

West Greenland Glass in Picrite – Low-pressure Cotectic Differentiates or Independent Compositions?

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Introduction

I reply to "[West Greenland was hot—how else can you make 22,000 km³ of picrites?](#)" [Larsen & Pedersen \(2009\)](#). The data set of [Larsen & Pedersen \(2000\)](#) is superb, and more closely bounds the problem of temperature than any other I have seen. I have enjoyed working with it. I welcome their comments on [Natland \(2008\)](#).

Title

The title of [Larsen & Pedersen \(2009\)](#) speaks to the tug that picrites have on the petrological conscience, extant at least since [O'Hara \(1968\)](#), but perhaps even before ([Drever & Johnston, 1957](#)), namely that all picrites represent great heat and high temperature. This idea today is rather entrenched. However, [Clague et al. \(1995\)](#) proposed that most picritic basalt from Kilauea volcano and offshore Puna Ridge resulted from mixing between crystal-rich sludges with abundant accumulated olivine and strongly differentiated liquids that had evolved by crystallization of plagioclase, clinopyroxene and olivine along a low-pressure cotectic. The picrites, then, erupted at temperatures of host liquids (glasses), far along a multi-phase liquid line of descent, and [Clague et al. \(1995\)](#) calculated these temperatures using geothermometry. I advocate this mechanism for Baffin Bay and West Greenland for exactly the same reasons. The disposition of melt and magma chambers was considered by [Clague et al. \(1995\)](#) and it is no different, no less complex a problem, than it is now; any solution to it must be contrived.

Mountains of picrite notwithstanding, the problem of the temperature of onset of crystallization of olivine phenocrysts requires precise consideration of the compositions of glass and minerals. What if deduced temperatures turn out to be low in all of them? There is plenty of low-temperature picrite at Hawaii, and abundant dunite xenoliths represent crystallization of a large volume of olivine-saturated liquids near the bases of volcanoes ([Clague & Denlinger, 1994](#)). The olivine in these crystallized at <1200°C, although olivine crystallized at higher temperature may now be re-equilibrated. [Clague et al. \(1995\)](#) was a serious warning shot across the bow of traditional interpretations. I have simply applied their approach to Baffin Bay and West Greenland.

Associated Silicic Rocks

Noting the presence of interbedded silicic tuffs and lavas was primarily a caution based on a prior interpretation of Iceland ([Natland, 2007](#)), except that for Baffin Bay and West Greenland, a geochemical signal of silicic contamination is more likely. The setting is intra-cratonic rather than intra-oceanic. If the Vaigat formation has no interbedded silicic lavas, but the overlying Maligât formation does, then the possibility remains that even lavas of the Vaigat formation might have experienced a small amount of assimilation of silicic continental crust, or mixing with rhyolitic partial melts derived from that crust. The situation recalls that of Theistareykir Volcano in Iceland, which was selected by

MacLennan et al. (2003) for study partly because of absence of interbedded silicic lavas. Theistareykir nevertheless carries a small signal of silicic contamination, which I modeled as basalt-rhyolite mixing (*Natland*, 2007). The presence of a small silicic contaminant in lava of the Vaigat formation is, however, not relevant to the question of temperatures.

ΔNb

Low ΔNb (*Fitton et al.*, 1997) in West Greenland basalt is a surprise, but again this has nothing to do with temperature. My comment was directed mainly to the many who prefer consistency of geochemistry in their geodynamic modeling, and who argue that high ΔNb necessarily supports presence of an enriched mantle source. Iceland is the standard of comparison, the type locality, if you will, of high- ΔNb . The surprise was discovering that the “plume-head” picrites of West Greenland, representing the onset of the North Atlantic Igneous Province in the plume hypothesis, and Iceland itself, the “plume tail”, are different.

Whereas *Fitton et al.* (1997) and more recently *Starkey et al.* (2009) see shifts in ΔNb as indicative of mantle source properties in this region, I believe instead that it is *mainly* an indicator of silicic contamination (*Natland*, 2007). That is, assimilation of silicic material or basalt-rhyolite mixing can explain interbedded basalts like these that have both positive and negative ΔNb (*Natland*, 2007). Without assimilation of, or mixing with, rhyolite I argued that Icelandic basalt, all of which *Fitton et al.* (1997) define as having $\Delta Nb > 0$, would more closely resemble mid-ocean-ridge basalt (MORB; $\Delta Nb < 0$). Accepting that an underlying geochemical continuity should exist across the North Atlantic Igneous Province, and noting that assimilation and mixing are both well documented in the field and by geochemistry, West Greenland supports my point. ΔNb , then, is non-diagnostic or, more bluntly, useless as a source discriminant especially in continental rift settings. I explained correlations of ΔNb with, e.g., light-rare-earth enrichment ($[La/Sm]_N$) and isotopes of Sr and Nd at Iceland by the same mechanism (*Natland*, 2007).

Crystallization along a Low-Pressure Cotectic

The fact remains that no basalt glass from Baffin Bay and West Greenland (nor from Iceland) so far analyzed exhibits olivine control. This has to be a serious drawback for backtrack calculations, undercutting its cardinal assumption. As first emphasized by *Wright & Fiske* (1971) for Kilauea, olivine control is only demonstrated when its consequences are evident on two-oxide variation diagrams for ALL oxides determined by chemical analysis of major elements. Failure of demonstration on any single diagram, for example TiO_2 -MgO, precludes it. [Natland \(2008\)](#) provided more than one variation diagram showing lack of olivine control among basaltic glass margins to picrites of Baffin Bay and West Greenland. A single diagram might coincidentally have data points aligned along some postulated olivine-control line, but other diagrams could, and in this case do, rule it out.

An important step in evaluation of variation diagrams is to show single-mineral or multiple-mineral control of crystallization differentiation; then the rocks belong to the same liquid line of descent and have a common parent. However, a sequence of basalt, perhaps sampled over a short time from eruptions of the same volcano (e.g., Kilauea) usually shows subtle parental variations. These will produce some dispersion on variation diagrams, and different places may contrast sharply (e.g., *Wright*, 1971), although general underlying shallow differentiation control can usually be seen. Glass from picrites of Padloping Island, Baffin Bay, and West Greenland (BWG glass) illustrates this nicely.

