Magmatism in the North Atlantic Igneous Province – an unconventional peridotite dominated plume or not a plume at all?

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1. **Introduction.**
2. **Global variations in mantle temperature – the current state of the art**
3. **Petrological methods of deriving T_P.**
4. **Peridotite solidi, olivine saturation and decompression melting pathways.**

4.1 Iron oxidation state in mantle peridotite and derivative primary magmas.

5. **Source lithology - Peridotite and Pyroxenite melting.**
6. **Trace element constraints on the source of NAIP lavas.**
    a. *Calculation of hybrid accumulative fractional melts – Methodology*
    6.2 Results of modelling
7. **Major Elements and a re-evaluation of T_P.**
8. **Olivine chemistry.**
9. **The alleged niobium anomaly (ΔNb)**
10. **Run out of steam**

ABSTRACT. Unlike many other Large Igneous Provinces, the North Atlantic Igneous Province (NAIP) is dominated by lavas that have compositions that are consistent with derivation largely from mantle peridotite. The major element compositions of NAIP lavas can be explained by melting of a mantle source that was ~90% peridotite and 10% silica deficient (SD) pyroxenite, and these melts subsequently underwent polybaric fractionation in the crust, over pressure ranges from 1.6 GPa to the surface. This general mechanism provides a simple explanation for the diversity of lava compositions seen both in the oceanic (Iceland) and continental (e.g. British Palaeocene Igneous Province; BPIP) sectors of the NAIP. To generate 20km thickness of crust at the northern coast of Iceland by this method requires T_P~1450°C. The Ni and Ca contents and Fe/Mn of olivine phenocrysts in NAIP lavas, which are often used as an indicator of pyroxenite source lithology, do not require a major contribution from pyroxenite to magmatism. However, because the major element compositions of melt produced from SD pyroxenite can be indistinguishable from peridotite-derived melts, major element geochemistry cannot be used to reliably identify pyroxenite source lithologies. Similarly, petrological methods which use major elements for estimating mantle potential temperature must be treated with caution because while pyroxenite and peridotite melts may be geochemically indistinguishable, melt productivity and melting behaviour may still be markedly different for each lithology. The trace element diversity of magmas that were not subjected to crustal contamination are explained by mixing of melt derived from peridotite with melt from a hybrid peridotite-pyroxenite source. A model is developed whereby isobaric accumulative fractional melts formed from hybrid olivine-bearing peridotite-pyroxenite lithologies, mix with shallower melts of mantle peridotite to generate melt with the observed range in Nb/La, Nb/Zr as well as other trace element characteristics. The hybrid lithology was likely to have been generated by solid-state interactions between recycled crust and peridotite, which would have provided ample opportunity to produce isotopic diversity within the NAIP mantle. The lithological structure of the NAIP mantle is therefore quite unlike that proposed for the Galapagos, Hawaii or Canary Island LIPs, where pure pyroxenite melts plus melts from a variety of hybrid lithologies must undergo complex mixing relationships to explain the major and trace element compositions of lavas and their component olivine phenocrysts. Since the pure pyroxenite components in the mantle are often considered represent recycled oceanic crust entrained in mantle plumes, then the absence of pure pyroxenite melts in

the NAIP coupled with the dominance of peridotite-derived magmas suggests that the NAIP is not a typical plume system.

1. Introduction.

The prodigious volumes of basaltic lavas and their associated differentiates that are erupted to form Large Igneous Provinces (LIPs) are generally considered to involve a flux of heat and material from deep in the mantle, possibly from the d'' layer core mantle boundary. Mantle potential temperatures (T_P) are considered to be higher by 200-300°C at LIPs than the ambient mantle that produces mid-ocean ridge basalts (MORB; T_P~1350°C). Elevated T_P provides an opportunity for melting of mantle peridotite to begin at the point where the adiabat for the appropriate T_P intersects the dry peridotite solidus, and for a system such as Hawaii with T_P ~1550°C, the initial pressure of melting (Pi^{Per}) is ~4.5 GPa depending on the peridotite solidus parameterizations that are used (Herzberg and Gazel 2009). Since melting of mantle peridotite occurs by decompression, the higher the value of Pi^{Per}, the larger the volume of basaltic magma that can be produced, and since the MgO content of primary melts of peridotite increase with increasing temperature, large volumes of picrite might be expected from mantle with high T_P. Conversely, it has also been argued that some CFBs (e.g Ferrar Province, and parts of the Central Atlantic Magmatic Province) may owe their origin to decompression melting of internally heated mantle, with plate boundary forces providing the mechanism for stretching the lithosphere and allowing passive upwelling of peridotite (Brandl et al. 2013; Coltice et al. 2009; Hole 2015). In these cases, internal warming of mantle insulated by continental lithosphere occurs, and does not necessarily require a mantle plume or long-lived point-source thermal anomaly.

Whatever mechanism causes melting to commence, melting will cease when there is no longer mantle peridotite available or when adiabatic ascent of mantle is no longer possible and in most cases this is likely to be at the asthenosphere-lithosphere boundary this is termed here the final pressure of melting (Pf^{Per}). For continental flood basalts (CFBs), the continental lithosphere will prevent decompression of mantle peridotite, and so the depth to the base of the continental lithosphere can have an important influence on the extent of melting of peridotite (Hole and Millett 2016). Additionally, interaction between hot mantle-derived melt and fusible sialic crust can modify erupted compositions, such that erupted compositions no longer carry a signature of melting of pure mantle peridotite in the geochemical composition (Thompson et al. 1980; 1986; Larsen and Pedersen 2000; 2009; Hole et al. 2015; Hole and Millett 2016).

Mantle peridotite is not the sole mantle lithology that contributes to magmatism at LIPs. The material that is included with the flux from the D’’ layer may consist of oceanic crust that was recycled during subduction (e.g. Sobolev et al. 2007). Transformations of oceanic crust in the lower and upper mantle results in the formation of a range of pyroxenite lithologies, with each portion of the recycled crust contributing a different mineralogical entity (Herzberg 2011; Lambart et al. 2013). Basaltic crust is transformed into silica-enriched (SE) pyroxenite, and cumulates formed from the generation of the basaltic crust will form silica-deficient (SD) pyroxenite. The solidus of pyroxenite is highly variable, but laboratory experiments show that some alkali-rich SE pyroxenite will melt at considerably lower temperatures for a given pressure compared to mantle peridotite. Consequently, if pyroxenite is entrained in a mantle plume, it will begin melting before peridotite. This gives rise to a suite of magmas at the surface that were generated by melting of pyroxenite alone, but also allows for the solid-state, melt-melt or melt-solid interactions between peridotite and pyroxenite to form hybrid lithologies. However, petrological phase equilibria also show that melts generated from SE pyroxenite cannot undergo mixing with melts of mantle peridotite at the same pressure (a thermal divide exists at mantle pressures), since one magma is Si-over saturated and the other Si-undersatrurated; mingling of melts from SE pyroxenite and peridotite will cause crystallization rather than mixing (Herzberg 2011).

Many of the basalts erupted during the shield-building phases of magmatism in the Hawaiian Islands are considered to be melts of near pure SE pyroxenite (Sobolev et al. 2006; Herzberg 2006) and pyroxenite melting may begin at >8 GPa, considerably deeper than $P\mathit{i}^{Per}$ for Hawaii (~4.5 GPa). Sobolev et al. (2005; 2007) argued that the rising mantle plume beneath Hawaii contains eclogite bodies that start melting at about 190–180km depth. Melting of eclogite produces high SiO_2 initial melts which infiltrate into, and react with, the adjacent peridotite eliminating olivine and producing a solid pyroxenite. Both the reaction pyroxenite and unreacted peridotite melt at depths between 140 and 100 km, producing hybrid magmas by mixing in conduits and crustal magma chambers. Nevertheless, data from Mauna Kea reveal that low- and high-SiO_2 pyroxenite-sourced magmas coexist as discrete batches along with peridotite-derived magmas (Rhodes et al., 2012; Sobolev et al., 2005; 2007) leading Herzberg (2011) to promote the hypothesis that melts of pure pyroxenite may segregate from their source without mixing and be erupted at the surface as lavas. SD pyroxenite may melt under very similar P-T conditions to mantle peridotite (Keshav et al. 2004; Lambart et al. 2016). Key differences between melting of peridotite and SD pyroxenite include the fact that in SD pyroxenite garnet is likely to be stable to shallower levels (~1.7 GPa) than for peridotite (~2.7 GPa depending on T), and laboratory experiments and thermodynamic parameterizations (e.g. Keshav et al. 2004; Lambart et al. 2012; 2013; 2016) show that melt productivity for pyroxenite is likely to exceed that of peridotite under the same P-T conditions. Unlike the case for SE pyroxenite, there is no thermal divide between melts derived from SD pyroxenite and those derived from peridotite, so melts can freely mingle and mix with one another.

Identification of a pyroxenite source in erupted lavas is generally approached by the combined use of major element geochemistry of whole-rocks and the trace element content of olivine phenocrysts. Hawaiian SE pyroxenite melts are deficient in CaO for a given MgO content compared to melts of mantle peridotite, and this fingerprint is useful for their identification in that setting (Herzberg and Asimow 2008; 2015). However, whether a pyroxenite melt is CaO deficient depends on D^{CaO} during melting, and this is not always >1, particularly for SD pyroxenite (e.g. Jennings et al. 2016). Furthermore, the melting experiments of Keshav et al. (2004) show that melts derived from SD pyroxenite can have almost identical geochemical compositions, including CaO contents, to melts generated by melting of mantle peridotite.

Olivine phenocrysts that form in pyroxenite melts tend to have higher Ni and Fe/Mn but lower Ca than olivine phenocrysts that form in equilibrium with peridotite-derived magmas, and this is often seen as a more reliable fingerprint for pyroxenite involvement in magmagenesis at LIPs than using major element geochemistry (e.g. Vidito et al. 2013; Herzberg et al 2016). However, crystallization and equilibration of olivines at high pressures in the crust, or even in the mantle, can generate Ni, Ca and Fe/Mn signatures that are similar to those for pyroxenite melting (Vidito et al. 2013; Hole 2017). Since crystallization of basaltic magma at high pressures is a common occurrence in CFBs (e.g. Thompson 1982; Villger et al. 2007; Whitaker et al. 2007; Hole 2017) then the use of olivine trace elements may not be a unique fingerprint of pyroxenite involvement in magmatism.

