured give approximate maximum backtracked parental MgO contents of as little as 10-12% (lower curving arrows in Figure 2A). *Sigurdsson et al. (2000)* obtained similar values for inclusions in Iceland spinel without adjusting FeO contents. This is about the MgO content calculated for the most primitive MORB, which the suite otherwise resembles (e.g., it has low  $TiO_2$  contents and nearly matches MORB cotectic differentiation). This backtrack estimate only can be attempted assuming that the inclusions were not also crystallizing plagioclase or clinopyroxene at any point, which is unlikely based on the cotectic trend of host glasses and the presence of similar glasses at Iceland. One inclusion has CaO higher (open triangle in Figure 2B), and both  $SiO_2$  and  $Na_2O$  lower, than the others, suggesting that it may have included calcic plagioclase before heating to 1275°C. Compositionally similar low-FeO inclusions at Iceland occur not just in olivine as magnesian as  $Fo_{92}$ , but also in Cr-spinel, clinopyroxene and plagioclase in the same rocks (e.g., *Gurenko & Chaussidon, 1991*; *Sigurdsson et al., 2000*). They cannot all be corrected for iron loss in the same way as inclusions in olivine, thus a simple olivine-control correction is inadequate. The minerals therefore crystallized from liquids represented by their associated melt inclusions with their current compositions at temperatures of about 1300°C, based on experimental comparisons (see also [Foulger et al., 2005](#), who estimated similar temperatures using the glass geothermometer of *Beattie, 1993*). Thus many West Greenland and Icelandic melt inclusions have low FeO and also low MgO; this gives the low crystallization temperatures at glass  $MgNo = 67-78$ . The Padloping Island picrites are similar in most respects of composition and mineralogy, but do not have low-FeO melt inclusions. Instead, on all variation diagrams, almost all the Padloping inclusions are similar to host glasses (but with MgO systematically elevated by experiment); they may not have been inclusions at all, but cross sections of open skeletal embayments in the olivine.

Olivine in picrite thus might have many origins, including some crystals that might be from ancient dunite cumulates or fragments of refractory upper mantle residues either of a modern or an ancient partial melting event. Olivine as magnesian as  $Fo_{92-93}$  especially could be xenocrystic, since this is the evidently global predominant composition of olivine in ultramafic xenoliths derived from Archean mantle, including Ubekendt Ejland, West Greenland (*Bernstein et al., 2006, 2007*), which is within the West Greenland picrite province (Figure 1). Such olivine is interpreted to be in residual ultramafic rock left from an Archean partial melting event. Mafic liquid in equilibrium with this olivine during the ancient partial melting event could have precipitated magmatic olivine just as magnesian. That olivine may then and there have precipitated in dunitic veins, pods and tabular magma bodies between melt sources and the Archean crust, and eons later been extracted by dike rocks of differentiated Paleogene magmas. Whether this is likely or unlikely is difficult to say. However, could it really be a coincidence that such magnesian olivine just happens to occur in the lithosphere where “plume

