1 **Phase equilibria and geochemical constraints on the**
2 **petrogenesis of high-Ti picrite from the Paleogene East**
3 **Greenland flood basalt province**

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17 **Abstract**

18 Phase equilibrium experiments have been performed on an extremely high-Ti (5.4 wt.% TiO₂) picrite
19 from the base of the Paleogene (~55 Ma) East Greenland Flood Basalt Province. This sample has a
20 high CaO/Al₂O₃ ratio (1.14), a steep rare-earth elements (REE) profile, is enriched in incompatible
21 trace elements, and is in chemical equilibrium with highly primitive olivine. This all suggests that the
22 picrite is a near-primary melt that did not suffer major chemical evolution during ascent from the
23 mantle source and through the crust. Near-liquidus phase relations were determined over the pressure
24 range of 1 atm, 1 to 1.5 GPa and at temperatures from 1094 to 1400°C. They provide an important
25 constraint on the petrogenesis of these lavas. The high-Ti picritic melt is multi-saturated with olivine
26 (Ol) + orthopyroxene (Opx) at 1 GPa but has only Ol or Opx on the liquidus at lower and higher
27 pressures, respectively. This indicates the primitive melt was last equilibrated with its mantle source
28 at relatively shallow pressure (~ 1 GPa). Melting probably started at 2-3 GPa and the picritic melt
29 was produced by 15-30% melting of the mantle source. Such a degree of partial melting requires a
30 mantle with a high potential temperature (1480-1530°C). The relatively low CaO content and high
31 FeO/MnO ratios of the most primitive East Greenland picrites, the high Ni content of olivine
32 phenocrysts and the presence of low-Ca pyroxene (i.e., pigeonite) at high pressure in our
33 experiments all suggest that the mantle source contained a major component of garnet pyroxenite.
34 Residual garnet in the source could adequately explain the low Al₂O₃ content (7.92 wt.%) and steep
35 REE patterns of the picrite sample. However, simple melting of a lherzolitic source, even with a
36 major pyroxenite component, cannot explain the formation of magmas with the very high Ti contents
37 observed in some East Greenland basalts. We therefore propose that magmas highly-enriched in Ti
38 were produced by melting of a metasomatized mantle source containing Ti-enriched amphibole

39 and/or phlogopite.

40 **Keywords:** Phase equilibria; high-Ti picrite; Flood basalt; Pyroxenite; Mantle metasomatism

1 Introduction

Large igneous provinces (LIP) represent extremely large accumulations of igneous rocks, including plutonic rocks (intrusive) or volcanic rock formations (extrusive). The source of many or all LIPs was variably attributed to mantle plumes or to processes associated with plate tectonics (e.g., [Foulger 2010](#)). LIPs are typically dominated by mafic rocks, with minor expression of ultramafic and/or silicic magmas.

[Cox et al. \(1967\)](#) first described the dichotomy between high-Ti and low-Ti basalts in the Karoo province: high-TiO₂ basalts from the northern Lebombo are enriched in K, Ti, P and other incompatible elements relative to those from the low-Ti basalts from the southern part of the province. Subsequently, this classification based on Ti was then adopted to distinguish basaltic rocks from other LIPs such as Paraná-Etendeka ([Duncan, 1987](#); [Marsh, 1987](#); [Gibson et al., 1996](#); [Mantovani et al., 1985](#)), Siberian Trap ([Ross et al., 2005](#)), North Atlantic Igneous Province ([Saunders et al., 1997](#)) and Queen Maud Land in Antarctica ([Ford and Kistler, 1980](#)), etc. Several petrogenetic models have been advocated to explain the geochemical variability and bimodality of basalts erupted in LIPs: (1) spatially variable (different place) partial melting within an upwelling mantle plume ([Arndt et al., 1993](#); [Campbell and Griffiths, 1990](#)) or lithospheric mantle with a variable asthenospheric component (e.g., [Peate and Hawkesworth, 1996](#); [Piccirillo et al., 1989](#)), followed by variable extents of crustal contamination ([Hawkesworth et al., 1988](#); [Petrini et al., 1987](#)); (2) ‘wet’ melting of a laterally heterogeneous subcontinental lithospheric mantle (SCLM) by heat supplied from an underlying mantle plume ([Gallagher and Hawkesworth, 1992](#)); (3) mixing of picritic mantle plume-derived melts with mafic ultrapotassic melts (lamproitic) as they rise through the SCLM ([Ellam and Cox, 1991](#)); (4) variable partial melting of different mantle source