Olivine-bearing hybrid lithologies formed by solid-state reactions between the recycled crust and the peridotite are another possible source for magmas at LIPs (Vidito et al. 2013; Herzberg 2011; Lambart 2017). Indeed, Lambart (2017) has recently proposed that an isochemical bulk mantle can produce contrasting trace element and isotopic melt compositions depending on the major-element compositions of the lithologies present in the source. This requires the existence of olivine-bearing hybrid lithologies formed by solid-state reactions between the recycled crust and the peridotite in the source of Icelandic basalts, and importantly, the Lambart (2017) model does not require pure pyroxenite melts to be involved in the generation of melt beneath Iceland. Similarly, Mathews et al. (2016) argued that the petrogenesis of Iceland Rift zone basalts required a three-lithology source (peridotite, pyroxenite and harzbugite) and that the 20 km thickness of basaltic crust in northern Iceland can be generated at $T_P=1480^{+37}_{-30}$°C from this lithologically heterogeneous source.

As has been previously shown (Hole 2017), olivine phenocrysts in NAIP lavas do not carry any substantial fingerprint of the involvement of pyroxenite in magmagensis, and indeed, petrological models for the estimation of T_P (e.g. PRIMELT3; Herzberg and Asimow 2008; 2015; Hole and Millett 2016) imply that the vast majority of melt produced beneath the NAIP was derived from mantle peridotite. However, since pyroxenite and peridotite melts may have very similar major element compositions, the use of trace elements in distinguishing between the contributions of different lithologies in the source are developed in this paper. The hybrid lithological model of Lambart (2017) for Iceland is investigated further, and developed to examine magma generation across the entire NAIP.

2. Global variations in mantle temperature – the current state of the art

The potential temperature (T_P) of the upper mantle is the temperature that a parcel of upper mantle would acquire if adiabatically brought to a near-surface pressure of 0 GPa. The commonly accepted value of T_P for ambient upper mantle is 1350±40°C, and is the T_P responsible for the generation of mid-ocean ridge basalts (MORB) at ridge segments that are distant in space and time from other thermal anomalies (e.g. Siqueiros Fracture Zone MORB; Herzberg & Asimow, 2015). The total global range of T_P required for the generation of Phanerozoic (<540 Ma) basaltic magmas is ~1350-1750°C (Fig. 1), with evidence for the highest temperatures coming from basalts at Tortugal in the Caribbean large igneous province (CLIP), where both petrological and olivine-spinel equilibration temperatures suggest T_P~1750°C (Trela et al. 2017). Archean komatiites are thought to have originated at T_P ~1730°C (Sobolev et al. 2016) which has led to arguments for a hotter mantle in the Archean compared to present-day.

In terms of the generation of primary magmas, T_P controls the pressure at which the initial intersection ($P\mathit{i}^{\mathrm{Per}}$) of the dry peridotite solidus and the mantle adiabat occurs. Since the MgO content of a primary magma formed by melting of dry peridotite increases proportionally with increasing temperature, T_P also controls the MgO content of a primary magma. From $P\mathit{i}^{\mathrm{Per}}$, fractional melting occurs along the adiabatic temperature–pressure melting path which is nearly coincident with the olivine liquidus (Herzberg and Asimow 2008). Fractional melting can be considered a series of small increments of equilibrium melt generated during decompression melting of a parcel of magma. At any pressure lower than $P\mathit{i}^{\mathrm{Per}}$, accumulated fractional melts (AFM) represent the integration of all the incrementally-produced small melt fractions over the pressure range of interest. Melting occurs at a rate of 8-10% GPa^{-1}, (Herzberg and Asimow 2015; Hole and Millet 2015) and at some final pressure of melting ($P\mathit{f}^{\mathrm{Per}}$) melting will cease. $P\mathit{f}^{\mathrm{Per}}$ is most likely to be the pressure at the base of the lithosphere and indeed, $P\mathit{f}^{\mathrm{Per}}$ has been used as a proxy for the asthenosphere-lithosphere boundary (e.g. Hole and Millett 2015; Matzen et al. 2016). The melting interval i.e. $P\mathit{i}^{\mathrm{Per}}$–$P\mathit{f}^{\mathrm{Per}}$, determines the extent of melting (F-AFM) which is expressed as a fraction of unity. The largest melt fractions possible must be produced from the hottest mantle with the thinnest overlying lithosphere. Whereas Trela et al. (2017) argue that melting to form picrites at Tortugal must have been initiated at > 10 GPa, T_P estimates for most picrites from the NAIP where probably generated at T_P ≤ 1550°C (Herzberg and Gazel 2009; Hole and Millett 2015) such that $P\mathit{i}^{\mathrm{Per}}$ ≤ 4.6 GPa. Maximum melt fractions for NAIP model primary magmas are occur at Baffin Island and which represent F-AFM up to 0.30.

3. Petrological methods of deriving T_P.

For lavas from NAIP, two main petrological approaches to deriving T_P have been applied; Firstly, a multicomponent melting model (Phipps Morgan 2001; Shorttle et al. 2014; Matthews et al. 2016) which is used to calculate the thermal structure and melt fraction for many individual melting regions with variable mantle temperature (T_P) and differing proportions of peridotite, pyroxenite and harzburgite. This model then utilizes the derived melt fraction to calculate crustal thickness. Forward models based on hypothetical end-member melts from the base and top of the melting region, representing absolute limits of fractional melt diversity, are considered from each melting region. The melting region which simultaneously satisfies crustal thickness, the amount of melt contributed by pyroxenite and the crustal thickness is then inverted to obtain T_P and the proportion of pyroxenite, harzburgite and lherzolite in the source. Mathews et al. (2016) reported $T_P=1480^{+37}_{-30}$°C for lavas from The Northen Volcanic Zone (NVZ) of Iceland and an estimate of 1318^{+44}_{-32}°C for MORB requiring a thermal anomaly of ΔT +160°C beneath Iceland compared to 'normal' MORB.

Hole and Millett (2016) utilized the PRIMELT3 model (Herzberg & Asimow, 2008; 2015) to generate primary magma compositions from major element data on lavas. PRIMELT software uses a mass balance solution to the primary magma problem calibrated to fertile peridotite KR-4003, derived from a parameterization of experimentally determined partial melt compositions (Herzberg & O'Hara, 2002; Herzberg, 2004a; 2004b; 2006; Walter, 1998). PRIMELT uniquely constrains the primary magma composition corresponding to given evolved lavas by computing a melt fraction that is common to both partial melts of mantle peridotite and to the primitive magmas from which the lava was derived. In this way, it differs from other methods that use olivine composition to constrain primary magma composition (e.g. Putirka, 2008; Putirka et al. 2007; Courtier et al. 2007). Critical aspects of the PRIMELT model are that melts must have had olivine as the sole liquidus phase during crystallization and melt must have been derived from volatile-deficient peridotite. Melts which show evidence of early augite crystallization can be effectively identified by the use of covariations between MgO and CaO, because augite causes significant fractionation of CaO relative to MgO, but olivine does not, and the data set used here has been filtered accordingly.

4. Peridotite solidi, olivine saturation and decompression melting pathways.

There are a number of parameterizations of the peridotite solidi in the literature that have been generated from laboratory experimental studies. The most commonly used are those of Hirschmann (2000) Katz et al. (2003) and Herzberg and Asimow (2008 modified in 2015). More recently, Sarafian et al. (2017) have determined both dry and 'damp' solidi that latter containing between 200 and 450 ppm H_2O but not being H_2O saturated.

Melting of mantle peridotite occurs where the adiabat for a given mantle potential temperature intersects the solidus. Thus, for the Katz et al. (2003) solidus and T_P=1480°C, melting would commence at the initial pressure of melting (Pi^{Per}) of 4.1 GPa and continue until the final pressure of melting (Pf^{Per}) when mantle peridotite is no longer able to undergo decompression melting, which is probably close to the asthenosphere-lithosphere boundary (Matzen et al. 2015; 2017). For the same conditions, with reference to both the Herzberg and Asimow (2015) and Sarafian et al. (2017) dry solidus, $Pi^{Per}=3.1$ GPa. Since the extent of melting is proportional to $P^{Per}i$–Pf^{Per} (Gazel et al. 2011; Herzberg and Asimow 2015; Hole and Millett 2016) for the same Pf^{Per} of about 1.5 GPa, the mantle peridotite melting along the Katz et al. (2003) solidus would produce a larger amount of melt than mantle peridotite melting along the Herzberg and Asimow (2015) or Sarafian et al. (2017) dry solidus. Melt generation approximates to 8-11%GPa^{-1} (Herzberg and Asimow; Matthews et al. 2016; Hole and Millett 2016) and so the Katz et al. (2003) solidus might be expected to yield 10% more melt than for mantle peridotite along the other two solidi.

Melting occurs by incremental addition of small melt fractions formed during decompression to form an accumulated melt fraction that separates from the mantle and is eventually emplaced as a lava or intrusion. Olivine saturation in a melt is a function of pressure temperature and composition (Putirka 2008). Once olivine saturation occurs, the magma will continue to ascend along a P-T pathway that relates to the olivine liquidus. The P-T pathway from Pi^{Per} to Pf^{Per} is therefore accompanied by the crystallization of olivine. Herzberg and Asimow (2015) argue that the olivine liquidus and decompression melting pathway are near coincident (Fig. 2) whereas the Putirka (2008) olivine saturation curve occurs at a higher T for a given P than the decompression melting pathway of Herzberg and Asimow (2015). The parameterizations used by Matthews et al. (2016) require a P-T pathway that has a lower T for a given P than the Herzberg and Asimow (2015) decompression pathway. Reasons for the differences in calculated P-T pathways are discussed elsewhere (e.g. Matthews et al. 2016) but what is important here is the effect they have on T_P estimates.