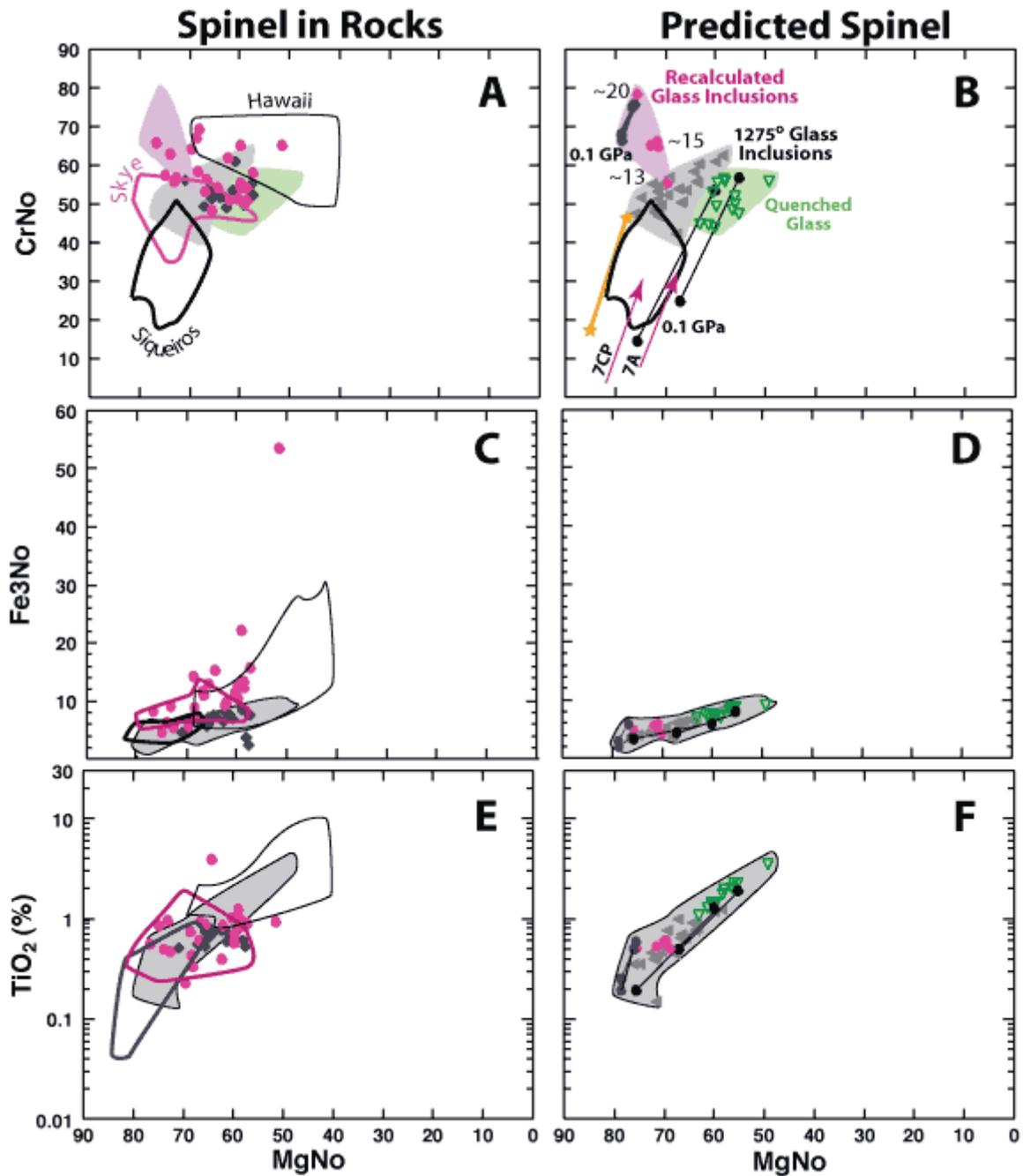
head” picrite erupts? At the minimum, simple existence of Fo<sub>92-93</sub> olivine in West Greenland and Baffin Bay picrites cannot be absolute proof that host basalt had a high-temperature precursor along any thread of a common liquid line of descent in Paleogene times. Some other means of assessment is needed.

## Spinel

Cr-spinel is a mineral of great petrological sensitivity, a ready indicator of aspects of magma composition and oxidation state (Allan *et al.*, 1988; Poustovetov & Roeder, 2000). Magmatic spinel is almost always euhedral and is most often enclosed in olivine, which it thus appears to precede in crystallization sequence. This can be tricky to judge, however, since olivine growth may trigger Cr saturation in the melt immediately next to it (c.f., Natland, 2003). The two minerals thus crystallize mainly along a two-phase cotectic, but during the very earliest crystallization history of any primitive basalt. We tend to think of olivine as being the first mineral to crystallize from basalt, but this is not technically correct. Dunite, ostensibly the cumulate produced during this stage, is never a “monomineralic” rock.

Poustovetov & Roeder (2000) developed a model from which the composition of chromite can be predicted from melt (glass) compositions. Their program is available as an Excel spreadsheet, *Melt-Chromite*. Spinel composition involves coupled substitutions between Fe<sup>2+</sup> and Al, and between Mg and Cr. Ti also partitions into spinel. In the structural formula for spinel, proportions of these five elements as well as minor elements that substitute for them allow Fe<sup>3+</sup> to be calculated from charge balance. For the crystallization model, however, the ratio of Fe<sup>2+</sup>/ (Fe<sup>2+</sup>+ Fe<sup>3+</sup>) in the glass must be assumed. For initial calculations and comparisons, I have taken this ratio to be 0.9, a common value for MORB (Christie *et al.*, 1986). I thus compare predicted spinel compositions for basalt glasses and either natural or experimental melt inclusions of samples from West Greenland and Padloping Island (Figure 3). The natural spinel compositions are in the left panels (Figure 3A, C, and E); predicted compositions are in the right three panels (Figure 3B, D, and F). The diagrams for CrNo (Cr/[Cr+Al]) and Fe<sub>3</sub>No (Fe<sup>3+</sup>/ [Fe<sup>3+</sup>+Al+ Cr + V]) versus MgNo (Mg/[Fe<sup>2+</sup> + Mg]) are conventional faces for the spinel prism. TiO<sub>2</sub> (rather than TiNo) is plotted semilogarithmically against MgNo in the lower two panels.