BWG glass exhibits MORB-like cotectic variation and shows no sign of olivine control. For example, both TiO_2 and total iron as FeOT increase as MgO decreases from 9.2 to 6.7%; in the same course of differentiation, both CaO and Al_2O_3 decrease. These variations cannot be accomplished simultaneously without plagioclase on the liquidus.

This was sufficiently shown on the diagrams in [Natland \(2008\)](#). Olivine control has only been demonstrated in extrusive glasses from Hawaii's Kilauea volcano and Puna Ridge ([Clague et al., 1995](#)); variation diagrams for those glasses have a sharp inflection where olivine-controlled differentiation joins the low-pressure cotectic at about 7% MgO content (Figure 1A). Figure 1B shows that BWG glasses have no such inflection, and that their overall trend is similar to trends seen in Indian Ocean MORB (Figure 1C). Although Indian MORB have fairly wide CaO variation at given MgO, they can be broken down into three types based on differences in parental Na₂O (Na₈ = Na₂O at 8% MgO content; [Klein & Langmuir, 1987](#)), which have different characteristic phenocryst assemblages in glass with highest MgO contents ([Natland, 1991](#)); viz:

- Type 1 = plagioclase, clinopyroxene and olivine;
- Type 2 = plagioclase and olivine;
- Type 3 = olivine alone.

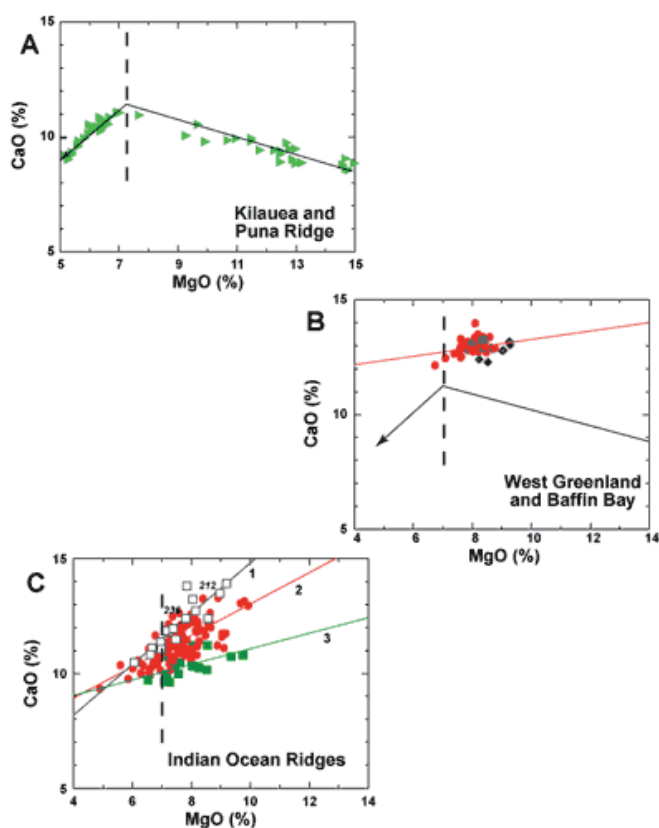


Figure 1: MgO versus CaO for basaltic glass from A. Kilauea and Puna Ridge ([Clague et al., 1995](#)); B. West Greenland ([Pedersen & Larsen, 2000](#)) and Padloping Island, Baffin Bay ([Francis, 1985](#); [Robillard et al., 1992](#)); and C. Central and Southwest Indian Ridges, and the Red Sea (Lamont-Doherty Petrology Data Base, *PetDb*). Note slightly different X-axis coordinates in A. In A, the bold line with an arrow is hand-drawn to show olivine and cotectic crystallization, with a break in slope at 7% MgO (vertical dashed line, repeated in B and C). In B, blue West Greenland glasses were used by [Larsen & Pedersen \(2000\)](#) for their backtrack calculations; red are all other glasses. Padloping Island glasses are dark blue diamonds. The red line is a linear regression through all data. In C, open squares are Indian Ocean Type 1 MORB; red dots = Type 2; green squares = Type 3. Lines are linear regressions through each type.

All three types carry Cr-spinel. All three have trends in Figure 1C *requiring* cotectic crystallization of mineral assemblages dominated by plagioclase, in order for CaO to decrease as MgO decreases, consistent with the phenocryst assemblages. From Figure 1B, BWG glasses most resemble Indian Ocean Type 1 MORB, overlapping Type 2. Note especially that glass-bearing basalt from DSDP Sites 212 and 236, labeled in Figure 1B, carries *abundant* plagioclase and clinopyroxene phenocrysts (Natland, 1991).

More generally, no MORB in the Indian Ocean is picritic, even if the glass has MgO > 9% and Mg# (= $100 * \text{Mg}/[\text{Mg} + \text{Fe}^{2+}]$) > 70. The three types have a general tectonic association: Type 1 occur mainly in what might be called early rift settings such as the Red Sea and locations near the Seychelles continental fragment, Type 2 are widely distributed along the modern Carlsberg and Central Indian Ridges, and Type 3 occurs mainly along the Southwest Indian Ridge near the Indian Ocean triple junction where spreading rates are very slow. BWG glass, then, most strongly resembles early-rift Indian-Ocean MORB, but verging on more typical Indian MORB compositions.

Most picrites also have negative Eu anomalies (West Greenland -0.53 ± 0.26 – 14 samples; Padloping Island -0.34 ± 0.27 – 36 samples; Starkey *et al.*, 2009). This can only happen with fractionation of plagioclase, as in the MORB suite. Assuming the anomalies indicate matrix proportions of rare-earth elements, similar to glass, and comparing to MORB with similar Eu anomalies, the rock groundmasses have ~7-8% MgO contents, like those of quenched BWG glass. Plagioclase must have been on their liquidus, consistent with reductions in CaO and Al_2O_3 as MgO content decreased along a cotectic, and prior to incorporation of olivine phenocrysts.