Fig. 2 illustrates two differing approaches to estimating T_P as derived by Herzberg and Asimow (2015) and Matthews et al. (2016) for lavas from Theistareykir and Kistufel. In Fig. a, the P-T pathways from Pi^{Per} to Pf^{Per} are coincident with the olivine liquidus (Herzberg and Asimow 2015) such that the highest pressure, hottest magmas from

Theistareykir required TP=1528°C and Pi^{Per}=4.1 GPa and for Kistufel T_P=1392 and P*i* 2.6 GPa. Olivine crystallizing at 0.8 GPa from the hottest Theistareykir and Kistufel magmas would realize olivine crystallization temperatures of 1460 and 1370°C respectively. Matthews et al. (2016) have proposed that fractional melt extraction from the mantle can generate multiple primary melts that are not fully mixed at the time of olivine saturation. This is in part corroborated by the work of Keiding et al. (2011) and Herzberg (2011) who described how incomplete mixing of fractional melts can cause primary magma MgO and FeO concentrations to be overestimated, which propagates into higher crystallization temperature estimates. Matthews et al. (2016) argued that the decompression melting pathway for an individual melting column has a shallower slope that that of Herzberg and Asimow (2015) and that melt derived from the same T_P might have very different olivine crystallization temperatures at 0.8 GPa, which depends both on olivine saturation temperature and the depth in the melting column from which the melt was derived. This is illustrated in Fig. 2. For a primary magma of composition 1 in Fig. 2b, formed at T_P=1480°C and Pi^{Per}=3.8 GPa, olivine saturation would occur at the point of melting. For melt 2 in Fig. 2b, T_P is the same, but the melt is derived from a shallower portion of the melting column and does not integrate with melt fractions from deeper in the column. Consequently, the decompression melting pathway must intersect the olivine saturation curves for the deepest and shallowest melts, such that crystallization temperatures of 1375°C and 1460°C (at 0.8 GPa) could be derived from the same melting column at T_P=1480°C. This provides an explanation for the wide range of aluminium-olivine-spinel partitioning temperatures derived for central Iceland lavas (Matthews et al. 2016). The apparent variations in T_P derived from PRIMELT3 for Theistareykir and Kistufel shown in Fig. 2 might also be explained by such an incomplete melt integration model, and when Iceland PRIMELT3 solutions are taken as a whole they scatter around the decompression melting pathways of both Matthews et al. (2016) and Herzberg and Asimow (2015).

4.1 Iron oxidation state in mantle peridotite and derivative primary magmas.

Estimates of primary magma composition require knowledge of the Fe_2O_3 and FeO contents of the lava. For a given analysis of total Fe, higher Fe_2O_3 requires that there be less FeO, which for petrological models such as PRIMELT3, propagates to a primary magma with lower MgO contents and lower source temperatures (Herzberg et al. 2007). However, it is the total iron content of a lava that is typically obtained by X-ray fluorescence analysis and this is reported as Fe_2O_3. In most cases it is assumed that

$Fe^{2+}/\Sigma Fe$ ~0.9, or other algorithms for $Fe^{2+}/\Sigma Fe$ can be utilized (e.g. Thompson 1982). $Fe^{3+}/\Sigma Fe$ is sensitive to low pressure fractionation as has been previously demonstrated for MORB tholeiites (Cottrell and Kelley, 2011; Shorttle et al. 2015). However, a significant difficulty with a discussion on iron oxidation state of basalts is the paucity of primary data for Fe^{3+} in basalts. Shorttle et al. (2015) determined $Fe^{3+}/\Sigma Fe$ for basalts from the Reykjanes Ridge using X-ray absorption near-edge structure spectroscopy (XANES) adding to the existing data base for MORB and OIB (Bezos and Humler 2005; Cottrell and Kelley 2009; 2011; Cottrell et al. 2013; Zhang et al. 2016; 2017; Helz et al. 2017). In addition, Breddam (2002) determined $Fe^{3+}/\Sigma Fe$~0.07 using Mossbauer spectroscopy on undegassed glasses from Kistufel. Because of the paucity of laboratory determinations of Fe^{3+} in basalts, Gaetini (2016) provided an internally consistent, thermodynamically-derived model for the behaviour of Fe_2O_3 and $Fe^{3+}/\Sigma Fe$ during partial melting of mantle peridotite and made three important observations; i) $DFe^{3+}_{\text{peridotite-melt}}$ is likely greater than the 0.1 that which is generally accepted (e.g. Canil et al. 1994); ii) $Fe^{3+}/\Sigma Fe$ in mantle peridotite decreases with increasing temperature and iii) $Fe^{3+}/\Sigma Fe$ increases with increasing extent of decompression melting.

Herzberg and Asimow (2008; 2015) recommended utilizing fixed Fe_2O_3/TiO_2 for estimating iron oxidation state when using PRIMELT3 for estimating primary magma compositions. The rationale here is that TiO_2 and Fe_2O_3 have similar compatibility ($DFe^{3+}_{\text{peridotite-melt}}$ ~0.1; Canil et al. 1994) during melting of mantle peridotite, such that fixed Fe_2O_3/TiO_2 accounts for variability in extent of melting in estimating Fe^{3+} of model primary magmas. However, Gaetini (2016) argued that the lack of a strong correlation between $Fe^{3+}/\Sigma Fe$ and indicators of extent of partial melting make such incompatible behaviour untenable. Indeed, Gaetini (2016) showed that $DFe^{3+}_{\text{peridotite-melt}}$ varies as a function of oxidation state of mantle peridotite (represented by ΔFMQ), pressure and temperature. For example, for spinel lherzolite with $Fe^{3+}/\Sigma Fe = 0.06$ that has undergone 5% partial melting, $DFe^{3+}_{\text{peridotite-melt}}$ increases smoothly from ~0.6 at 0.8 GPa and $\Delta FMQ = 0$, to ~1.1 at 2.8 GPa and $\Delta FMQ = -2.3$. Additionally, Gaetini (2016) showed that during decompression melting, ΔQFM correlates negatively with pressure of melting such that shallower, larger melt fractions are more oxidized than deeper, small melt fractions. Consequently, the argument that Fe_2O_3/TiO_2 can be used as a proxy for iron oxidation state of a magma may not be petrologically sound.

The uncertainty in petrological estimates of T_P (±40°C; Herzberg and Asimow 2015) using PRIMELT3, which are propagated from data from melting experiments, and those of 1480^{+37}_{-30}°C derived from the integrated geophysical-petrological method of Matthews et al. (2016) are consistent with a broad range of model primary magma $Fe^{3+}/\Sigma Fe$. For example, using PRIMELT3 and Fe_2O_3/TiO_2=1.0, lavas from Theistareykir yield model primary magmas which required TP=1486±40°C and $Fe^{3+}/\Sigma Fe$=0.069±0.016. These calculated T_P values are within the range of uncertainty for the Mathews et al. (2016) method and $Fe^{3+}/\Sigma Fe$ are equivalent to ΔQFM~–2.5 for melting at the reasonable pressure of ~2.8 GPa for this T_P (Hole and Millet 2016). $Fe^{3+}/\Sigma Fe$ estimates from PRIMELT3 for Theistareykir lavas are also close to the laboratory determinations of Breddam (2002) for glasses from Kistufel, but are significantly lower than the $Fe^{3+}/\Sigma Fe$~0.15 (ΔQFM~–0.7 at 1.0 GPa) estimated for primary magmas to Reykjanes Ridge MORB glasses (Shorttle et al. 2015). However, because ΔQFM of a primary magma varies as a function of both pressure and temperature, higher $Fe^{3+}/\Sigma Fe$ in Reykjanes Ridge MORB glasses compared to glasses from central Iceland is a predictable consequence of higher T_P and thus higher pressure of melting for central Iceland compared to the Reykjanes Ridge.

It can be concluded, therefore, that whereas variations in $Fe^{3+}/\Sigma Fe$ have an effect T_P estimates for a given suit of magmas using petrological models such as PRIMELT3, the uncertainty in T_P estimates is about the same as, or slightly larger than, the likely variability that might be caused by variations in $Fe^{3+}/\Sigma Fe$ in primary magmas derived from mantle peridotite of 'normal' oxidation state i.e. $Fe^{3+}/\Sigma Fe$=0.04-0.06 (Geatini 2016). However, absolute TP determinations for a given suite of basalts increase by about 50°C for a change in $Fe^{3+}/\Sigma Fe$ of 0.04-0.12.

5. Source lithology - Peridotite and Pyroxenite melting.

It is generally accepted that the source lithology for CFB and their oceanic equivalents is not exclusively mantle peridotite, and that magmas derived from partial melting of pyroxenite are an important component of CFB magmatism (e.g. Sobolev et al. 2009; Herzberg, 2011; Rhodes et al. 2012; Vidito et al. 2013; Trela et al. 2015). The pyroxenite source components in LIPs are generally considered to have been derived from oceanic crust recycled to lower mantle depths during subduction tectonics (Herzberg, 2011; Sobolev et al., 2007). The transformation of basaltic crust produces quartz or coesite eclogite in the upper mantle (Fig. 3), and these are termed stage 1 pyroxenites (Herzberg 2011). In the lower mantle, stage 1 pyroxenite will consist of Mg- and Ca-perovskite, stishovite and silica-poor sodium-rich aluminous phase with a calcium ferrite structure (CF). Cumulates of dunite, troctolite and olivine gabbro will yield olivine pyroxenite lithologies (Ol+Cpx+Gt) in the upper mantle and stishovite-perovskite lithologies in the lower mantle. Solid-state reactions occur between mantle peridotite and recycled oceanic crust because silica-bearing recycled crust is not stable in an olivine rich peridotite host, yielding stage 2 pyroxenites which will be coincident with the CaTs-En plane in the upper mantle and with the MgPv-CF plane in the lower mantle.

In the broadest terms, stage 1 pyroxenites can be classified into two distinct lithologies; silica enriched (SE) and silica deficient (SD). Primary or very primitive melts derived from SE pyroxenite (silica+pyroxene) are recognized because they plot on the SiO_2-rich side of the orthopyroxene-calcium Tschermaks component (Opx-CaTs) join in the molecular projection of the system olivine–Opx–CaTs–silica (Fig. 3). Conversely, melts derived from SD pyroxenite (olivine+pyroxene) plot to the SiO_2-poor side of the same plane. Since the Opx–CaTs join is a thermal divide at pressures associated with mantle melting i.e. > ~1.7 GPa (Kogiso et al. 2004; Lambart et al. 2013) derivative magmas from the melting of high and low SiO_2 pyroxenite have divergent fractionation histories. At near melting pressures, low SiO_2 pyroxenite may undergo mixing with melts of mantle peridotite to form hybrid magmas, whereas melt derived from high SiO_2 pyroxenite cannot (Sobolev et al. 2005; Herzberg, 2011).