*Figure 3: (Overleaf) Comparison of natural spinel in West Greenland and Padloping Island picrites (left panels, A, C, and E) with spinel predicted to be in natural glasses and glass inclusions (right panels, B, D, and F) by the technique of Poustovetov & Roeder (2000). Data sources are as in A, as are symbols in left panels. Symbols in right panels are given by labels to shaded fields which are replicated without labels or symbols in A. Data fields labeled in A by place are for picrites and one plagioclase-phyric basalt from Siqueiros fracture zone (Natland, 1989); Hawaii – Kilauea and Puna Ridge (Clague *et al.*, 1995); and Skye, Scotland (J. Allan & J. Natland, unpublished data). Gray fields in D and F bound full data fields, rather than the distinct shaded fields in B, and are repeated in C and E. Lines link predicted selected spinel compositions calculated at 1 bar and 0.1 GPa. These data points are for natural glasses (black) and selected recalculated and bulk rock parental inclusions with MgO contents range from 13-20, as indicated near the light purple data field in B (blue; Yaxley *et al.*, 2002).*



In Figure 3B, predicted spinel compositions are given for quenched glasses, experimental glass inclusions at 1275°C of *Yaxley et al.* (2002), and various recalculated or estimated parental compositions to a maximum MgO content of 20.2% (*Yaxley et al.*, 2002; *Putirka*, 2008). The data fields, respectively green, gray, and lavender, are repeated in Figure 3A. In the lower panels, I show just a single all encompassing field (gray within a black line), which I replicate in the right panels, Figure 3, C and E. For several compositions, calculations at 1 bar and 10000 bars (10 kilobars or 0.1 GPa) are linked by tie lines. The predictions have the following features:

- Pressure increase produces strong variations in CrNo (Figure 3B) with MgNo for glass compositions with relatively low MgO contents;
- Pressure variations produce weak variations in CrNo if glass has high MgO contents (Figure 3B);
- High glass MgO contents (13-20%) produce high CrNo (Figure 3B);
- Cotectic crystallization causes CrNo to increase somewhat, and MgNo to decrease (inclined light green field in Figure 3B);
- Some spinel crystallized from glass with  $Fe^{2+}/(Fe^{2+} + Fe^{3+}) < 0.9$  (data points above gray field in Figure 3C); adjusting this downward from 0.9 in glass used in the model calculations would increase MgNo (by increasing melt MgNo at constant total iron) and CrNo in predicted spinel compositions.

The general result is that high glass MgO-contents (>15%) produce a restricted predicted field (lavender) on Figure 3B, one fairly insensitive to pressure, with spinel compositions having both high MgNo and high CrNo. Most of this field does not overlap natural spinel from West Greenland and Padloping Island, nor the boundaries of three other data fields provided for comparison (Hawaii, Skye, and Siqueiros Fracture Zone in the eastern Pacific). Predicted individual spinel compositions within the lavender field of Figure 3B, also have lower  $Fe^3No$  and  $TiO_2$  (purple and blue data points linked by lines in Figure 3, D and F) than natural spinel. Nearly horizontal trends of spinel in Figure 3, A and E, result from mixing between spinel-bearing high-MgO and low-MgO (differentiated) magmas.

*Poustovetov & Roeder (2000)* suggested that compositional variability in glasses, especially in  $Al_2O_3$  contents, could explain the variability of MORB spinel, and especially invoked the role of plagioclase fractionation in producing such variability. They thought differences in glass  $Al_2O_3$  content of as little as 2%, by weight (with  $Al_2O_3$  varying from 15-17%) would be sufficient to explain variability in CrNo. Since many primitive MORB have abundant plagioclase phenocrysts, this conclusion might be justified. However, the samples from Siqueiros Fracture Zone are picritic, with no plagioclase phenocrysts, thus the variability of spinel CrNo there seems likely to indicate the effect of pressure. All spinel at West Greenland and Padloping Island, however, has fairly restricted and high CrNo, suggesting that it all crystallized at shallow levels in the crust.