63 compositions ([Basu et al., 1998](#)). Although the implications of the high-Ti and low-Ti classification
64 include the recognition of distinct petrological processes, source heterogeneity or source
65 contamination of basaltic rocks within the same temporal, spatial and geological setting.

66 Among high- and low-Ti lavas, detailed studies of more primitive high-Mg rocks (e.g., picrite)
67 could provide key constraints on the formation of the dichotomy and potentially LIPs because some
68 compositional features of these magmas—such as low-Al, high Fe, variable Ti, and Ocean Island
69 Basalt (OIB)-like trace element and isotopic signatures—have been considered indicative of
70 generation at high P–T conditions with the involvement of pyroxenite/eclogite components in their
71 mantle sources ([e.g., Gibson et al., 2000; Sobolev et al., 2007; Beccaluva et al., 2009; Sobolev et al.,](#)
72 [2009; Buslov et al., 2010; Kamenetsky et al., 2012; Heinonen et al., 2013, 2014](#)). Furthermore, high-
73 Ti primitive lavas are usually derived from higher temperature or deep source and thus more
74 informative in term of reconstruct the entire evolution processes.

75 High-Ti picritic and basaltic lavas occur in East Greenland ([Hansen et al., 2002](#)). The Ti
76 content in picritic magmas (~5 wt.%) is extremely high and probably higher than those high-Ti lavas
77 in all the other LIPs. This may indicate a unique melting process in the mantle source. Besides, its
78 phase relationships could thus shed new light on mantle melting processes that produce lavas, and
79 especially the Ti-rich ones, erupting in LIPs. Consequently, in this paper, we present a series of phase
80 equilibria experiments on such a high-Ti picritic sample. This rock represents a primitive and
81 perhaps near primary liquid, which has suffered only minor degrees of crystallization or crustal
82 modification during magma ascent.

2 Geological settings and petrography

Palaeogene flood basalts are the dominant geological feature of the central East Greenland between 68° and 70°N (Fig. 1), and cover an area of about 65000 km² (Pedersen et al., 1997) which is mainly covered by pre-Cretaceous rocks. They represent the largest onshore outcrop of Paleogene flood basalts in the North Atlantic region. Previous studies have shown that the North Atlantic flood basalts are the result of the activity of the Paleo-Icelandic plume within the rifting Pangaeon continent (e.g., Brooks, 1973; White and McKenzie, 1989; Saunders et al., 1997). The East Greenland flood basalts are closely associated to other igneous bodies such as dykes, sills and intrusions, and to numerous fault systems related to the evolution of the continental margin (e.g., Karson and Brooks, 1999). Volcanism occurred in three major phases according to Storey et al. (1996) and Tegner et al. (1998), i.e., early volcanic rocks emplaced before 58 Ma, the plateau basalts formed between 56 and 54.5 Ma, and the late volcanics erupted later than 54 Ma. The term 'plateau basalts' here is used as a collective term for the Milne Land, Geikie Plateau, Rømer Fjord and Skrænterne Formations (e.g., Hansen et al., 2002), representing the largest volume of basalts.

In East Greenland, the largest volume of tholeiitic flood basalts occurs in the inland region of Prinsens af Wales Bjerge (Hansen et al., 2002) and contains a distinct high-Ti flow suite, with the rest rocks made of tholeiitic low-Ti flow and more evolved lavas, alkaline basalts, and volcanoclastic sediments, etc. The high-Ti flows can be grouped into a lower suite and an upper suite. The high-Ti rocks in the lower suite are porphyritic, with euhedral to subhedral olivine crystals occurring as phenocrysts, and clinopyroxene and ilmenite being interstitial phases between olivine phenocrysts in the fine-grained volcanic matrix (Hansen et al., 2002). These rocks can be classified as tholeiitic basalts (avg. MgO = 8 wt.%) or picrites (avg. MgO = 13 wt.%) and have TiO₂ contents between 5 and