Since BWG glass follows a low-pressure multi-phase liquid line of descent similar to MORB, we can calculate the usual parental parameters for MORB (Klein & Langmuir, 1987). Figure 2 shows how extremely similar they are to MORB when comparing Si_8 , Ti_8 and Fe_8 to Na_8 . BWG glass partially or largely overlaps Types 1 and 2 Indian Ocean MORB on all diagrams, especially glass from Padloping Island. West Greenland glass has generally lower Si_8 and somewhat higher Ti_8 than the most nearly comparable Indian MORB. In these respects, BWG glass is also very similar to basalt glass from Iceland, with West Greenland glass predominantly corresponding to higher- TiO_2 Iceland basalt (defined as having $\text{Ti}_8 > 1.4\%$; Natland, 2007), and most glass from Padloping Island resembling low- TiO_2 Iceland basalt. Picritic or not, Baffin Bay and West Greenland basalt glass is no different from that of moderately differentiated Icelandic basalt, and neither is very different from early-rift Indian Ocean MORB.

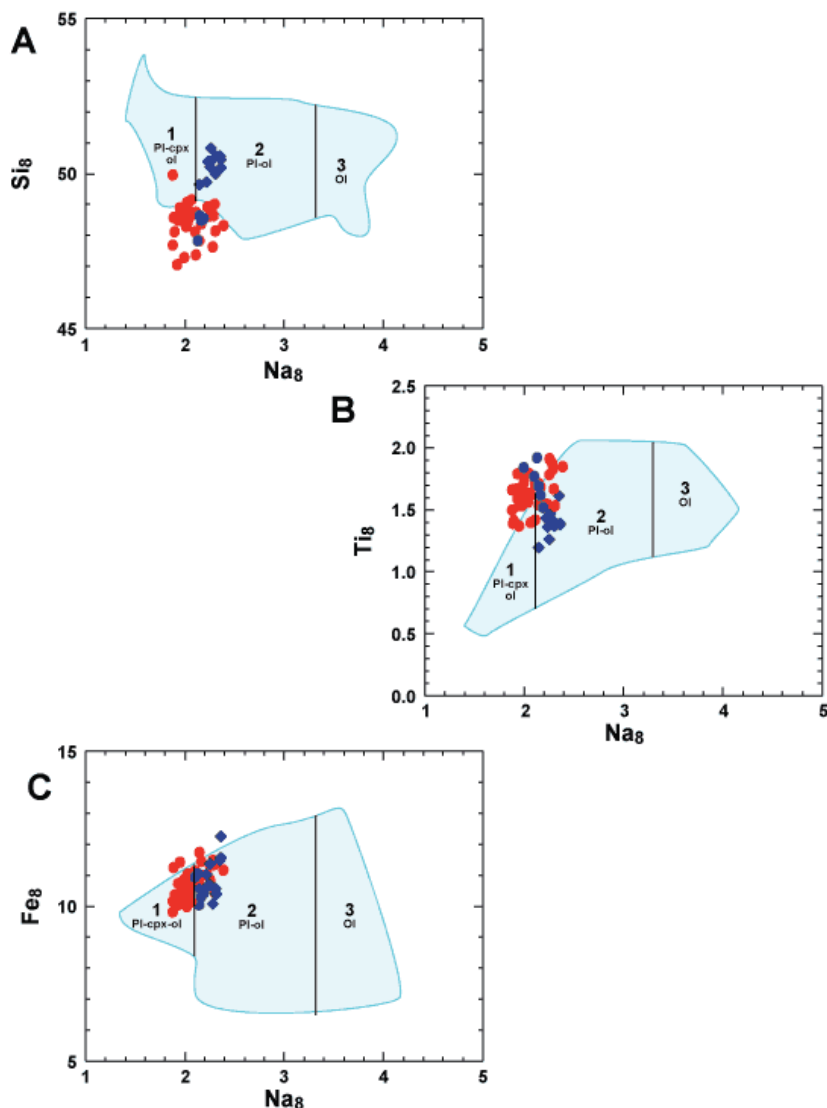


Figure 2: Parental attributes of glass from West Greenland and Padloping Island compared with data fields for the Indian Ocean MORB array. Na_8 versus A. Si_8 ; B. Ti_8 and C. Fe_8 were calculated according to Klein & Langmuir (1987). Symbols and data sources are as in A. Indian Ocean MORB are subdivided into Types 1, 2 and 3 (cf., Figure 1C) at values of Na_8 given by the vertical lines (Natland, 1991).

Figure 3 shows diagrams of CaO versus Al_2O_3 for the same three data sets, except blown up in comparison to Figure 4 of [Larsen & Pedersen \(2009\)](#). The diagrams are to the same scale, but have different values along the abscissas and ordinates. In Figure 3A, for Kilauea and Puna Ridge, the inflection between olivine-controlled and cotectic crystallization is very sharp. No such inflection occurs either for BWG glass (Figure 3B) or Indian MORB. Again, BWG glass resembles fairly primitive to moderately differentiated Type 1 Indian MORB (light blue data field), including strongly plagioclase-phyric basalt from DSDP 212 and 236. Whereas the data cluster for BWG glass is compact and without a strong trend, it cannot be olivine-controlled (thin arrow). Instead plagioclase and clinopyroxene must combine to produce trends like those shown by the bold arrows, which go in the opposite direction and are oblique to an olivine-control line drawn through the data. Vertical dispersion of data points in Figure 3B results mainly from parental diversity, glass from Padloping Island having higher Al_2O_3 than most glass from West

Greenland. This simply means that although all glasses are cotectic compositions, they do not fall along the same cotectic liquid line of descent. They do not have a common parent.