The melting behaviour of pyroxenites is key to their involvement in the genesis of CFBs. It is a commonly held perception that pyroxenite is likely to melt at a lower temperature for a given pressure than mantle peridotite, such that for a given T_P the pyroxenite solidus is intersected deeper in the mantle than the peridotite solidus. Consequently, deep-sourced melts that escape to the surface have the potential to be derived entirely from pyroxenite. This observation led to studies which argue that a pyroxenite contribution to flood basalts should be greatest for mantle plumes situated beneath thick continental lithosphere (Sobolev et al. 2007). However, Lambart et al. (2016) showed that the Na, K and Ti content of pyroxenite has a profound effect on melting behaviour, and that some SiO_2 poor pyroxenite may have very similar melting behaviour to mantle peridotite. This is illustrated in figure 3 from which it is evident that the temperature at which pyroxenite begins to melt ($T_{5\%}$ of Lambart et al. 2016) at a given pressure decreases with increasing alkali index, and additionally, alkali index broadly increases with CaTs content of solid pyroxenite. In general, high-SiO_2, alkali-rich pyroxenites will melt at a higher pressure for a given T_P than low SiO_2, alkali poor pyroxenite. Consequently, the relationship between pyroxenite contribution to magmatism and the thickness of the overlying lithosphere is not straightforward and is strongly dependent on the bulk composition of the pyroxenite in the source region.

The MeltPX simulations of Lambart et al. (2016) show that melt productivity from pyroxenite is enhanced when its solidus is lower than that of peridotite and it begins to melt before peridotite. This is because when the low-solidus lithology starts to melt, heat diffuses into that lithology, enhancing the amount of melting of the low-solidus lithology and delaying the melting of the high-solidus lithology. If the pyroxenite solidus has a gentler slope than that of peridotite, the pyroxenite will stop melting once the mantle ascends to depths where the peridotite melts. This occurs because the P-T path for melting peridotite has a steeper slope than that of the solidus of the residuum pyroxenite. So much heat is consumed by peridotite melting that the pyroxenite remains below its solidus during further ascent. If the pyroxenite solidus gradient is only slightly gentler than that of the peridotite, then pyroxenite melting will persist, but at a greatly reduced rate, once peridotite melting begins (Phipps Morgan 2001).

Once a pyroxenite lithology begins to melt it is likely to produce a spectrum of melt compositions that depends on the temperature, pressure and extent of melting as well as the initial bulk composition of the pyroxenite (Kogiso et al. 1998; 2003; Keshav et al. 2004; Lambart et al. 2012; 2013; 2016). Consequently, using major or trace major element geochemical data to try and distinguish between pyroxenite- and peridotite-derived magmas in the lavas of LIPs is fraught with problems. Nevertheless, a commonly used indicator of the involvement of pyroxenite-derived melt in basalt petrogenesis is that pyroxenite-derived magmas tend to be deficient in CaO at a given MgO content compared to melts derived from volatile-free mantle peridotite (Herzberg and Asimow, 2008). However, fractionation of CaO relative to MgO is sensitive to the bulk mineral/melt partition coefficient for CaO (DCaO) in a magma, and some pyroxenites appear to have DCaO > 1 and others < 1 regardless of the initial concentration of CaO in the solid pyroxenite (e.g. Herzberg, 2011; Jennings et al. 2016; Hole 2017). Residual clinopyroxene during melting of pyroxenite is likely to result in the generation of CaO deficient derivative lavas and therefore depends on the extent of melting at which clinopyroxene is totally consumed (Lambart et al., 2016). An additional difficulty with using CaO and MgO contents of mafic lavas to identify pyroxenite involvement is that lavas derived from CaO-poor pyroxenite will fall in the same region of an MgO-CaO diagram as peridotite-derived melts that have fractionated Ol+Cpx at crustal pressures of about 1 GPa (Villger et al. 2004; 2007; Hole 2017). Crystallization of peridotite-derived magmas at such pressures tends to drive liquids to progressively more SiO_2-undersaturation with decreasing Mg#. Consequently, crystallization of a peridotite-derived melt at ~1.0 GPa can result in a CaO depleted, Si-undersaturated liquid composition that is indistinguishable from a magma derived from the melting of low SiO_2 pyroxenite such as those found at the Canary Islands and Madeira (Herzberg, 2011; Gurenko et al., 2013). Recycled pyroxenite is also known to be far more heterogeneous than mantle peridotite, both chemically and mineralogically, and pyroxenite melting can produce a diversity of derivative magmas, some of which may be very similar to peridotite-derived magmas (Keshav et al., 2004; 2007; Lambart et al. 2012; 2013; 2016;).

Fig. 4 shows CaO-MgO relationships and molecular projections in the systems Ol–An–Qz for melts of pyroxenite generated during laboratory experiments. The system Ol–An–Qz has been used to isolate the ranges of potential melts of peridotite across the known spectrum of Phanerozoic melting conditions on Earth (Herzberg and O'Hara 2002; Herzberg and Asimow 2008; 2015). If derived from peridotite, lavas that have primary magmas compositions which plot above the peridotite-pyroxenite divide in Fig. 4a) must also plot in one of the regions reserved for peridotite melts in Ol–An–Qz space. Large extents (>70%) of melting of pyroxenite 77SL-582 at 2.0-2.5 GPa (coincident with the CaTs-En-Di plane in Fig. 3; alkalinity index=0.29) produces liquids that are both within the peridotite melting region on a plot of CaO *versus* MgO and could have been generated by ~15% melting of spinel peridotite in terms of the Ol–An–Qz projection (Fig. 4b). But, consistent with their known pyroxenite lineage, these large melt fractions plot close to the 3 GPa eutectic in the system Ol–CaTs–Qz (Fig. 3b). So, for pyroxenite 77SL-582, and consequently for melting of pyroxenite in nature, major element geochemistry of lavas may not be capable of distinguishing between some pure SD pyroxenite-derived and peridotite-derived melts.

However, an observation can be made that near-pure melts of pyroxenite from Hawaii can be distinguished from peridotite melts based on relationships between Mg# and SiO_2/CaO (Fig. 5). Melts derived from SE pyroxenite have considerably higher

SiO_2/CaO than peridotite-derived melts, which results from elevated SiO_2 in the pyroxenite source coupled with low CaO abundances, which might in part be due to higher D^{Ca} during pyroxenite melting than during peridotite melting. Because continental flood basalts may undergo crystallization of the assemblage L+Ol+Cpx at high pressures in the crust, or even in the mantle (Herzberg and Asimow 2008; Kashev et al. 2007; Hole 2017), then SiO_2/CaO will be modified by early Cpx crystallization. Inspection of Fig. 5 shows that SiO_2/CaO increases when olivine and clinopyroxene are together on the liquidus (e.g. MPLF lavas; Fig. 5c), whereas crystallization at of Ol+Pl at low pressures maintains relatively consistent SiO_2/CaO. This means that it is not possible for a near-primary SE pyroxenite-derived mantle to attain the low SiO_2/CaO (<~5) characteristic of peridotite-derived melt at any Mg# for any pressure of crystallization. However, peridotite-derived near-primary magmas that crystallize at >1.5 GPa can inherit similar SiO_2/CaO and Mg# to differentiates of SD pyroxenite melts that form over a range of pressures (e.g. MPLF lavas in Fig. 7a and SR083-17.5 in Fig. 7c). Nevertheless, at high Mg# (>75) SD pyroxenite melts from Hawaii still have higher SiO_2/CaO than peridotite-derived melts, and the distinction between melts derived from the two sources is clear. Whereas these observational relationships allow the identification of pyroxenite in the source of Hawaiian lavas, when experimentally generated melt compositions from SD pyroxenite are considered, the relationships are more problematical. Laboratory generated melts of SD pyroxenite SL77-582 (Keshav et al. 2004) have similar Mg# and SiO_2/CaO to peridotite derived melts (Fig. 5) and therefore differentiates of these melts formed at any pressure will mimic the composition of peridotite melts in the same manner as is also evident from Figs 3 and 4. Therefore, the conclusion that can be drawn from Figs 3-5 is that near-pure melts of SE pyroxenite cannot have contributed significantly to magmatism in the NAIP, but a significant role for SD pyroxenite is not precluded. Furthermore, since hybrid pyroxenite-peridotite sources are likely to be important in the genesis of NAIP basalts (Matthews et al. 2016; Lambart 2017) then the use of major elements in distinguishing different source lithologies is further clouded. Another method of distinguishing source lithology must be sought, and the trace element variability of NAIP lavas will now be examined in this connection.

6. Trace element constraints on the source of NAIP lavas.

Lambart (2017) argued that the observed crustal thickness of the Iceland Rift Zones is consistent with about 10% recycled crust in the form of SD pyroxenite in the source, but also showed that there is no need to involve a contribution of 'pure' melts derived from a recycled basaltic component to explain the compositional variability of Iceland Rift Zone basalts. Instead, the spectrum of melts seen in Icelandic rift zones is more likely to be the result of the melting of hybrid lithologies formed by solid-state reactions between peridotite and pyroxenite i.e. Stage 2 pyroxenite. Lambart (2017) used T_P=1480°C from Matthews et al. (2016) and provided models for melting of hybrid pyroxenite-peridotite lithologies that produced 20km of new basaltic crust, that is, the thickness of the crust at the northern coast of Iceland. Here, the hybrid lithology model of Lambart (2017) is developed and applied to trace element systematics to examine the geochemical compositions of lavas across the NAIP.

A key feature of the trace element distributions in Icelandic and other NAIP basalts and picrites is the significant variability in ratios of highly incompatible elements (e.g. Nb/La 0.5-1.5, Ba/Th <1 to >200). Crustal contamination causes lowering of Nb/La and variable Ba/Th in NAIP continental lavas, but this can readily be accounted for by using isotopic compositions because the NAIP crust is generally of great antiquity and so Nd- and Pb-isotopic compositions are particularly good indicator of crustal interaction (Thompson et al. 1980; 1986; Thompson 1982; 1982; Kerr et al. 1999; Hole et al. 2015). For Iceland it is assumed that there is no interaction with continental crust (but see Torsvik et al. 2015). Since Nb/La does not vary significantly with progressive melting it is most likely to reflect differences between inter-element ratios of source components (Stracke and Bourdon 2009; Lambart 2017). Variability in other incompatible element ratios (e.g. Nb/Zr) are influenced by both source heterogeneity and progressive melting (Stracke and Bourdon 2009). For Iceland a 'depleted' component must exist which has Nb/La <0.5 but and second 'enriched' component with Nb/La up to 1.7 is also required to account for the variability in Icelandic rift and rift-flank magmas (Shorttle et al. 2013; Maclennan et al. 2003; Peate et al. 2010). Nb/Zr systematically increases tenfold (~0.01 to 0.10) with increasing Nb/La in the same lavas, but since Zr is generally more compatible than Nb during melting, variability in Nb/Zr reflects both mantle source compositions and melting dynamics. Additionally, high-precision Nb, Zr (and Y – see below) data are available for many NAIP lavas, and the majority of Nb determinations used here, at levels <2 ppm, were carried out by ICP-MS.