106 6 wt. %, indicating that the Ti content of the lavas is not correlated to MgO, thus suggesting that
107 these picritic rocks represent real melt composition. The high-Ti picritic rocks also have high $\text{FeO}_{\text{total}}$
108 of 12–13 wt. %, high P_2O_5 contents (0.5–0.6 wt. %) and low CaO (9–10 wt.%) and Al_2O_3 (7–9
109 wt. %), with a CaO/ Al_2O_3 ratio of 1.07–1.26 (avg. 1.1; Hansen et al., 2002). No other tholeiitic rocks
110 in East Greenland with similar MgO-content reach such high- TiO_2 concentrations.

111 One of the most TiO_2 -enriched picrite [sample 436412 from Hansen et al. (2002)] from lower high-
112 Ti flow in Milne Land Formation was selected as the starting material in this study. $^{40}\text{Ar}/^{39}\text{Ar}$
113 degassing method yielded an age of 55 Ma. The sample is one of the most magnesian aphyric rocks
114 and likely represents the primary magmatic liquid composition. The whole-rock TiO_2 content is 5.43
115 wt. % with an $\text{Mg}^\#$ [molar $\text{Mg}^{2+} / (\text{Mg}^{2+} + \text{Fe}^{2+})$] = 0.68 which is in equilibrium with olivine with Fo
116 content 87.6. This value almost equals to the Fo content of the most magnesian olivine phenocryst in
117 the starting material (87.74), suggesting a near primary melts. The major element composition and
118 Ni content (461 ppm) are given in Table 1.

119

120 **3 Experimental methods**

121 *3.1 Starting material*

122 About 200 grams of the high-Ti picrite sample (chips) was cleaned and then crushed and
123 subsequently ground in an agate vibratory disc mill down to grain sizes $<60\mu\text{m}$, and dried overnight.
124 It was then fused to a melt on a platinum (Pt) loop for several times (about 1 g powder every time) in
125 a vertical atmospheric gas-mixing furnace at 1300 °C and equilibrated for 14 hours under oxygen
126 fugacity conditions ($f\text{O}_2$) of the fayalite–magnetite–quartz solid buffer (FMQ). The product then was

127 crushed again to 200 μm mesh and dried overnight at 110 $^{\circ}\text{C}$ before it was used as starting material
128 for the experiments. The composition of the super-liquidus run (NAB-13) is presented in [Table 2](#).

129 3.2 *Low-pressure experiments*

130 Phase equilibria experiments (1 atm) were performed in a vertical Deltech rapid quench gas-mixing
131 furnace in University of California (UC) in Davis, USA. $f\text{O}_2$ was controlled by $\text{H}_2/\text{Ar}-\text{CO}_2$ gas
132 mixtures. Gas flows were regulated with digital Brooks mass flow controllers and a $\text{PtRh}_6\text{-PtRh}_{30}$
133 (type B) thermocouple was used for temperature measurements. The thermocouple has previously
134 been calibrated against the melting point of gold (1064 $^{\circ}\text{C}$). Yttrium-stabilized zirconia electrolyte
135 (SiO_2) was used for the $f\text{O}_2$ measurements inside the furnace. Uncertainties on $\log f\text{O}_2$ values and
136 the temperature do not exceed ± 0.2 and ± 2 $^{\circ}\text{C}$, respectively. Forty to fifty mg of the powder were
137 pressed into a pellet with elvanol. The pellet was sintered to a FePt alloy loop pre-saturated with 7–9
138 wt. % Fe to minimize iron loss between the silicate charge and the wire loop during the experiment
139 ([Grove, 1981](#)). Experimental durations varied from 18 hours at near-liquidus conditions (1354 $^{\circ}\text{C}$) to
140 48 hours at the lowest experimental temperature (1139 $^{\circ}\text{C}$).