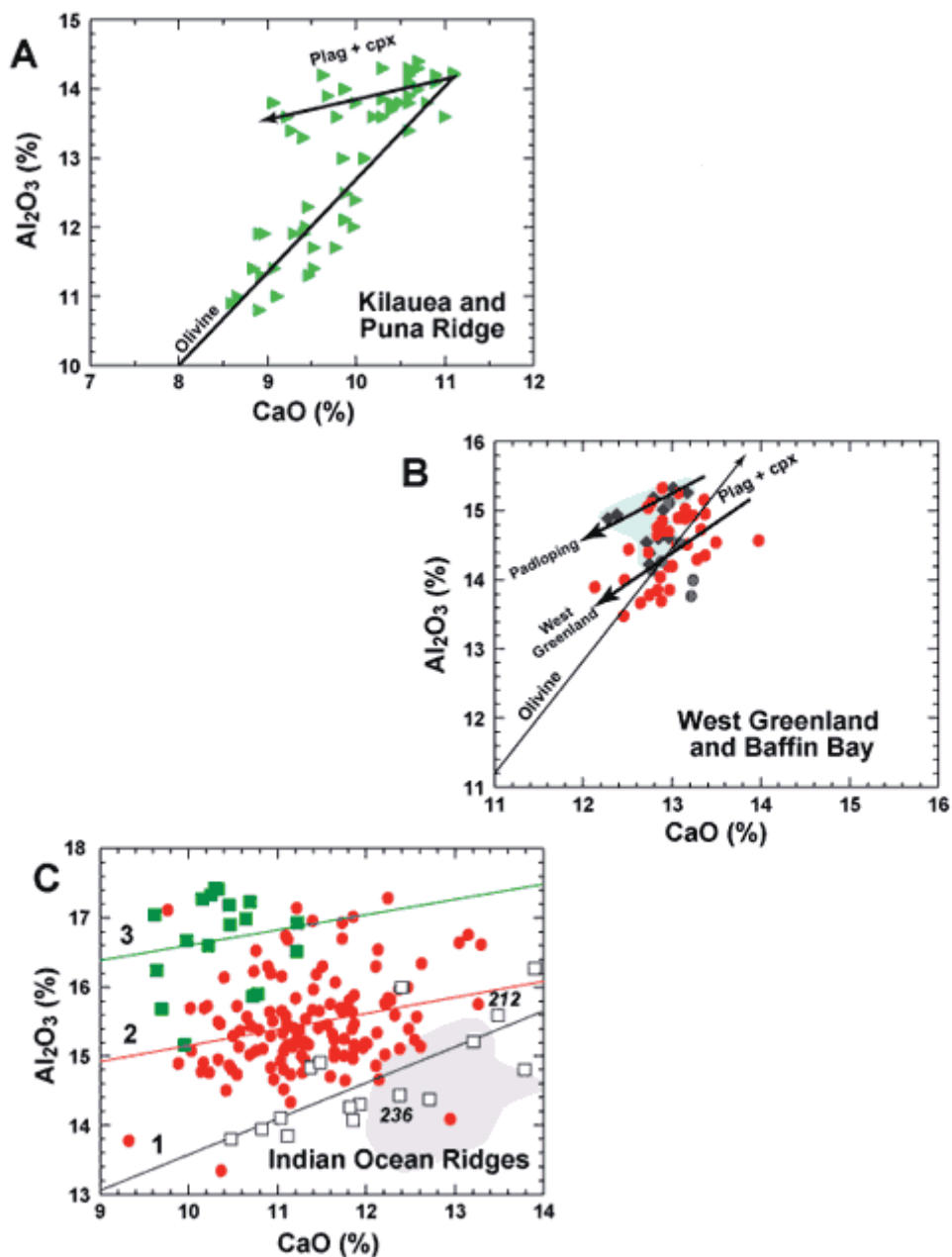


Figure 3: The contrast between olivine-controlled and cotectic glass compositions: CaO versus Al₂O₃ for basalt glass from A. Kilauea and Puna Ridge; B. West Greenland and Padloping Island, Baffin Bay; and C. Indian Ocean Ridges. Symbols and data sources are as in Figure 1. Lines with arrows in A and B are generalized hand-drawn trends (see text for discussion). In C lines are linear regressions through each data set for Types 1, 2 and 3 Indian MORB. Note that although the diagrams are to the same scale, values of abscissas and ordinates are not the same. In C, the area for Baffin-West Greenland glasses in B is shown as a light purple field. Data points for two sites of the Deep Sea Drilling Project, DSDP 212 and 236, are labeled.

From these considerations, I reiterate that analyzed glass from Baffin Bay and West Greenland is unsuitable for backtrack calculations to estimate temperature. They are cotectic compositions. The further out along the cotectic trend a glass is in terms of iron enrichment, the greater will be the tendency to overestimate both crystallization temperatures of any olivine in the glass, and then potential temperature.

The Larsen & Pedersen (2009) Alternative

[Larsen & Pedersen \(2009\)](#) suggest that every analyzed glass of the Vaigat Formation is *precisely* at the end of olivine control—just at the point where clinopyroxene, plagioclase, or both join the liquidus, and no further along it. In other words, each is an independent composition upon eruption, derived from some high-MgO liquid that *intrinsically* had different TiO₂, total FeO, CaO, Al₂O₃, and Na₂O contents—all the oxides conventionally used on variation diagrams to reveal shallow cotectic crystallization, as I have just done. That is, the contents of these oxides varied *at the source* in such a way as to mimic shallow cotectic crystallization after all olivine had separated out polybarically upon ascent from the mantle, but yet which reached proportions of 20-30% accumulation in the same hosts.

I find it implausible that no cotectic variation occurred, that every BWG glass formed uniquely from an independent picritic parental composition almost as we see it, and that only some olivine needs to be added back to it to have it in its original state. Even less plausibly, the glass compositions now are precisely such as to fool us into thinking that they exhibit a shallow, MORB-like, plagioclase-dominated variation *complete with negative Eu-anomalies*.