The MeltPX parameterizations of Lambart et al. (2016) allows predictions to be made relating to the contribution from peridotite- and pyroxenite-derived melt in a mixed peridotite-pyroxenite source region. The general melting characteristics of three different peridotite-pyroxenite pairs are given in Fig. 6, and Fig. 7 shows the situation for melting of peridotite and SD pyroxenite 77SL-582 at varying T_P and proportion of pyroxenite in the source. In the three cases shown, the total thickness of basaltic crust that can be formed is ≥20 km (Lambart 2017). The key observations that can be made from Figs 6 and 7 are as follows;

1) For SD pyroxenite 77SL-582, peridotite and pyroxenite both start melting together at temperatures in the range 1450-1500°C and over the range of 10% to 50% pyroxenite in the source.
2) For 77SL-52, increasing T_P from 1450°C to 1550°C increases both P_i^{Pyr} and P_i^{Per} from ~3.5 to ~5.0 GPa and increases the maximum values of tc^{Pyxte}/tc from ~0.55 to 1.00 although the melting interval where $tc^{Pyr}/tc > 0.50$ is relatively small (~1.2 GPa). The total crustal thickness formed at 1500°C for this configuration is 27 km.
3) For pyroxenite M5-40 there is a melting interval of approximately 0.5 GPa where pyroxenite is the sole contributor to magmatism. From 3.5 GPa both pyroxenite and peridotite contribute to magmatism
4) For pyroxenite B-ECL1 there is a significant pyroxenite-only melting interval from ~7 to ~3.5 GPa reaching a maximum of ~50% melt at 3.5 GPa. Consequently, pure pyroxenite melt might be expected to erupt at the surface.

Based on the observations made above, and following from Lambart (2017), for trace elements modelling, the configuration of 10% SD pyroxenite 77SL-582 and 90% peridotite at 1450°C, which gives precisely 20 km of crust has been used as the basis for trace element modelling.

6.1 Calculation of hybrid accumulative fractional melts - Methodology

1) For mantle peridotite, the source composition used was the source of mid-Atlantic Ridge depleted MORB outside the area of influence of Icelandic magmatism (Hole et al. 2015). This is close to the composition of KLB-1 used by Lambart (2017). Multi-element primitive mantle-normalized (Sun and Mcdonough 1988) for KLB-1 and the peridotite source used here are given in Fig. 8 and Table 1.
2) The composition of incremental fractional melts at 1% melt intervals over the range Pi^{Per} to Pf^{Per} were calculated based on partition coefficients (Table 2) given in McKenzie and O'Nions (1995) and Pertermann et al. (2004) which are also the default values used in alphaMELTS (Smith and Asimow 2005).

3) Accumulated fractional melts derived from mantle peridotite were calculated by integration of 1% increments of melting over the pressure interval for the appropriate T_P. The peridotite solidus used was that of Katz et al. (2003) as this is the parameterization used in MeltPX (Lambart et al. 2016). For T_P=1450°C, P_i^{Per}=3.5 GPa and melting was assumed to occur at 10% GPa^{-1}. Garnet-out was set at <2.7 GPa.
4) For pyroxenite melting, partition coefficients were taken from Pertermann et al. (2004) for compositions appropriate to SD pyroxenite (Table 3)
5) The trace element composition of pyroxenite (Fig. 7; Table 1) is close to that used by Lambart (2017) for KG2, the composition KG2 being modified iteratively to produce best fits for NAIP data as will be discussed below.
6) Pyroxenite melt productivity at a given T_P and proportion of pyroxenite in the source over the pressure interval of interest was taken from MeltPX.
7) Isobaric accumulated fractional melts were calculated for a given pressure using the proportional contribution to total melt thickness from pyroxenite and peridotite at that pressure, as described by Lambart (2017). This utilizes the parameters tc^{Pyr}/tc and tc^{Per}/tc in MeltPX, which are the proportion of the total thickness of crust derived from pyroxenite and peridotite melts respectively (Figs 5 and 6).

6.2 Results of modelling

Fig. 9 shows the relationships between Nb/La and Nb/Zr for accumulated fractional melts derived from melting of mantle peridotite and SD pyroxenite 77SL-582 (Table 3) at T_P=1450°C. Pure pyroxenite melts have higher Nb/La and Nb/Zr than peridotite melts regardless of melt fraction. Isobars represent mixing curves between peridotite- and pyroxenite-derived melts at the same T_P assuming they were generated separately, and were subsequently able to undergo perfect mixing. Because all melts of SD pyroxenite and peridotite generated in this model have similar levels of Si-saturation, there is no petrological barrier to mixing of melt (Herzberg 2011). The isobaric accumulative fractional melting (I-AFM) curve represents the trace element composition of the melt produced from a parcel of mantle comprising 90% peridotite and 10% pyroxenite decompressing to that pressure, with the composition weighted according to the melt productivity rate for each individual lithology, which was derived from MeltPX. At the point of intersection of the I-AFM curve with the 1.0 GPa isobar, sufficient melt would be generated to produce 20km of new crust (point 1 in Fig 9a). Points along the I-AFM curve represent the compositions of the predominant melt formed by decompression melting from a hybrid peridotite-pyroxenite lithology at a given pressure. Thus, a parcel of mantle comprising 90% peridotite and 10% pyroxenite that melts by decompression to 3 GPa would be expected to have Nb/Zr~0.134 and Nb/La~1.41 (point 2 in Fig. 9a). At this point, approximately 70% of the trace element budget in the melt would be from a peridotite source (Fig. 9a). Increasing the proportion of pyroxenite in the source moves the I-AFM trajectory closer to the pure pyroxenite melting curve. Data points which fall at lower Nb/La than the I-AFM curve require less participation from pyroxenite than the isobaric melts, and most likely require the mixing of pyroxenite-rich melts with pure peridotite melt, as shown by the isobars in Fig. 9b. The I-AFM melting trajectory does not change significantly over the temperature interval T_P=1350-1500°C (Fig. 9b) showing that the mixing relationships are not strongly temperature dependent. However, TP of ~1450°C must be maintained to generate 20 km of basaltic crust.

Data for basalts from the mid-Atlantic Ridge between latitudes 57°N and 60°N, which are outside the influence of the Icelandic melting regime (Fitton et al. 1997), fall close to low pressure isobars in Fig. 9b and require a mantle source that has a trace element contribution that is more than 80% peridotite. The position of these data is not strongly temperature dependent in relationship to the modelled paramters. Basalts from the East Pacific Rise (EPR) at 10°N have a similar range in Nb/La to, but slightly higher Nb/Zr than, MAR basalts, suggesting a higher mean pressure of melting at the EPR compared to the MAR. However, a more likely explanation for these compositional differences is that the peridotite source for EPR basalts has a slightly higher Nb/Zr than that of the MAR. This is also reflected in differences between Atlantic and Pacific MORB in Zr/Y-Nb/Y space (Fitton 2008). Extending the MAR data set to the higher latitude of 63°N, where the MAR is influenced by the Icelandic melting regime, extends the data array shown in Fig. 9b to higher Nb/La and Nb/Zr (Fig. 10a). Indeed, some MAR samples at ~63°N require a significant pyroxenite component in their source according to Fig. 10a, and some data points extend into the region reserved for near-pure pyroxenite melt. Increasing Nb/La of the pyroxenite component from 1.2 to 1.4 would shift the I-AFM curve to higher Nb/La values and could result in the most pyroxenite-rich MAR basalts falling along the I-AFM. However, if this is achieved by increasing the Nb concentration in the pyroxenite source, this will also decrease Nb/Zr, and the MORB data would no longer fall on the low pressure isobars. Decreasing La would be able to achieve the desired effect, but this would then require a pyroxenite with a positive Nb spike on primitive mantle normalized diagrams compared to adjacent elements.

Data for the WVZ (Reykjanes Peninsula) exhibits a similar distribution to data from the MAR (Fig. 10b), but at slightly higher Nb/Zr which may equate to a higher pressure of melting, a predictable consequence of higher T_P beneath the WVZ than most of the MAR. NVZ lavas from Theistareykir form a broad array emanating from the I-AFM curve towards more peridotite-rich compositions (Fig. 10c). A small number (8 samples out of 185 shown) might represent near-pure pyroxenite melts. Thesitareykir data require pressures of melting of ~1.0 to ~3.2 GPa consistent with a melting column at ~1450°C. Primitive mantle-normalized multi-element diagrams for the most peridotite- and pyroxenite-rich model liquids are compared to data for basalts from Iceland in Fig. 11. The samples from the WVZ and Theistareykir have been chosen based on their positions in Fig. 10, that is the lowest and highest Nb/La lavas in each case. Two features of Fig. 11 are significant. Firstly, lavas from the NVZ and WVZ selected in this way are nearly identical for all 19 trace elements used in the modelling. Secondly, there is good agreement between the modelled compositions in Fig. 11b and the lavas shown in Fig. 11a. Consequently, based on 19 trace elements of variable compatibility, the trace element variability of all Icelandic basalts could be produced by melting of a bi-lithological, hybrid peridotite-pyroxenite source. Data for Rift Flank lavas from the East of Iceland at Snaefellsness (Fig. 10c), and from the West of Iceland at Snaefellsjökull plot close to the I-AFM curve at 3.5-3.0 GPa. At around this pressure, the I-AFM and pure pyroxenite melting curves intersect, making the Rift Flank magmas potentially the most pyroxenite-rich in Iceland.

Lavas from Baffin Island (Fig. 12a) form an array between the I-AFM and peridotite melting curves subparallel to 2.5 to 3.0 GPa isobars. The 'E-MORB' lavas of Starkey et al. (2009) are, according to the current model, the most pyroxenite-rich lavas and plot close to the I-AFM melting curve. However, these data still only require a trace element contribution from pyroxenite of <20%. At Disko Island (Fig. 12b), lavas are peridotite dominated and no lavas require more than ~20% contribution to their trace element budget from pyroxenite. Enni Formation lavas from the Faroe Islands Basalt Group (FIBG; Fig. 12c) have a bimodal distribution, which is the previously described distinction between high TiO_2, high Nb/La and low TiO_2, Nb/La types (Soager and Holm 2011; Millett et al. 2017). This might require up to 30% pyroxenite in the source of the high TiO_2 basalts, but virtually no contribution from pyroxenite to the low TiO_2 basalts. Lavas from the British Palaeocene Igneous Province (BPIP) are again peridotite dominated, which is consistent with the conclusions of other studies (e.g. Fitton et al., 2000; Hole et al. 2015; Kerr et al. 1999).