Spinel thus fairly positively indicates that the compositions of the basaltic glass from which it precipitated around Baffin Bay had MgO contents no higher than ~15%. No natural spinel anywhere matches compositions for any estimated parental basaltic liquid with MgO content >18%, as predicted for 26 out of 29 island locations by *Putirka (2008)*. Even the estimated parental glass for Siqueiros Fracture Zone of *Putirka (2008)* seems too high in MgO contents (13.2%) to explain the spinel variability in picritic rocks there (Figure 3B). The entire spinel field in Figure 3B is to the right of the estimated parental spinel compositions, and crystallized from melt with lower MgO content.

## Summary

Figure 4 summarizes estimated predicted temperatures of crystallization at atmospheric pressure for compositions of glass margins and inclusions in picritic samples from West Greenland. The temperature algorithm is that of *Beattie (1993)*, which is available in the program Melt-Chromite of *Poustovetov & Roeder (2000)*. This gives temperature estimates a bit lower than the scheme of *Ford et al. (1993)*. For example, the average temperature of inclusions heated experimentally to 1275°C by *Yaxley et al. (2002)* is about 25° less than experimental with the Beattie procedure. But all those compositions cluster tightly in both T and MgO content on this diagram, as they should. In contrast, five compositions recalculated from these with

higher MgO content project linearly to excessively high T from compositions of natural glasses and the homogenized inclusions. The maximum limit for MgO content and temperature based on spinel is given by the dashed outer boundary of the light blue field. The solid line is a more realistic limit for spinel crystallization at most places, including Samoa (Natland, 2008), Iceland and Padloping Island, with only two West Greenland spinel compositions falling outside it.