Now variation diagrams are a powerful means of getting at differences in parental compositions at the source, by sorting out the effects of shallow crystallization. I claim to see shallow effects in BWG glass whereas [Larsen & Pedersen \(2009\)](#) deny them. To justify their temperature backtrack calculations, they pass over the manner in which petrologists have treated obvious trends on such diagrams for more than 40 years. I simply say that whether or not any antecedent magma crystallized at high temperature, the host glasses cannot be used for backtrack calculations. They are, by the principle of least astonishment, cotectic liquid compositions. Two questions remain, how do you get at temperature, and where is the plagioclase?

Melt Inclusions

[Larsen & Pedersen \(2009\)](#) and I agree that melt inclusions are too complicated to use for backtrack calculations. However, to the extent that those with > 6% total iron as FeO roughly mimic low-temperature cotectic crystallization, *and* occur in forsteritic olivine, they might give clues about temperature. I pointed out that all of them from West Greenland and most of them from Baffin Bay have lower iron than host glasses to olivine. Then, on the assumption that the inclusions are olivine-controlled liquids, backtrack estimates *suggest* lower temperatures than inferred from host glasses. The assumption of olivine control is unproven, but the general point remains that low-iron melt stems contributed to the overall composition of the final aggregated picrite, and olivine as forsteritic as Fo_{92.8} contains these inclusions. Low iron means low temperature in backtrack calculations. The inclusions confirm that the particular olivine that holds them did not crystallize along a common liquid line of descent leading to the current host. The relevant liquid seen as an inclusion was demonstrably different.

Magma Mixing

Figure 4 shows spinel compositions for several West Greenland basalts of the Vaigat Formation ([Larsen & Pedersen, 2000](#), supplemental data). The data may be viewed in aggregate, or alternatively that they reveal populations of spinel with different

compositions in the same rocks, and thus are evidence for mixing between primitive and differentiated liquids. One population, Type A, seen in three of the samples shown, has high Cr# and high Mg#, and is occluded within forsteritic olivine. ($>Fo_{90}$). Another population in the same three samples, Type B, has both lower Cr# and Mg# and occurs within less forsteritic olivine. It resembles spinel from Indian Ocean Type 1 and 2 basalt (red boundary). Rare spinel with high TiO_2 , and falling outside both populations in Figure 3, occurs in two samples and crystallized in strongly differentiated liquids.

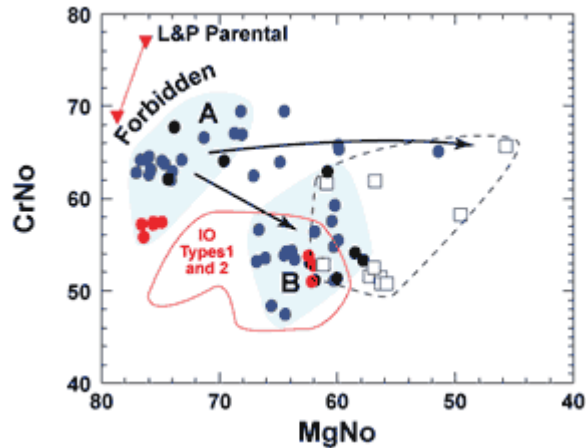


Figure 4: A portion of the standard diagram of Mg# versus Cr# for Cr-spinel in four samples from West Greenland (Larsen & Pedersen, 2000), to show that spinel compositions comprise two principal clusters of data. Samples are Vaigat Formation 136943, blue dots; 332771, black dots; 362148 red dots; and 326789, open squares. Blue backgrounds indicate Type A, spinel with high Cr# and Mg# in forsteritic ($>Fo_{90}$) olivine, and Type B, spinel with lower Cr# and Mg# in more iron-rich olivine. Sample 326789 (dashed boundary) lacks Type A spinel, and includes some that have both lower Mg# (higher iron) and higher TiO_2 than Type B spinel. Lines with arrows indicate possible trends for magma mixing. Type B spinel are similar to spinel in Types 1 and 2 Indian MORB (red boundary; J. Natland, unpublished data). Also shown are two spinel compositions (inverted brown triangles) calculated using Poustovetov & Roeder (2000) from a high-MgO parental composition derived by backtracking (Larsen & Pedersen, 2000). These have higher Cr# and Mg# than any Type A spinel, including spinel within the olivine used for the backtrack calculation.

One sample lacks forsteritic olivine, all its spinel is Type B and it occurs in olivine as iron rich as Fo_{84} . This is fully consistent with the host liquid lying along a low-pressure cotectic, as also indicated by the similarity between Type B spinel compositions and those predicted from host glass (Natland, 2008) using the procedure of Poustovetov & Roeder (2000). This demonstrates that the host liquids were along a low-pressure cotectic, and that most spinel in the samples (Type B) crystallized from melts like the host liquids.

I suggest now that mixing with something containing very primitive olivine and spinel (Type A) and primitive affiliated melt, also occurred. Olivine-rich crystal sludges were incorporated into differentiated basaltic liquid. Preliminary calculations for Padloping Island based on variation diagrams using new data (Starkey *et al.*, 2009), suggest that the average composition of the primitive melt there had about 0.8% TiO_2 , 8% FeO_T and 10.5% MgO (which is in equilibrium with $\sim Fo_{90}$ olivine), consistent with compositions of melt inclusions in Fo_{88-92} olivine. The liquid represented about 50% of a mixture that otherwise contained olivine of average composition $Fo_{87.8}$ (240 analyses from Larsen & Pedersen, 2000, Supplemental Data; 15 analyses from Francis, 1985). Somewhat more magnesian precursors to this average primitive liquid precipitated olivine with $>Fo_{90}$ compositions. Picritic whole-rock analyses thus do not strictly define simple olivine

addition control trends; mixing involved olivine and *two* liquids, as originally suggested for Kilauea by *Clague et al.* (1995). This by itself precludes using picrite FeO-MgO variation as the basis for an incremental olivine-addition backtrack procedure to construct olivine fractionation curves and estimate parental compositions and temperatures. The present host glasses were also not precisely cotectic liquids, but hybrids between cotectic liquids and more primitive liquids; the hybrids thus lie within the olivine primary phase volume and no longer have plagioclase on the liquidus.