141 3.3 *High-pressure experiments*

142 High-pressure equilibrium experiments were carried out under pressures of 1.0, 1.2 and 1.5 GPa and
143 temperatures between 1150 and 1400 $^{\circ}\text{C}$ ([Appendix 1](#)). We used a 0.5-inch end-loaded solid medium
144 piston cylinder apparatus ([Boyd and England, 1960](#)) in UC Davis, USA. About 10 mg of the starting
145 mix were loaded into a graphite crucible (7.93 mm outer diameter, 6.33 mm inner diameter) with a
146 graphite lid and placed in a platinum capsule that had been triple crimped, welded shut, and flattened
147 on one end. Graphite powder was packed in above and below the crucible ([Elkins et al., 2000](#)) and

148 the top of the outer Pt capsule was crimped and welded shut. The capsule was fitted in an alumina
149 sleeve and positioned in the hotspot of a graphite furnace using MgO spacers. BaCO₃ cells wrapped
150 with Pb foil were used as the pressure medium. The pressure was calibrated against the reaction
151 anorthite + gehlenite + corundum = Ca-Tschermak (Hays, 1966), and is accurate to ± 50 MPa.
152 W₉₇Re₃-W₇₅Re₂₅ D-type thermocouples were used to monitor and control temperature, with no
153 correction for pressure effects on thermocouple electromotive force. A 20 °C correction was applied
154 to account for the temperature difference between the sample in the hotspot and the location of the
155 thermocouple. This temperature difference was measured by offset thermocouples. The temperature
156 is thought to be accurate to ± 10 °C. The fO_2 imposed by the graphite capsule in our piston cylinder
157 experimental assembly has not been measured directly, but fO_2 buffering along the C-CO buffer
158 translates to fO_2 conditions between QFM-1.5 and QFM-3.5 under the pressure and temperature
159 conditions of our experiments.

160

161 **4 Analytical methods**

162 Analyses of the experimental products were performed by electron probe micro analyzer (EPMA) at
163 UC Davis. A Cameca SX100 EPMA equipped with five WDS detectors was used. Operating
164 conditions were an accelerating voltage of 15 kV and a 10 nA beam current. We used a focused beam
165 (1 μ m) for minerals and a defocused beam (5–20 μ m) for glasses. The counting times on peaks were
166 10 s for Si, Ti, Al, Fe, Mn, Mg and Ca, and 8 s for the alkalis. Na, K, Si, Ca and Fe were measured
167 first. Spectrometers were calibrated with the following standards for K α X-ray line measurements:
168 albite for Na, orthoclase for K, wollastonite for Si and Ca, TiO₂ for Ti, Fe₂O₃ for Fe, MgO for Mg,

169 Mn₃O₄ for Mn, Cr₂O₃ for Cr and NiO for Ni. Raw data were corrected using the PAP routine. The
170 precision for oxide concentrations was better than 1%. No significant alkali loss (within uncertainty)
171 was detected during measurements.

172

173 **5 Results**

174 *5.1 Attainment of equilibrium*

175 The high diffusivity of major element cations at high temperature in basaltic-picritic systems
176 (Villiger et al 2004) ensures fast equilibration between silicate melts and crystallizing mineral
177 phases. This is especially true for experiments conducted under near-liquidus conditions. Mg/Fe
178 partition coefficients (Kd_{ol-liq}^{Fe-Mg}) are 0.27±0.04 at 1 atm and 0.30±0.02 at 1 GPa (Table 2). These
179 values are in agreement with those commonly reported in the literature (0.30±0.03; Toplis, 2005;
180 Tuff et al., 2005; Takahashi and Kushiro, 1983; Walter, 1998), therefore attesting the equilibrium of
181 the experiment. Moreover, our samples also show evidence of equilibrium based on textural and
182 compositional characteristics: (1) the distribution of mineral phases is homogeneous throughout the
183 samples; (2) the morphology of the crystals does not indicate any quench crystallization (tails or
184 skeletal forms); (3) the minerals and glasses have homogeneous compositions; (4) the crystallization
185 sequence, compositions and proportions of phases in experiments performed at decreasing
186 temperature follow systematic trends. (5) The quality of mass balance calculations performed from
187 all the elements (including Fe) also suggests equilibrium as the sum of squared residuals is low in
188 most cases (<0.7; Appendix 1).