East Greenland lavas exhibit a slightly different set of trends from the rest of the NAIP (Fig. 13). Data for some Prince af Wales Bjerge lavas plot close to the high pressure (~3.0 GPa) I-AFM curve, whereas others require less pyroxenite and overall lower pressures of melt formation. In common with lavas from Sortebrae (Fig. 13b) there is evidence for a steep trend of variable Nb/La (0.7-1.5) but with little concomitant change in Nb/Zr (0.07-0.10) such that groups of lavas form arrays oblique to the isobars. This could be achieved by mixing between small melt fractions derived from mantle peridotite and melts falling along the I-AFM melting curve. Lavas from Kangerlussuaq exhibit a similar relationship (Fig. 13a).

Jan Mayen Island lies approximately 650 km NNE of Iceland at the locus of a fracture zone which cuts the MAR, and is made of Miocene-Recent ankaramites, basalts and associated differentiates. The mafic lavas of Jan Mayen plot very close to the intersection of the model I-AFM and pure pyroxenite melting curves at ~3.4 GPa (Fig. 13b) and fall in a similar position to some lavas from the Icelandic Rift Flanks.

7. Major Elements and a re-evaluation of T_P.

According to the current model, lavas from Jan Mayen Island and Rift-Flank magmas from Iceland represent the most pyroxenite-rich melts in the region, and may indeed represent pure pyroxenite melts. These lavas have been investigated using PRIMELT3 to identify those which are likely to have undergone only olivine fractionation (Herzberg and Asimow 2008; Hole and Millett 2016; Hole 2017). Using the systems CaO-MgO-Al_2O_3-SiO_2 (CMAS; O'Hara 1972) and Ol-CaTs-Qz (Fig. 3), and following the method of Herzberg (2011), olivine has been added to the selected lavas to estimate the likely major element composition of potential near-primary pyroxenite melts, and these are included in Figs 3 and 4 and summarized alongside pyroxenite melt compositions from Hawaii in Table 4. If it is assumed that Snaefellnes and Jan Mayen lavas are the closest approximation to pure pyroxenite melt in the region, the major element geochemistry of these melts cannot be distinguished from accumulative melt fractions of ~0.1 derived from mantle peridotite leaving a garnet peridotite residue (Fig. 4). Consequently, the apparent change in T_P from

~1500°C to ~1450°C from Iceland Rift Zones to Rift Flanks required from PRIMELT3 models (Hole and Millett 2016) is more likely to reflect changes in the proportion of pyroxenite in the source region in the two locations. A re-evaluation of the PRIMELT3 T_P calculations for Icelandic lavas using only samples which have Nb/La <1.25, and are thus most likely to have been dominantly derived from peridotite, yields 17 solutions with T_P=1477±22°C a value which is entirely comparable with the value of T_P= 1480^{+44}_{-32} °C for Iceland Rift Zone lavas derived by Matthews et al. (2016). Similarly, using only low Nb/La samples from other locations in the NAIP yields PRIMELT3 temperature estimates of T_P=1538±34°C for Baffin Island, 1526±26°C for Disko Island and 1523±39°C for the Enni Formation of the FIBG.

8. Olivine chemistry.

Equilibrium olivines that crystallize from pyroxenite-derived melts tend to have higher Ni, lower Ca and higher Fe/Mn than olivine in equilibrium with peridotite-derived magmas (Fig. 14). A substantial data base for trace elements in Icelandic lavas exists, so and samples from Theistareykir and Snaefellsness are used here to illustrate the principals involved. Crystal lines of descent (CLDs) have been calculated for olivine in equilibrium with 'pure' pyroxenite melts from Hawaii, using the Ni data given in Sobolev et al. (2005). Melts of high SiO_2 pyroxenite (Makapuu stage of Koolau) crystallized olivine with considerably higher Ni at a given Fo content than any olivine phenocrysts found in lavas from Iceland, supporting the notion of peridotite-dominated melting at Iceland (Hole 2017). CLDs for olivine in equilibrium with Hawaiian low SiO_2 pyroxenite have Ni contents which approach those with the highest Ni content in Icelandic lavas. However, it has been shown using whole-rock data and olivine trace element data (Hole 2017) that some olivine in lavas from Theistareykir crystallized at ~1.0 GPa, the early crystallization of the assemblage L+Ol+Cpx causing higher Ni and lower Ca contents at a given Fo compared to olivine that crystallized from the assemblage L+Ol at low pressures. This is a predictable result of $k_D{}^{Cpx}{}_{Ni} < k_D{}^{Ol}{}_{Ni}$ (Herzberg and O'Hara 2002) and the removal of CaO from the melt during Cpx crystallization. The combination of moderate Ni and low Ca in Theistareykir olivine can thus be satisfactorily explained by differences in the pressure of equilibration of olivine. In addition, Fe/Mn for all olivine phenocrysts in lavas from Theistareykir are within the limits expected for melts derived from mantle peridotite. Some olivine in lavas from from Snaefellsness (Enni; Sobolev et al. 2007) are deficient in Ca compared to those from Theistareykir (Fig. 14b), and fall close to the compositions of olivine found in Hawaiian low SiO_2 pyroxenite-derived melts. However, these same olivines do not have elevated Ni contents and it is more likely that they crystallized at pressures > 1 GPa. The slightly higher Fe/Mn in low Ca olivines from Snaefellsness (Fig. 14c) can also be satisfactorily accounted for by their crystallization at >1.0 GPa.

A primary peridotite-derived melt formed at T_P=1480°C would contain 16.7 wt% MgO and ~510 ppm Ni (Herzberg et al. 2016). Olivine that crystallized at depth from this primary magma would contain ~3100 ppm Ni at Fo ~91.4, if the magma was subsequently rapidly emplaced at the surface (Matzen et al. 2017). A near primary pyroxenite melt with ~18 wt% MgO is estimated to contain ~800 ppm Ni (Sobolev et al. 2005). Olivine that crystallized at depth from this magma would be expected to contain up to 4000 ppm Ni at Fo ~90.4, if the magma was subsequently rapidly emplaced at the surface (Fig. 14a). These differences in olivine Ni content are sufficient to generate the separation between the pyroxenite and peridotite CLDs shown in Fig. 14. Since the contribution from pyroxenite-derived sources to most NAIP magmas is <30%, then the total Ni budget of a high pressure I-AFM would be expected to be ~600 ppm Ni, which would result in the crystallization of olivine with ~3400 ppm Ni. This level of Ni in olivine can also be readily achieved by melting of mantle peridotite at ~17.5 wt% MgO or T_P~1500°C. Consequently, there may not be any expression of pyroxenite melting in olivine phenocrysts.

9. The alleged niobium anomaly (ΔNb)

Fitton et al. (1997) presented a much-used diagram of Nb/Y *versus* Zr/Y (Fig. 14) for MORB and OIB and argued that the higher Nb/Y for given Zr/Y seen in Icelandic neo-volcanic zones compared to MORB, was the result of a high Nb component in the Iceland mantle plume. The value ΔNb, or the vertical displacement of an individual data point from the lower bound shown in Fig. 14, has subsequently been used widely to assess the contribution of Iceland plume mantle to NAIP magmatism, and indeed in a global context. Fig. 14 shows that the I-AFM melting curve for the hybrid peridotite-pyroxenite configuration used in the present modelling falls above the lower bound of the Iceland plume region. That is, all hybrid I-AFM melts have positive ΔNb (0.0 to 0.7) whereas peridotite melts all have negative ΔNb (–0.6 to –0.3). Furthermore, isobars for mixing of I-AFM melts with peridotite-derived melts have a curvature that is close to the observed data for the MAR (~1 GPa) and Baffin Island (2.0-2.5 GPa). Consistent patterns also emerge for MPLF lavas, the majority of which are peridotite dominated according to Figs

12 and 14, as also concluded by Hole (2017). The near I-AFM melt compositions from the MPLF are also known as the Coire Gorm tholeiites (Kerr 1999) and these occur close to the top of the Mull lava pile. Chambers and Fitton (2000) argued that these lavas represented the first manifestation of the Iceland Plume beneath western Scotland, but it is now concluded that these lavas are simply I-AFM melts.

Perhaps most significantly, the pyroxenite trace element composition used in this study does not have a well-pronounced positive anomaly at Nb relative to adjacent elements (Fig. 8). Indeed, the distribution of trace elements shown in Fig. 8 is like that for basalts from some Oceanic Plateaus (e.g. Ontong Java; Fitton et al. 2004).

10. Run out of steam

Figure Captions.

Fig. 1. Summary of global petrological estimates of T_P (diamonds ±40°C) and ranges of olivine crystallization temperatures for magnesian olivine (T_{OL}) derived from the Al-in-olivine geothermometer. The lower cross-hatched region represents the range of T_{OL} for olivine which crystallized from near-primary magmas formed at ambient T_P~1350°C. The upper cross-hatched region represents the mean maximum T_{OL} for olivine phenocrysts in basalts from Tortugal, the highest recorded Phanerozoic T_{OL} to date.

Fig. 2 P-T diagrams giving comparisons of the methodology relating to the calculation of T_P and olivine crystallization temperatures (T_{cryst}) using the PRIMELT3 model of Herzberg and Asimow (2008; 2015) and the method described by Matthews et al. (2016). a) PRIMELT3 solutions for lavas from Theistareykir and Kistufel with Fe_2O_3/TiO_2=1.0 which for Theistareykir gives $Fe^{3+}/\Sigma Fe$=0.069±0.016. For Theistareykir, PRIMELT3 gives T_P=1486±23°C and for Kistufel T_P=1410-1424°C. T_{cryst} at 0.8 GPa would be ~1375°C (Fo~91.2) and ~1460°C (Fo~91.5) for the lowest and highest temperature melts from Kistufel and Theistareykir respectively. The black star represents the composition of MORB primary magmas from the Siqueiros Fracture Zone (Herzberg and Asimow 2015). b) P-T diagrams for magmas formed at the T_P estimate of Matthews et al. (2016) of T_P~1480°C. For the Herzberg and Asimow model; point 1 is the intersection of the dry peridotite solidus with the 1480°C adiabat, which represents the initial pressure of melting (P*i*) at ~1570°C for a primary magma with 16.7 wt% MgO; the grey dot is the mean pressure of melting (P*m*) as described by Hole and Millett (2016); the black dot is the final pressure of melting (P*f*). P*m* and P*f* both fall along the decompression melting path which intersects the base of the crust at 0.8 GPa and gives a crystallization temperature for olivine (T_{cryst}) of 1440°C. For the Matthews et al. (2016) model; point 1 is P*i*. If a magma that formed at an infinitesimally lower pressure than P*i*, was able to segregate and reach the base of the crust, it would fall along the olivine saturation curve giving T_{cryst} ~1460°C (Fo~91.3). The Matthews et al. (2016) decompression melting path from point 1 is of a shallower slope than that described by Herzberg and Asimow (2008; 2015) such that magma ascending along this pathway would intersect the base of the crust at T_{cryst} ~1375°C (Fo~91.0). The Matthews et al. (2016) model therefore allows two magmas with the same T_P to have widely differing T_{cryst} at any given pressure of crystallization. This requires that melts segregate from differing depths in the same melting column such that accumulative fractional melting is not efficient. Using the Herzberg and Asimow (2008;2015) decompression melting pathways and calculating up-pressure from the highest T_{cryst} estimate (1460°C) of Matthews et al. (2016) gives a primary magma composition with T_P~1510°C (point 1) and 17.9 wt% MgO. Applying the same criteria to the lowest T_{cryst} estimate (1375°C) from Matthews et al. (2016) gives a primary magma composition with T_P~1400°C (point 2) and 13.7 wt% MgO. c) schematic 1-dimensional melting columns for Iceland rift zones and MORB illustrating mantle lithology (after Hole and Millett 2016).