The sample with only Type B spinel is an example of a differentiated picrite of the kind common in Hawaiian, Juan Fernandez and Samoan basalt (*Natland, 2003; 2008*). Such picrite belies the notion that even a full mountain of picrite necessarily represents high temperature eruption in its totality. Of the 20,000 km³ of Vaigat Formation picrite, how much of it is like this?

Primitive Olivine and Spinel

Since the backtrack-technique is questionable, we must focus on the origin of primitive spinel and coexisting forsteritic olivine to try to understand temperature. I suggested (*Natland, 2008*) that very forsteritic olivine could represent something that crystallized very long ago, and which might be present as a reconstituted xenocryst, perhaps originally in an ultramafic cumulate, or as part of peridotite restite left over from formation of Archaean continental crust. There is no doubt that the latter material, with appropriately magnesian olivine, is present in the region (*Bernstein et al., 2006, 2007*). I noted the suggestion of prior workers (especially *Francis, 1985*) that the most forsteritic olivine at Padloping Island (Baffin Bay) is kink-banded or it has subgrains, thus is xenocrystic.

In the Vaigat formation, however, olivine with composition $>Fo_{90}$ has magmatic compositions (with small but significant amounts of CaO and Cr_2O_3), has glass inclusions, no subgrains, and is not kink-banded (*Larsen & Pedersen, 2009*). This is all to the good. Thus we may use compositions of coexisting olivine and spinel (*Larsen & Pedersen, 2000* Supplemental data) to understand the compositions of the liquids that produced them. The compositions of parental liquids estimated by *Larsen & Pedersen (2000)* predict spinel compositions with both higher Cr# and Mg# than any spinel they analyzed. *Spinel in the very olivine used in the calculation is not as magnesian as predicted.* My estimate from spinel was that both the olivine and the spinel crystallized from somewhat less magnesian liquids, with ~15% MgO rather than ~20% MgO. This makes a difference of about 150° less in crystallization temperature, and it mainly means that the liquids chosen for backtrack calculations have too much iron, as they should, since they are along a low-temperature cotectic.

In Figure 4, the difference between the locations of most primitive spinel (highest Cr# and Mg#) and spinel predicted from postulated parental compositions represents the “forbidden zone” in Figure 4 of *Natland (2008)*. It is not an interval of “intratelluric crystallization”, as suggested by *Larsen & Pedersen (2009)*, for the reason just given. The gap simply means that the glass used in the computation had too much iron, as indicated above, thus higher MgO content is needed to get a liquid into equilibrium with the forsteritic olivine in the sample. Lower the initial iron, as indicated by spinel and the melt inclusions, and then even assuming the propriety of the backtrack procedure, which I do not, the estimated parental MgO drops to about 15%. This still gives fairly high temperatures (~1350°C), but nowhere near the 1500-1550°C of *Pedersen & Larsen (2000)*.

Xenocrysts?

Are mantle xenocrysts represented in Vaigat Formation picrites? I remain struck by the coincidence that the glasses are MORB-like and yet neither open-ocean MORB nor abyssal peridotite have olivine with composition $>Fo_{91}$. No MORB glass is olivine

controlled (*Presnall & Gudfinnsson, 2008*). Something very specific to Baffin Bay and West Greenland produced such olivine in an otherwise ocean-crust setting. A plume head bringing each and every geochemical variation from depth is kind of a magical solution, but like any *deux ex machina*, not a very satisfactory explanation. I think the actual mechanism was more prosaic.

Thus olivine of composition Fo_{91-93} does occur in harzburgite-dunite xenoliths of Archaean lithosphere in this region. Is it unreasonable to suppose that such olivine, present here but not beneath spreading ridges, is in some way the source of very forsteritic olivine in West Greenland picrite? Unfortunately, that olivine has low CaO. I offer three additional possibilities to account for the higher CaO in BWG olivine phenocrysts, forewarning at the outset that all are contrived. The three might act in tandem.

First, from experiments, olivine is known to re-equilibrate rapidly with melt in which it is immersed (*Gaetani & Watson, 2000*). Although those experiments mainly dealt with effects of re-equilibration on FeO, effects on other elements cannot be ruled out. Here is an example. My own data on compositions of olivine in Samoan xenoliths and phenocrysts show quite variable and overlapping CaO contents from 0.03-0.51% in residual harzburgite (Type 1; *Frey & Prinz, 1978*) and magmatic dunite-pyroxenite (Type 2) ultramafic xenoliths. CaO in most olivine ($<Fo_{90}$) in the dunite-pyroxenite suite precisely matches that seen in West Greenland olivine of composition $>Fo_{90}$ (0.25-0.45%), but even olivine in some residual harzburgite has high CaO. Were some of the Samoan olivine crystals in xenoliths re-equilibrated at their source by percolating magma?

Second, could the Archaean upper lithosphere in this region contain old ultramafic cumulates having Fo_{91-93} olivine, precipitated perhaps from komatiitic magma? Presumably that olivine originally had magmatic concentrations of CaO. Would that have been lost upon re-equilibration to "mantle conditions", or did it remain the way it usually does in the ultramafic portions of ancient layered igneous intrusions?

Thus forsteritic olivine in West Greenland picrites may have begun as xenocrysts originally precipitated from ancient ultramafic magmas and retaining the properties of magmatic minerals. The sources were in shallow and relatively (over billions of years) cold sub-continental lithosphere that was broken and fragmented during the early stages of continental rifting, and traversed by MORB.