189 *5.2 Phase equilibria*

190 Details of the experimental products and phase proportions are listed in [Appendix 1](#). The
191 crystallizing phases are composed of olivine (Ol), Cr-spinel (Sp), augite (Aug), pigeonite (Pig),
192 orthopyroxene (Opx), plagioclase (Plag) and Fe-Ti oxide including ilmenite (Ilm) and
193 pseudobrookite (Psb). No compositional zoning is observed in the crystals as reflected by
194 microprobe analysis. The determined stability fields of the minerals are plotted as a function of
195 temperature (in the range from 1100 to ~ 1350 °C) and pressure (from 1 atm to 1.5 GPa) in [Fig. 2](#). As
196 shown in the phase diagram, Ol is the liquidus phase at 1 atm. It is joined by orthopyroxene on the
197 liquidus at 1 GPa and is replaced by orthopyroxene at higher pressure (1.2 and 1.5 GPa). At 1 atm,
198 Sp is the second liquidus phase over a wide range of temperature, followed by Aug and Ilm, and
199 eventually Plag joined the liquidus at the lowest temperatures of our experiments. At 1 GPa, the
200 sequence of liquidus phases is Ol + Opx, followed by Pig, Aug, Ilm and Plag. Above 1 GPa, the
201 sequence of phase appearance is Opx followed by Pig and Ilm ([Fig. 2](#)).

202 *5.3 Compositions of liquids and minerals*

203 *5.3.1. Liquids*

204 The chemical compositions of the experimental liquids are reported in [Table 2](#) and presented in
205 [Fig 3](#) where they are compared with published experimental melts produced by partial melting of
206 pyroxenitic and peridotitic sources. [Fig 3](#) illustrates variations in the concentrations of major
207 elements against MgO at different pressure, with temperature variation color-coded. Notably, the
208 super-liquidus compositions at 1 atm have less Na, more Si, less P, less K which could be attributed
209 to loss of alkalis during the experiments. At a given temperature, the incompatible elements have
210 higher concentrations in the high-pressure runs because they have a higher crystallinity than the low-
211 pressure ones. At 1 atm, the SiO₂ content in the liquids stays relatively constant during

crystallization, except for the experiment performed at the lowest temperature (1094°C). At high-pressure conditions (1-1.5 GPa), SiO₂ contents in the liquids are lower and generally decrease gradually with decreasing temperature. The evolution of CaO is similar for all pressure conditions. With decreasing MgO, CaO first increases to a maximum (9-10 wt.%) and then decreases after augite and/or pigeonite saturation. TiO₂ in the melt initially increases and then drops after ilmenite saturation. The maximum TiO₂ concentration is about 7.0, 8.0 and 9.5 wt.% at 1 atm, 1 GPa and 1.5 GPa, respectively. The evolution of FeO_{total} is controlled by saturation in spinel and ilmenite. At 1 atm, FeO_{total} content firstly increases down to 1200°C, temperature at which spinel appears. It then slightly decreases until the saturation of ilmenite. Crystallization of ilmenite results in a stronger decrease of FeO_{total} with crystallization. At higher pressure (1-1.5 GPa), spinel is never a liquidus phase and FeO_{total} contents steadily increase until ilmenite saturation. The higher concentrations are ~ 13.30 of 14.70 and 13.30 wt.%, respectively.

5.3.2. Olivine

The forsterite content of olivine [Fo; molar ratio of 100*Mg/(Mg + Fe)] ranges from Fo₇₆ to Fo₈₇, in 1 atm experiments and from Fo₇₁ to Fo₈₄ in 1 GPa experiments ([Table 2](#)). Only trace amounts of olivine were observed at 1.2 GPa where it has a composition of Fo₈₂. As shown in [Fig. 4a](#), the Fo content of olivine decreases continuously with temperature. At any given temperature it is higher in the 1 atm experiments than in the 1 GPa experiments. This is related to both the lower degree of crystallinity of the 1 atm experiments and their lower FeO content due to co-crystallization of spinel. The relatively constant K_d values in our experiments lead to linear relationships between the Fo contents of olivine and the Mg# values of the equilibrium melts ([Fig. 4b](#)).