Fig. 3. a) Molecular projection from or towards diopside onto the plane Fo–Qz–CaTs (Forsterite–Quartz–Calcium Tschermak's molecule) showing the areas occupied by stage 1, silica-deficient (SD – orange field) and silica-enriched (SE – blue field) pyroxenites. The pyroxene-garnet plane is used for taxonomical purposes to identify high- (SE) and low-SiO_2 (SD) pyroxenites on the SiO_2-rich and -poor sides, respectively. SE pyroxenites are formed by the transformation of basaltic oceanic crust to quartz or coestite eclogite, whereas cumulates in the same crust will form SD pyroxenite (Ol+Cpx+Gt). Pyroxenites formed in the upper mantle and Transition Zone by the

solid-state reaction of MORB with fertile peridotite (stage 2 pyroxenites) have compositions that are coincident with the Diopside–Enstatite–CaTs plane (pyroxene–garnet plane in pink). Pyroxenites formed in the lower mantle by the solid-state reaction of MORB with fertile peridotite have compositions that are coincident with the Ca and Mg perovskite–CF plane (red), where CF is an aluminous phase with the calcium ferrite structure. Blue lines are cotectics for L+Cpx+Gt±Opx±Cpx at the pressures indicated in blue. After Herzberg (2011). b) the same diagram as in a) but showing the projection of melt compositions produced during melting experiments on a range of SD pyroxenites. Estimated compositions of near-primary pyroxenite melts from Hawaii (orange dots, SD; blue dots SE) are taken from Herzberg (2006) and Sobolev et al. (2005). Near-primary pyroxenite-derived melts from Snaefellsness (black dots) and Jan Mayen Island (grey dots) were calculated using the method of Herzberg (2011) which is discussed in the text. c) Temperature at which 5% melting occurs for a given pyroxenite at 3.0 GPa *versus* alkalinity index ((1−Mg#)+10(Na+K)−Ca+Al as molecular proportions). Calculated using data from MeltPX (Lambart et al. 2016). 1500°C is the intersection of the 1450°C adiabat with the dry peridotite solidus of Herzberg and Asimow (2015) and 1440°C the same for the solidus of Katz et al. (2003).

Fig. 4. a) Weight% CaO v*ersus* MgO and b) molecular projection from or towards diopside onto the plane Ol–An–Qz for the pyroxenite compositions shown in Fig. 3. In a) the two solid curves represent the area occupied by primary melts of mantle peridotite KR4003 (Herzberg and O'Hara 2002) and the grey line represents the boundary between peridotite and pyroxenite derived melts given in Herzberg and Asimow (2008). This boundary relies on the fact that $D^{CaO}_{Pyr} > 1$ which is an over-simplification. In b), the stippled, blank and cross-hatched fields represents the regions occupied by primary melts of mantle peridotite leaving a garnet, spinel and harzburgite residues respectively. Note that a significant number of melts of SD pyroxenite have major element compositions that are indistinguishable from some melts derived from mantle peridotite. However, eclogite-derived melts are apparently always too Si-rich to have be derived from peridotite.

Fig. 5. SiO_2/CaO *versus* Mg# for a) peridotite-derived magmas from West Greenland (grey dots) and Iceland rift zones (black dots), and lavas of the Mull Plateau Lava Formation that underwent crystallization at up to 1.6 GPa (white dots). b) near-pure pyroxenite-derived magmas from Hawaii compared to peridotite-dervied magmas from Iceland; Hawaii Scientific Drilling Project 2 (HSDP-2, grey dots) and Koolua (Makapuu Stage; white dots). c) liquid lines of descent of near-primary peridotite- and pyroxenite-derived magmas at pressures of 1 atm, 1.0 GPa and 1.5 GPa. In c), various stars are the estimated near-primary magmas for pyroxenites from Koolau (Makapuu Stage), Koolau Scientific Drilling Project (Sobolev et al. 2005), and HSDP-2 sample SR083-17.5 (Herzberg 2006). Black diamonds are PRIMELT3 model primary magmas to lavas from the Rift Zones of Iceland (Hole and Millett 2016). Grey dots represent the compositions of experimentally derived melts of low SiO_2 pyroxenite 77SL-582 (Keshav et al. 2004). Liquid lines of descent were calculated using Petrolog3 (Danyuchevsky and Plechov 2011) and applying the mineral melt models of Herzberg and O'Hara (2002) for olivine, and Danuchevsky (2001) for plagioclase and clinopyroxene. Melts were assumed to be at the QFM buffer of Kress and Carmichael (1988). The grey line 'Pyr-Per' is the approximate dividing line between the liquid lines of descent of pyroxenite-derived magmas and peridotite-derived magmas at 1 atm.

Fig. 6. Results of MeltPX modelling for pyroxenite lithologies 77SL-582, M5-40 and B-ECL-1 (see Figs) and peridotite KLB-1 at T_P=1450°C. In each case, the pyroxenite lithology makes up 10% of the mantle source and the subsolidus peridotite is assumed to contain a fraction of 0.2 Cpx. Each of the models presented can generate 20km of new basaltic crust. For M5-40 and B-ECL-1 the pressure interval for pyroxenite-only melting is indicated. a) melt productivity (%GPa^{-1}). Horizontal shading is the region of pyroxenite melting and the box shading that for peridotite melting; b) total melt production (wt%) for pyroxenite (solid lines) and peridotite (pecked lines); tc^{Per}/tc which is the proportion of crust generated from peridotite melt relative to the total thickness of crust generated.

Fig. 7. Results of MeltPX modelling for pyroxenite lithology 77SL-582 and peridotite KLB-1 at varying T_P (upper frames) and varying proportion of pyroxenite in the source (lower frames).

Fig. 8. Trace element compositions of peridotite and pyroxenite used in the modelling. Data for perditotie KLB-1 and pyroxenite KG1 are from Lambart (2017)

Fig. 9a Calculated melting and mixing trajectories for Nb/La and Nb/Zr for the model peridotite and pyroxenite compositions shown in Fig. 8, at T_P=1450°C. Isobaric accumulated fractional melts (I-AFM, black diamonds) are melts formed by the melting of pyroxenite and peridotite at the same pressure (depth in the melting column) weighted according to the total volume of melt production at that pressure, as given in the MeltPX parameterizations of Lambart et al. (2016). Isobars (grey lines, indexed in GPa) are mixing lines between pure pyroxenite- and peridotite-derived melts at that pressure. Note that I-AFM melts must also fall along the isobars. Pecked black lines are contours representing the % of peridotite-derived melt in the mixed magma, indexed at the end of each contour. I-AFM at this T_P requires a contribution of more than 60% melt from peridotite. Dotted curves are mixing lines resulting from polybaric mixing of small melt fractions of peridotite with I-AFM melts (see also Fig. 13). b) the same diagram as in a) but for additional T_P of 1350°C and 1500°C. Basalts from the mid-Atlantic Ridge between 57°N and 60°N i.e. those outwith the influence of Icelandic magmatism, all require ≥80% contribution from peridotite-derived melt. The white dot is the intersection of the 0.5 GPa isobar and 80% peridotite melt contour at 1350°C.

Fig. 10. Plots of Nb/La *versus* Nb/Zr showing data for basalts from a) Theistareykir (NVZ) and Snaefellsness (rift-flank); b) WVZ and c) mid-Atlantic Ridge 57-63°N. Pyroxenite, I-AFM and peridotite melting trajectories are shown for 1450°C, as are isobaric mixing lines and are the same as those in fig. 12. The black pecked line represents the contour for 80% contribution from peridotite melt.

Fig. 11. Primitive mantle-normalized (Sun and Mcdonough 1989) multi-element diagrams for a) basalts from the NVZ and WVZ of Iceland that fall at the extreme ends of the arrays in Fig. 12. b) calculated trace element compositions of liquids formed by melting of peridotite at 1.0 GPa and for I-AFM at the pressures indicated. The 3.0 GPa I-AFM melt and the 1.0 GPa peridotite melt are also indicated in a).

Fig. 12. Plots of Nb/La *versus* Nb/Zr showing data for basalts from a) Baffin Island; b) Disko Island and c) Faroe Islands Basalts Group (FIBG) and British Palaeocene Igneous Province (BPIP).

Pyroxenite, I-AFM and peridotite melting trajectories are shown for 1450°C, as are isobaric mixing lines.

Fig, 13. Plots of Nb/La *versus* Nb/Zr showing data for basalts from East Greenland and Jan Mayen Island by stratigraphical unit and location; a) Prince af Wales Bjerge and Kangerlussuaq; b) Sortebrae Lava Field and Jan Mayen Island. Pyroxenite, I-AFM and peridotite melting trajectories are shown for 1450°C, as are isobaric mixing lines.

Fig. 14. Nb/Y *versus* Zr/Y for a) Baffin and Disko islands; b) BPIP Mull Plateau Lavas; c) mid-Atlantic Ridge and Iceland NVZ and d) FIBG Enni Formation. The two dotted lines are the upper and lower bounds of the "Iceland array" of Fitton et al. (1997). Crosses, peridotite melts; filled diamonds, I-AFM; open diamonds pyroxenite melts. Pecked lines are isobars at the pressure indicated. Note that data for Baffin Island and the mid-Atlantic Ridge exhibit similar distributions, but at different pressures, and agree well with the pressures and mixing lines shown in Figs 9 and 11.