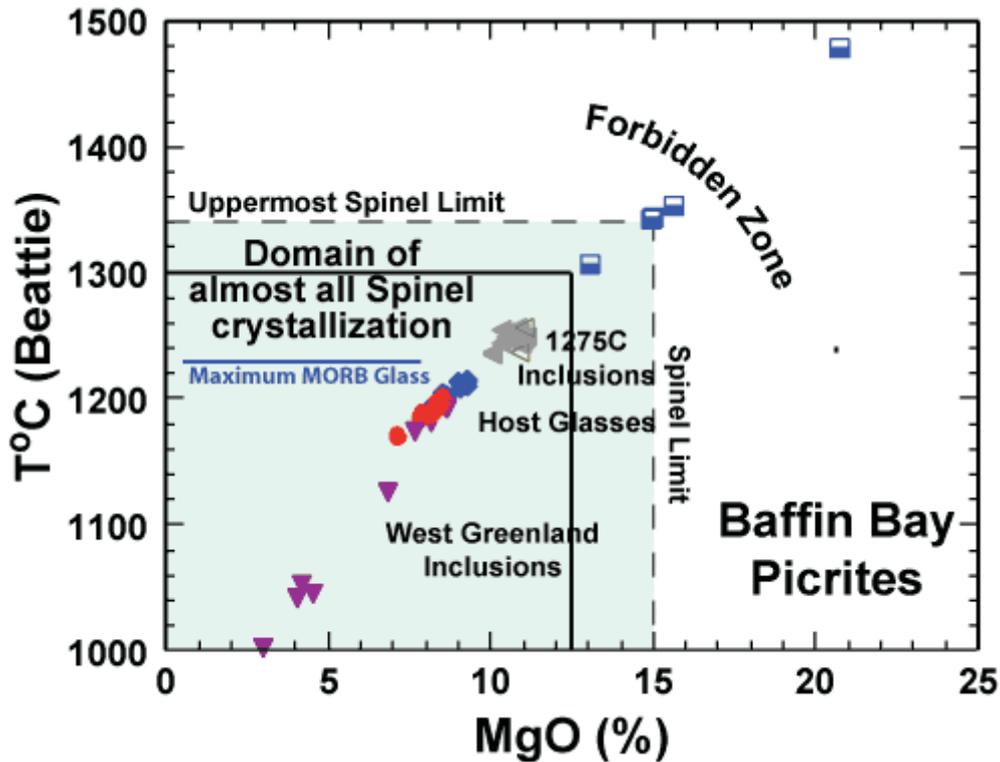


Figure 4: MgO content of natural glasses and inclusions from West Greenland and Padloping Island (data symbols and sources as in Figure 2) versus estimated temperature ( $T_c$ ) of crystallization (Beattie, 1993), calculated from Melt-Chromite (Poustovetov & Roeder, 2000). From Figure 3, the upper limits of T and MgO content, representing the most magnesian spinel composition, are given by the outer boundary of the light blue shaded field (Spinel Limit). Also from Figure 3, any more spinel in individual samples crystallized from hybrid and/or more differentiated liquids at  $<1300^\circ\text{C}$  and 12.5% MgO (dashed line), including a majority at temperatures and MgO contents represented by host glasses.

Spinel populations in given samples, like those of olivine, appear to reflect a diversity of parental compositions in their makeup (lateral spread in MgNo at relatively fixed CrNo in Figure 3A), some crystallizing from more magnesian melt strains, or more differentiated liquids, or both, than others. Host basalt is always an amalgam of many parental strains, and is also influenced by mixing with differentiated liquid. As a result, spinel compositions predicted from glass margins will always plot further to the right in Figure 3A than many natural spinel compositions in the same samples. One can force a host composition to some higher MgO content by adding olivine back into it, but this diagram indicates how artificial that procedure can be. Furthermore, such addition does not produce a realistic antecedent composition from which all olivine and spinel

co-precipitated along a common liquid line of descent, as *Putirka* (2008), for example, assumed in his calculation of a parent to Siqueiros picrite. The rocks say that this does not occur, and that at best only some parental magma strains anywhere had higher MgO contents than about 12.5%. In most rocks, the contributing melt strains, and certainly the average of them, had MgO content less than this.

Once again, the sample distribution obeys the Central Limit Theorem (*Anderson*, 2003). A consequence is that we should not “correct” a mean composition to what we consider to be an upper statistical limit. Perhaps also the upper limit has a plausible explanation, such as that the mantle source is lithologically heterogeneous (*Natland*, 1989, 2007; *Sobolev et al.*, 2005) or that the more magnesian inclusion compositions are a consequence of pressure (*Falloon et al.*, 2007a, b; *Presnall & Gudfinnsson*, 2008). Then, even the statistical limit should not be subject to correction. Thus a few strains in some places (e.g., Hawaii; *Clague et al.*, 1995), but not all (e.g., Samoa; [Natland, 2008](#)), might have been olivine controlled and had 13-16% MgO contents, but no higher, and they would have been no hotter if erupted individually than about 1340°C (Figure 4). So say spinel compositions.

## **Discussion and Conclusion**

One reviewer of an early draft of this webpage advised me to keep the technique of olivine addition to basaltic melt compositions as a means of evaluating parental MgO content, eruptive temperature and potential temperature. “Don’t throw the baby out with the bath water,” he advised. West Greenland and Padloping Island adjacent to Baffin Bay seemed to offer the most likely challenge to my view that picrites are complicated rocks; here, if anywhere, the procedure was bound to work. However, this region turns out to be no different than any other, and the rocks are hybrids that are indeed too complicated to see how the procedure can be valid even here. Some other procedures, outlined above, are even less appropriate. I have carefully searched the bath water, but find no baby there to throw.

To recapitulate, the Paleogene picrites of west Greenland and Baffin Bay were no hotter on eruption than picrite at Iceland, and eruptive temperatures at Iceland itself are little, at the very most about 100°, hotter than a normal spreading ridge (*Foulger et al.*, 2005; *Presnall & Gudfinnsson*, 2008). This probably rules out a plume from the core-mantle boundary by criteria of *Putirka* (2008), and may only require variations in pressure or mantle heterogeneity, rather than differences in temperature at a common source depth, to explain. Although other places need to be evaluated individually, this must be approached through a careful examination of crystallization histories of likely parental and near-parental compositions. To date, I have not found any rock from Samoa, Hawaii, Juan Fernandez, West Greenland, Skye or Iceland in which even any individual parental liquid magma strain exceeded 16% MgO content, and the average MgO content of hybrid and amalgamated host liquid was less than that. The search for olivine-controlled glasses should continue. If it is successful in future, it may modify the present picture, but we have not reached this stage yet.