233 5.3.3. Pyroxene

234 The compositions of experimental pyroxenes are plotted into the triangular diagram wollastonite
235 (Wo) – enstatite (En) – ferrosilite (Fs) (Fig. 4c) where they are compared with experimental
236 pyroxenes produced by crystallization of ferropicrite under similar pressure conditions (Tuff et al.,
237 2005). Experiments at 1 atm only contain Aug (Wo₃₈₋₃₉En₄₆₋₄₈Fs₁₁₋₁₂). Notably, the Mg# of Aug
238 shows a negative correlation with temperature and Mg# of melts, could be attributed to the co-
239 crystallization of Fe-Ti oxide. At a given temperature or Mg# of melts, higher Mg# of Aug is
240 observed (Appendix 2). Experiments at 1 GPa contain Opx (Wo_{3.8-5.0}En₇₅₋₈₂Fs₁₄₋₂₀), Pig (Wo₁₀₋₂₀En<sub>62-
241 73</sub>Fs₁₅₋₂₄) and Aug (Wo₃₄₋₃₇En₄₈Fs₁₄₋₁₆). We use 6% Wo (wollastonite) content as the distinction
242 between the co-existing Opx and Pig, thus in the Fig. 4c, less data is plotted into the Opx field.
243 Experiments at 1.2GPa contain Opx (Wo₆En₇₇Fs₁₆) and Pig (Wo₁₈En₆₄Fs₁₇). Experiments at 1.5 GPa
244 contain Opx (Wo₅En₇₂₋₈₀Fs₁₅₋₂₃) and Pig (Wo₁₉₋₂₄En₅₃₋₆₄Fs₁₆₋₂₀). The Mg# of Pig and Opx are
245 positively correlated to the Mg# of the equilibrated melt.

246 5.3.4. Plagioclase

247 Plagioclase was observed only in two low-temperature runs at 1 atm (NAB-15) and 1 GPa (PIV-
248 5). Their compositions are reported in Table 2 and plotted onto the triangular diagram An–Ab–Or
249 (Fig. 4d). The plagioclase produced at 1 atm falls in the labradorite field and has a higher An content
250 [molar Ca/(Ca+Na)] than the plagioclase observed at 1 GPa (andesine). This is in agreement with the
251 effect of pressure on the An content of plagioclase (Longhi et al., 1993). However, the change in
252 plagioclase composition in our experiments could also be related to Na loss in the 1 atm experiments
253 (Tormey et al 1987).

254 5.3.5. *Cr-spinel + Fe-Ti oxide*

255 Cr-spinel was only observed in 1 atm experiments below 1230 °C ([Appendix 1](#)). The Cr₂O₃ and
256 Al₂O₃ concentrations of spinel vary from 23.7–33.4 wt.% and 6.1–7.5 wt.%, respectively. Fe-Ti
257 oxide was observed in some experiments performed at 1 atm, 1 GPa and 1.5 GPa. Under 1 atm, in
258 samples NAB-1, NAB-16, and NAB-15, there are pseudobrookite, which are characterized by high
259 TiO₂ content (~65 wt.%), 20 wt.% FeO, 8-9 wt.% MgO, 2 wt.% Cr₂O₃. Two experiments under
260 higher pressure have ilmenite with ilmenite is strongly enriched in TiO₂ (54-55 wt.%) compared to
261 FeO (~35 wt.%) ([Table 2](#)).