Fig. 15. a) Ni (ppm) b) Ca (ppm) and c) Fe/Mn v*ersus* Fo (wt%) for olivines from Theistareykir (grey and white dots), Snaefellesness (black dots) and Mauna Kea low SiO_2 pyroxenite (grey shading). For Theistareykir, high pressure olivines are those which crystallized at ~1 GPa and low pressure olivines crystallized close to the surface (see Hole 2017 for details). Also shown are calculated crystal lines of descent (CLD) for equilibrium olivines crystallizing from model primary magmas from Theistareykir derived from PRIMELT3 at 1 atm, 1.0 GPa and 1.5 GPa. CLDs for estimated near-primary pyroxenite-derived magmas from Hawaii (Herzberg 2006; Sobolev et al. 2005) are shown at 1 atm. Forward crystallization models for near-primary magmas were calculated using Petrolog3 with the following parameters; QFM Buffer of Kress and Carmichael (1988); plagioclase and clinopyroxene equilibria of Danyushevsky (2001). CLD's for Ni were calculated using the method of Herzberg and O'Hara (2002) with modifications from Hole (2017). For Ca, the method of Gevrilenko et al. (2016) was used. The thick grey line labelled '0.55 wt% NiO' represents the 1 atm CLD for pyroxenite melt #368 of Keshav et al. (2004) with a Matzen olivine calculated at 2.5 GPa and 1480°C containing 0.55 wt% NiO. This CLD gives a good approximation to the CLDs for low SiO_2 Hawaiian pyroxenite and is equivalent to a near-primary melt containing 750ppm Ni and ~17 wt% MgO. At the same MgO content, a peridotite-derived magma would be expected to contain ~525 ppm Ni using the calculation method of Herzberg et al. (2016). The two pecked horizontal lines on the Fe/Mn diagram is the expected range of olivine compositions in equilibrium with primary magmas derived from melting of dry peridotite.

Table X – Summary of melting experiments on the pyroxenite lithologies referred to in the text.

Authors	**Starting materials**	**Lithologies**	**Pressure range GPa**	**Temperature range °C**
Keshav et al. 2004	77SL-582; Mix G1	Garnet pyroxenite	2.0-2.5	1280-1495
Kogiso & Hirschmann 2006	B-ECL1; B-ECL-1-OL; MIX G1; G2	Eclogite (Cpx-Gt) Eclogite + Ol	3.0-5.0	1450-1675
Kogiso et al. 1998	KG1; KG2	Mixed peridotite and basalt	1.5-3.0	1300-1525
Kogiso et al. 2003	MIX 1G; G2	Garnet pyroxenite	3.0-7.5	1400-1800
Lambart et al. 2009	M5-130; M7-16	Pyroxenite	1.0-1.5	1165-1450
Lambart et al. 2012	M5-40; M7-16	Pyroxenite	1.0-2.5	1250-1400
Lambart et al. 2013	M5-40; M7-16	Pyroxenite	2.0-2.5	1200-1475
Pertermann & Hirschmann 2003	CRB72-31; GA1; CL G46; SMB6	Eclogite	2.0-3.0	1250-1500
Takahahshi et al. 1998	Columbia River basalts	Basalt	0.0-3.0	1050-1475

Table 1 – End-member compositions of peridotite and pyroxenite

	Depleted Iceland	KLB-1	Pyroxenite this study	KG1	KG2	EPR MORB source
Rb	0.04	0.02	0.6	0.604	0.409	0.02
Ba	0.8	0.227	10.0	9.774	6.591	0.227
Th	0.004	0.004	0.08	0.07	0.048	0.004
U	0.0013	0.0018	0.03	0.0239	0.0165	0.0018
Nb	0.06	0.0864	2.7	3.1082	2.1009	0.0864
La	0.08	0.134	2.2	1.415	0.988	0.134
Ce	0.32	0.42	5.0	4.291	3.001	0.421
Pr	0.08	0.065	0.8	0.67	0.47	0.08
Sr	7.5	6.092	65.0	52.101	36.765	7
Nd	0.6	0.483	4.02	4.429	3.114	0.56
Zr	6	4.269	30.0	34.635	24.513	6.5
Hf	0.18	0.127	0.90	0.918	0.654	0.1792
Sm	0.28	0.21	1.3	1.26	0.91	0.22
Eu	0.125	0.086	0.45	0.571	0.409	0.086
Tb	0.085	0.064	0.3	0.382	0.276	0.061
Dy	0.6	0.471	2.0	2.436	1.781	0.42
Y	4	3.129	12.0	13.925	10.326	2.6
Yb	0.45	0.348	1.4	1.874	1.365	0.32
Lu	0.07	0.056	0.21	0.214	0.161	0.05

Table 2 - Mineral-melt partition coefficients - Peridotite

	Ol	Cpx	Opx	Gt	Sp	DGt	Dsp
Rb	0.005	0.02	0.005	0.07	0	0.0130	0.0066
Ba	0.005	0.02	0.005	0.07	0	0.0130	0.0066
Th	0.005	0.02	0.005	0.07	0	0.0130	0.0066
U	0.005	0.02	0.005	0.07	0	0.0130	0.0066
Nb	0.005	0.02	0.005	0.07	0	0.0130	0.0066
La	0.0004	0.054	0.001	0.01	0.01	0.0057	0.0072
Ce	0.0005	0.098	0.003	0.021	0.01	0.0108	0.0130
Pr	0.0008	0.098	0.003	0.054	0.01	0.0147	0.0133
Sr	0.00019	0.13	0.0007	0.0011	0	0.0102	0.0158
Nd	0.001	0.21	0.0068	0.087	0.01	0.0280	0.0277
Zr	0.01	0.1	0.03	0.32	0	0.0567	0.0258
Hf	0.001	0.22	0.01	0.44	0	0.0700	0.0295
Sm	0.0013	0.26	0.01	0.217	0.01	0.0476	0.0347
Eu	0.0016	0.31	0.013	0.32	0.01	0.0640	0.0417
Tb	0.0015	0.31	0.019	0.75	0.01	0.1147	0.0432
Dy	0.0017	0.33	0.022	1.06	0.01	0.1526	0.0465
Y	0.005	0.5	0.005	2.11	0	0.2618	0.0280
Yb	0.0015	0.28	0.049	4.03	0.01	0.4959	0.0477
Lu	0.0015	0.28	0.06	5.5	0.01	0.6673	0.0507

Table 3 - Mineral-melt partition coefficients - Pyroxenite

	Ol	Cpx	Opx	Gt	D	D[1]
Rb	0.005	0.002	0.022	0.00825	0.003	0.003
Ba	0.005	0.0025	0.013	0.005	0.003	0.006
Th	0.005	0.005	0.01	0.0041	0.005	0.0034
U	0.005	0.0065	0.01	0.023	0.009	0.0077
Nb	0.005	0.008	0.002	0.012	0.009	0.006
La	0.0004	0.0285	0.002	0.0043	0.025	0.023
Ce	0.0005	0.058	0.02	0.0105	0.051	0.047
Pr	0.0008	0.075	0.005	0.012	0.066	
Sr	0.00019	0.0645	0.06	0.013	0.057	0.058
Nd	0.001	0.1345	0.03	0.17	0.140	0.152
Zr	0.01	0.1325	0.18	0.546	0.195	0.188
Hf	0.001	0.245	0	0.447	0.275	0.269
Sm	0.0013	0.2515	0.05	0.356	0.267	0.221
Eu	0.0016	0.217	0.05	0.42	0.248	0.224
Tb	0.0015	0.5	0.019	1.5	0.650	
Dy	0.0017	0.542	0.15	2.83	0.886	0.858
Y	0.005	0.618	0.18	4.12	1.144	1.13
Yb	0.0015	0.769	0.34	8.97	2.000	1.994
Lu	0.0015	0.8275	0.42	10.6	2.293	2.248

1.bulk mineral–melt partition coefficient for garnet pyroxenite from Pertermann et al. (2004).

Table 4 – estimated primitive pyroxenite melt compositions for Iceland and Hawaii

	Jan Mayen	Iceland	Hawaii low SiO_2			Hawaii high SiO_2	
Source	1	1	2	2	2	2	3
SiO_2	45.86	46.10	46.72	47.63	46.60	49.45	51.30
TiO_2	1.85	1.68	1.98	1.85	0.92	1.84	1.50
Al_2O_3	9.42	11.68	10.52	10.54	10.38	10.49	11.30
FeO_T	9.49	9.86	11.76	11.53	11.86	10.49	9.99
MnO	0.15	0.17	0.17				
MgO	20.78	18.28	17.42	17.44	18.44	16.56	15.00
CaO	9.09	9.51	8.62	8.79	8.43	8.87	7.00
Na_2O	1.69	1.76	1.81	1.69	1.83	1.78	2.40
K_2O	1.18	0.55	0.29	0.28	0.28	0.28	0.40
P_2O_5	0.29	0.31	0.16	0.14	0.15	0.16	0.30
Molecular projection coordinates							
Ol	52.1	44.3	42.9	41.5	46.9	37.0	32.6
CaTs	29.5	29.4	26.5	25.4	25.2	24.6	24.4
Qz	18.4	26.3	30.7	33.1	27.9	38.4	43.0
Ol	65.8	55.6	50.4	48.5	56.3	42.4	37.4
An	27.3	31.3	27.9	26.6	19.5	25.3	22.7
Qz	6.9	13.1	21.7	24.9	24.2	32.3	39.9
P (GPa)	3.8	3.3	3.4	3.6	3.8	3.7	3.2

Sources; 1. This study; 2. HSDP-2, Herzberg (2006); 3. Koolau (Makapuu), Sobolev et al. (2005)

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Accumulated fractional melts.

Liquid composition $\frac{C_L}{C_0} = \frac{1}{D_0}(1 - {}^{PF}/_{D_0})^{(\frac{1}{P-1})}$

Residue $\frac{C_S}{C_0} = \frac{1}{(1-F)}(1 - {}^{PF}/_{D_0})^{(\frac{1}{P})}$

Pyroxenite.

P = 3.65 – 1.5 GPa from MeltPX calculations

$P = 0.0002F^2 - 0.0441F + 3.6751$ from MeltPX calculated in 1% increments

C_L/C_0 and C_s/C_0 calculated for 1% increments of melt.
Accumulated fractional melts calculated from the sum of the product of the % melt fraction and C_L/C_0