Finally, what would be the course of differentiation of primitive MORB magma injected into and reacting with very refractory peridotite encountered during continental rifting? MORB liquid reacts with abyssal peridotite by precipitating olivine as dunite and extracting pyroxene (*e.g., Dick & Natland, 1996*). In general, provided the mass of transported liquid remains high, (i.e., it is constantly resupplied and doesn't diminish in volume during the course of differentiation), the peridotite buffers the Mg# of the liquid, and pulls it away from cotectic precipitation of olivine and plagioclase. Concentrations of incompatible oxides and trace elements (TiO_2 , Zr, etc., but perhaps also Cr, depending on oxidation state?) will increase depending on the volume of mantle that reacted with the magma. Primitive plagioclase-saturated Type 1 or Type 2 Indian MORB reacting with Archaean harzburgite, then, might shift to a composition in an olivine primary phase volume, in which case forsteritic olivine and associated Cr-spinel, but no other phases, would crystallize. The serious questions are whether olivine phenocrysts of composition Fo_{91-93} and affiliated spinel with both high Cr# and Mg# could form by such a mechanism, and what would be the pressure and temperature of their crystallization?

The key to evaluating this is very careful petrographic and mineral study. It is not enough to state that the most magnesian olivine in West Greenland picrite lacks subgrains or kink bands, and is euhedral or skeletal, thus is obviously magmatic. The question turns specifically on the morphology and composition of crystals with composition Fo_{91-93} .

and the manner and composition of their intergrowths of spinel and melt-inclusions. Are the olivine grains broken? Are there internal grain boundaries? Do they have euhedral overgrowths surrounding irregularly shaped cores? Are the melt inclusions in embayments in outer zones of the mineral grains (zoning itself is well documented by *Larsen & Pedersen, 2000*), or do they extend into their cores? Are the embayments skeletal in origin, and formed at elevated undercooling, or did they originate by resorption? For all the data that exist on these rocks, these questions have not yet been answered. If they can be answered, then all questions about backtrack techniques will be moot.

Cr₂O₃ in Olivine

Cr₂O₃ in olivine is not routinely measured to high precision; *Larsen & Pedersen (2000, 2009)* show its potential utility. At this point, I will simply say that temperatures estimated for olivine using this parameter (*Li et al., 1995*) can only coincidentally correspond to the backtrack estimates of *Larsen & Pedersen (2000)*; those I consider to be unreliable for the reasons I have emphasized above.

Li et al. (1995) performed experiments in the four-component systems, MgO-SiO₂-Cr-O, most notably without FeO. The temperature isopleths they calculated were for this system only; the general tendency of additional components to lower equilibrium temperatures remains a possibility. At that, Figure 2 of *Larsen & Pedersen (2009)*, which shows Fo (%) vs. Cr₂O₃ in olivine, does not formally show higher Cr content in West Greenland olivine than anywhere else, although *Li et al. (1995)* summarize scant information for MORB olivine, saying that it has < 0.02% Cr₂O₃. The general conclusion of *Larsen & Pedersen (2009)*, that high Cr₂O₃ in West Greenland olivine indicates high temperature, may be correct, but it needs to be qualified with better understanding of how natural systems function, more explicit comparisons to hypothetical lower-temperature places like spreading ridges, and a question about how much hotter than usual West Greenland olivine equilibration temperatures actually had to be, in order to get the elevated Cr₂O₃ concentrations in olivine. In short, what uncertainty is there to isopleths in Figure 3 of *Larsen & Pedersen (2009)*?

A Suggestion

Figure 5 presents a generalized sketch showing possible schematic relationships between a Paleocene magma chamber emplaced in high-level peridotite but below granite at West Greenland. The difference in elevation might represent the difference between the floor of a rift valley lake, and a volcano emplaced at high levels on a structural accommodation zone. It is based on geometrical models proposed to explain why picrite occurs in structural depressions such as Siqueiros fracture zone (*Natland, 1989*) and in deep water at the tip of Kilauea's Puna Ridge (*Clague et al., 1995*), whereas differentiated lava predominantly occurs higher.

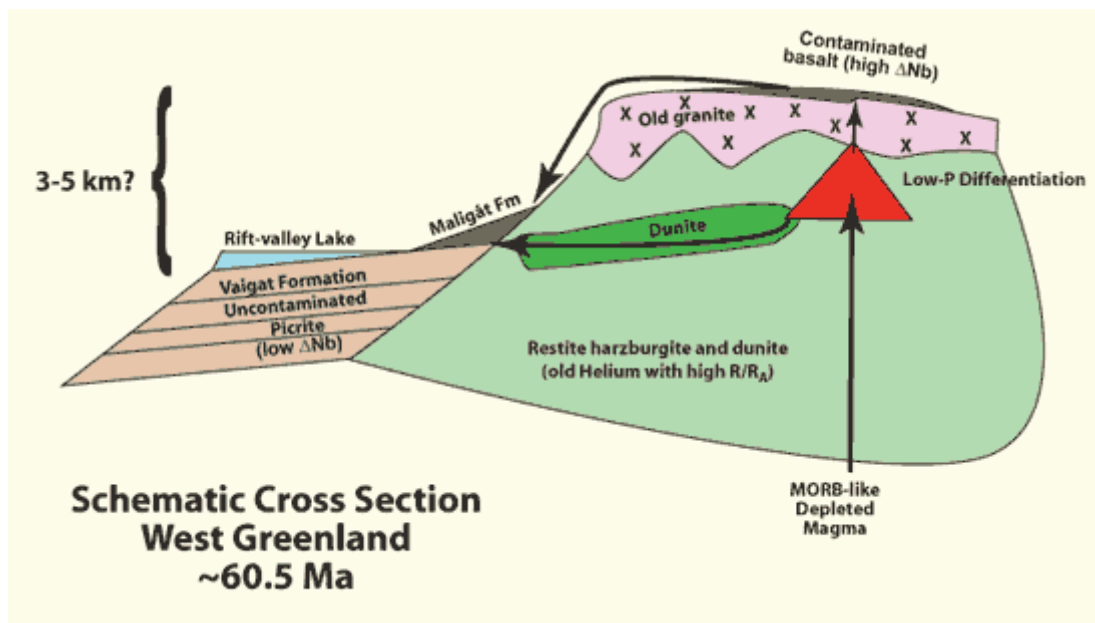


Figure 5: Schematic cross section through the rifted West Greenland margin during the Paleocene (~60.5 Ma Storey et al., 2004). See text for discussion.