262

263 6 Discussion

264 6.1 Comparisons with previous experimental study of high-MgO magmas

265 [Tuff et al. \(2005\)](#) conducted low- to high-pressure experiments (1 atm to 7 GPa) on a ferropicritic
266 composition which is relatively similar to our starting compositions but that has a much lower in Ti
267 content (1.75 wt. TiO₂). Their starting composition has also higher MgO, FeO_{total}, Al₂O₃ and Na₂O
268 contents and lower TiO₂, K₂O and CaO contents than ours (see a detailed comparison in [Table 1](#)).
269 Due to its higher MgO content, the composition investigated by [Tuff et al. \(2005\)](#) has a higher
270 liquidus temperature than our starting composition (50-100°C from 1 atm to 1.5 GPa). Despite these
271 differences, the phase relations of the ferropicrite show similarities with our experimental results
272 ([Fig. 5a](#)). The ferropicrite also has Ol as the first liquidus phase at low pressure (1 atm to 1.5 GPa).
273 Similar to what is observed in our experiments on the high-Ti picrite, olivine is replaced on the
274 liquidus by a pyroxene at higher pressure. However, the change between olivine and pyroxene occurs

275 at higher pressure (between 1.5 and 2 GPa) than for the high-Ti picrite (between 1 and 1.5 GPa).

276 Another major difference between experiments on the high-Ti picrite and the ferropicrite is that the
277 ferropicrite never crystallizes orthopyroxene while orthopyroxene is a major phase crystallizing from
278 the high-Ti picrite. In the ferropicrite, orthopyroxene is replaced by augite. An intuitively simple
279 reason to account for such a difference would be the difference in the CaO contents of the starting
280 compositions. However, the CaO content and the CaO/Al₂O₃ ratio of our starting composition are
281 higher than those of the ferropicrite liquids (Fig. 5b). Similarly, at a given temperature and at any
282 pressure (between 1 bar and 1.5 GPa) our experimental liquids have higher CaO/Al₂O₃ ratios than
283 those produced by crystallization of the ferropicrite. We therefore believe that the CaO contents of
284 the starting compositions is not the only parameter controlling the stability of pyroxenes as already
285 discussed in many previous publications (Grove and Juster 1989; Lindsley 1983; Sack and Ghiorso,
286 1994).

287 6.2 *Liquid line of descent of the high-Ti picrite*

288 A practical way to use our new experiments to better constrain the conditions of formation of
289 flood basalts in East Greenland is to compare experimental liquids with East Greenland mafic lavas.
290 Natural lava compositions have been presented in a number of publications, including Hansen et al.,
291 (2002) and are plotted, along with the experimental liquids in Fig. 3. In these diagrams, our
292 experimental data, and especially those produced at 1 atm, match relatively well with most natural
293 lavas. This indicates that flood basalts probably originate from differentiation of a picritic melt at low
294 pressure. There is however a group of natural samples low TiO₂ and FeO concentrations that is not
295 reproduced by our experiments. These samples correspond to low-Ti flood basalts and probably

296 formed by differentiation of a primitive melt that had a lower TiO₂ content than our starting
297 composition.

298 6.3 Mantle melting conditions

299 When plotted in a MgO-FeO_{total} diagram and considering a $K_D(\text{Fe-Mg})_{\text{ol-liq}}$ of 0.30, all the
300 Urjbergite high-Ti picritic lavas align on a trend in equilibrium with an olivine Fo₈₅₋₈₈. Our starting
301 composition (436412; 12.88% MgO) falls on the Fo₈₈ line. This value is very similar to the
302 composition of the most primitive crystals observed in the natural rock (Fo₈₈). Addition or removal
303 of equilibrium olivine from the melt would produce minor variations in Fe* relative to MgO, and we
304 therefore consider that the melt has not undergone significant modification by olivine fractionation
305 or accumulation since melt generation in the mantle. This is further supported by the comparison
306 with primary melts in Icelandic lavas which have relatively similar MgO contents (13-14 wt.%;
307 [Shorttle et al., 2014](#)).