The high Fo content of some Baffin and West Greenland olivine, reaching Fo<sub>92</sub> in five samples and as high as Fo<sub>92.9</sub> in one of them, places unusual and restrictive conditions on magma genesis. Fo<sub>92.93</sub> is more magnesian than average olivine in residual abyssal peridotite from transform faults, even those near presumed hot spots on ridge axes (*Dick*, 1989). Fo<sub>92.93</sub> is more magnesian than olivine found in Samoan residual harzburgite xenoliths, from which clinopyroxene has been exhausted by partial melting (*Wright*, 1987; *Dieu*, 1995). In fact, Fo<sub>92.93</sub> olivine is only found in the most refractory continental peridotite xenoliths from which almost all *orthopyroxene* has been extracted by melting and what is left in the melt source is mainly dunite (*Bernstein et al.*, 2006, 2007). In experimental analogs to peridotite considered with respect to partial melting, this is equivalent to the invariant point enstatite-out, when additional melt extraction ceases even as temperature increases (e.g., *Presnall*, 1969). In nature, Fo<sub>92.93</sub> thus

is likely the most magnesian olivine in any peridotite in the lithosphere, ancient or modern, but among peridotite xenoliths it *only* occurs, so far as we know, in Archaean (not Proterozoic or younger) mantle rocks so refractory that they may have been involved in a widespread or global magma ocean (*Bernstein et al.*, 2007). The extent of partial melting was likely 40% or more. Orthopyroxene exhaustion in mantle sources thus is the minimum partial melting condition required to produce a basaltic or picritic partial melt capable of crystallizing  $>Fo_{92}$  olivine.

This is an extraordinary condition to impose on a modern mantle plume, yet here it is postulated at a place where  $Fo_{92-93}$  olivine happens to be prevalent in the particular ancient lithospheric mantle through which basaltic magma had to pass (*Bernstein et al.*, 2006). Backtracking liquid compositions to the extent that they should be in equilibrium with  $Fo_{92-93}$  thus implies that the Baffin plume head was as hot as anything that has existed on the surface of the Earth through geological time, that it ignored the upper mantle through which its magmas had to ascend, entraining none of it, and yet that it produced olivine phenocrysts with the same composition as in only that type of mantle. Furthermore that olivine now survives only in variably enriched differentiated basalt so little different from transitional to enriched abyssal tholeiite or Icelandic basalt that even picritic calculated parental compositions are only about 21% partial melts (*Putirka*, 2007) rather than the 40% partial melts entailed by the petrological condition. This seems unlikely.

Finally, the high values of  $^3He/^4He$  (R/Ra) in gases taken from olivine in several West Greenland samples (*Graham et al.*, 1998) must be viewed with caution. Bulk-rock compositions were not determined. Rock descriptions in *Graham et al.* (1998) indicate that olivine in some of the samples was as magnesian as  $Fo_{92}$ , although whether this was a particular host to the high- $^3He/^4He$  cannot be known. Even if such radiogenic helium is present in less forsteritic olivine, spread of "mantle" helium to highly differentiated liquids in flanking rift systems, there to work its way into olivine as iron rich as  $Fo_{75}$ , must be allowed (*Natland*, 2003). In this case, the rocks have low concentrations of helium in general, which is consistent with their olivine being picked up when shallow degassed differentiates in dikes traversed crystal-rich sludges or mantle wall rock. It also fits with the evidence for shallow crystallization of spinel in other rocks. The isotopic signal of helium thus could well be dominated even by a minor concentration of gases cracked out of ancient olivine-rich materials *en route*. Thus if helium is present in very forsteritic olivine in a picrite, then the source of that olivine, especially if it has kink bands and subgrains, must be carefully evaluated. The olivine is almost certainly xenocrystic.

Olivine is a natural time capsule for "old" helium with high  $^3He/^4He$ , and in cumulates such as dunite, which are often deformed, helium isotopes could retain original values without the influence of U and Th, excluded from olivine, in the same facies, for eons. One might simply claim that such helium arises with plume magmas (*e.g.*, *Graham et al.*, 1998). However, this follows from the assumptions that the mantle is homogenous, and that all olivine in a sample lies along a common liquid line of descent. But when refractory Archaean mantle is traversed by later basalt, one has to wonder whether a plume attribution for this particular olivine, thence helium isotopes, is justified.

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