At West Greenland, I suggest that basaltic differentiation reached and proceeded along the low-temperature multi-phase cotectic in the magma chamber (red). A body of dunite (dark green) is shown to lie along a curving listric fault extending into the adjacent depression (cf., Clague & Denlinger, 1994), but it may be more than one body. If it was there to begin with, it provides xenocrysts. Alternatively, magma is emplaced along it as olivine-rich sills (Boudier et al., 1996; Kelemen et al., 1997), with olivine concentrating by flowage differentiation or forming by reaction of the magma with harzburgite. When picrite erupts, differentiated magma from the near the base of the chamber that is uncontaminated by silicic continental crust flows along the listric fault, encounters olivine-laden crystal sludges, scavenges or entrains olivine and spinel, and pushes out onto the floor of the depression. There it flows into lakes to form pillows and aquagene tuffs with quenched glass. Basalt contaminated by continental crust erupts above the magma chamber. Later, some of it flows far enough to reach the depression (Maligât Formation), but only after much picrite has erupted (Vaigat Formation). Helium with high (old) $^3\text{He}/^4\text{He}$ (R/R_0) is incorporated into primitive but otherwise uncontaminated magma migrating through restite lithosphere (Natland, 2008). The ultimate source of primordial helium and its mobility remain controversial (Starkey et al., 2009).

Finally, I suggest that the magma chamber was not a massive body of circulating melt, a type of Skaergaard intrusion, but instead was largely crystalline, much like that at spreading ridges, and in which the network of crystals effectively served as a filter to sieve out phenocrysts (Natland & Dick, 1996). Crystallization differentiation proceeded by precipitation of crystalline material within the network of pre-existing crystals, and which had low melt porosity. Compaction from overburden and tectonic deformation assisted migration of the interstitial melt, and it vacated the magma body into the listric fault zone by means of a type of filter pressing. It left without phenocrysts, and shifted away from any additional precipitation of plagioclase by continued reaction with mantle rock. The sign of cotectic crystallization in BWG glass is thus a response to cryptic variation during adcumulus growth; it is no different from the mechanism that so consistently produces eruption of strongly differentiated yet aphyric basalt on the summit of the East Pacific Rise (Natland & Dick, 1996).

Helium

Where the main concern is temperature, I can only touch on noble gases. *Starkey et al.* (2009) concluded that the helium isotopic ratios of picrite from this region, which reach some extraordinarily high values of $^3\text{He}/^4\text{He}$, have nothing to do with other trace elements and isotopes; those instead could reflect other source variations (*Starkey et al.*, 2009) and/or, as I believe, crustal contamination. *Starkey et al.* (2009) consequently invoke an independent non-lithospheric source for helium with high- $^3\text{He}/^4\text{He}$. To my mind, this is a matter of where and when. The key to very high- $^3\text{He}/^4\text{He}$ is age. Wherever the He is, it retains an isotopic ratio consistent with something that happened very early in Earth history, perhaps nearly from the accretionary dawn of time. I also consider that old helium was likely separated from U and Th, and in general from the magmaphilic isotopes of Sr, Nd and Pb, by magmatic processes. The problem here is not one of the partition coefficient between helium and olivine, as people most commonly tend to treat it, but of partitioning between melt and vapor; then the nucleation of bubbles on growing crystals; and finally the separation from molten liquid of crystal-rich rocks (cf. *Natland*, 2003).

So picture some very old event—a magma ocean? komatiites? formation of continental crust?—and the upper mantle that supplied those rocks. In short, picture a process of separation of He from U and Th in crystal cumulates or segregations, that is, in rocks that are entirely possible to lie now at deep levels beneath Archaean continental crust that is billions of years old. That is where helium with very high- $^3\text{He}/^4\text{He}$, and isolated from U and Th, could still be. Being a mobile volatile, it may no longer be precisely where it was to begin with, but that is why helium with high- $^3\text{He}/^4\text{He}$ is found in some residual mantle xenoliths (*Poreda & Farley*, 1992) and why it seeps into formation fluids in cold sedimentary basins (*Castro et al.*, 2009). Thus it was in the pathway of migrating BWG picrites during the Paleocene (Figure 5). Olivine is a helium time capsule, and the bubbles within it are easily cracked by migrating magma and even by simple fracturing.

Summary

I hold to my views. Backtrack calculations using olivine-liquid FeO-MgO equilibria to estimate temperatures of crystallization of olivine in liquids postulated once to have been in equilibrium with the mantle are suspect. Even for Baffin Bay and West Greenland, where picrite abounds, the actual details of their crystallization and mixing histories confound the basic assumptions behind the backtrack technique, viz:

1. that all olivine in a picritic sample crystallized along a single olivine-controlled line of descent;
2. that olivine-addition control lines are strictly that;
3. that host liquids to picrites themselves represent olivine-controlled liquids.

These assumptions result in calculation of parental liquids with >20% MgO contents and crystallization temperatures in excess of 1500°C, but they are wrong. The procedure of *Poustovetov & Roeder* (2000) that relates liquid and spinel compositions to temperature instead suggests maximum MgO content of ~15% for West Greenland picrite, and about 1350°C for the temperature of crystallization of spinel intergrown with the most magnesian olivine found in any West Greenland sample ($\text{Fo}_{92.8}$). A technique utilizing Cr_2O_3 in olivine suggests higher temperatures than those estimated from spinel crystallization, but differences between the two techniques are not yet resolved. Additional detailed petrographic and compositional studies of olivine Fo_{91-93} and its inclusions of spinel and glass are warranted.

I suggest instead that BWG picrites result principally from injection of fairly typical depleted MORB into and through ancient subcontinental lithosphere and perhaps lower continental mafic crust during initial continental breakup. Interaction with that lithosphere and crust slightly perturbed the bulk compositions of the magmas and added abundant

olivine phenocrysts. The geochemical signal is mainly that of depletion, but with some effects (e.g., change in ΔNb and $^3\text{He}/^4\text{He}$; Figure 5) incorporated from the rocks through which the magmas passed.

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