308 Our experimental results show that the high-Ti picritic liquids are multi-saturated with respect
309 to Ol + Opx at pressures between 1 and 1.5 GPa. Considering that the melt formed by near perfect,
310 fractional melting, this suggests that melting started at shallow depth (2-3 GPa; 65-100 km) if the
311 multiple saturation point records the average pressure–temperature conditions of the melting column
312 ([Asimow and Longhi, 2004](#)). The degree of partial melting (F) that is necessary to produce such a
313 melt is unknown and cannot be constrained from our experiments. However, by analogy with
314 melting experiments on lherzolitic and pyroxenitic mantle sources, it can be suggested that such melt
315 with 13 % MgO at 2-3 GPa can be formed by 15-30 % melting (Baker and Stolper, 1994;
316 Hirschmann et al., 1998; Walter, 1998; Lambart et al., 2009). These values require a mantle hotter

317 than the average upper mantle (Shorttle et al., 2014). To calculate the potential temperature of the
 318 mantle source we correct the average pressure and temperature of picrite genesis to zero pressure
 319 using a mantle adiabatic gradient ($-P[\delta T/\delta P]$) of 13.3 K/GPa. Following Putirka (2007), we corrected
 320 for the effect of latent heat of fusion on the temperature using the expression $\Delta T = F(\Delta H_{\text{fus}}/C_p)$. We
 321 considered a mantle source with a heat capacity (C_p) of 192.4 J/mol.K⁻¹ and $\Delta H_{\text{fus}} = 128.3$ KJ/mol.
 322 We also considered an adiabatic gradient ($-P[\delta T/\delta P]$) of 13.3 K/GPa. Using these values, we
 323 calculate a mantle potential T_p values of 1480-1530°C. This suggests an excess mantle T_p of 150-
 324 200°C assuming a MORB mantle T_p of 1330°C. This in agreement with recent estimates for Iceland
 325 and lower than estimates of T_p for ferropicrites in Karoo (1540-1640°C). This is also in the range of
 326 mantle T_p for mantle plumes beneath Galapagos, Gorgona, Emeishan and Curacao (Herzberg and
 327 Asimow, 2015; Trela et al., 2017).

328 We cannot fully exclude that our starting composition (436412) did not experience any
 329 fractionation on the way from the mantle source to the Earth's surface. If we assumed olivine
 330 fractionation, we can recalculate the composition of a melt in equilibrium with Fo₉₀. Such a melt
 331 would have ~ 18 wt.% MgO and would have been produced in a mantle with a even higher T_p (~
 332 1670°C) than previously calculated.

333 6.4 Presence of pyroxenite in the mantle source of the Early Paleogene Greenland basalt

334 High CaO/Al₂O₃, low SiO₂ and high FeO, TiO₂ contents are essential features of our starting
 335 composition and of high-Ti picrites from East Greenland. One possible explanation for the genesis
 336 these picrites is that they formed by simple partial melting of a lherzolite with olivine left in the
 337 residue. However, the relatively high Ni (461 ppm) and MgO (12.88 wt.%) content of our sample

338 indicates that either that typical olivine-rich mantle lithologies are not sources of parental melts or
339 that the melts have changed their composition after extraction (e.g., [Matzen et al., 2017](#); [Sobolev et](#)
340 [al., 2005](#)).

341 The source of the high-Ti picrite could also have a component of garnet-bearing pyroxenite
342 (e.g., [Petermann and Hirschmann, 2003](#); [Lambert et al., 2013](#)). This may explain the low Al_2O_3
343 content (~ 8 wt. wt.%) of the picrites and their steep HREE patterns. In the following we will test the
344 hypothesis of a mantle source containing a pyroxenite component for East Greenland picrites. If
345 conclusive, we will also try to estimate the contribution of pyroxenite melting to the formation of
346 high-Ti picrites. [Herzberg and Asimow \(2008\)](#) provided computational methods and experimental
347 data to discriminate between melts produced by melting of pyroxenite and lherzolite. These results
348 show that melts derived from a pyroxenite source relative to those derived from a peridotite source
349 have lower CaO contents at a given MgO content ([Herzberg and Asimow, 2008](#); [Petermann and](#)
350 [Hirschmann, 2003](#)). Our starting composition, as shown in [Fig. 6a](#), plots within the field of melts
351 produced by pyroxenite melting. [Herzberg \(2011\)](#) also suggested that the FeO/MnO ratio at a given
352 MgO content is different for melts produced by melting of a peridotite and melts produced by
353 melting of a pyroxenite. FeO/MnO ratios of melts derived from a peridotite range from 50 to 60,
354 whereas FeO/MnO ratios of melts derived from a pyroxenite are > 60 . As shown in [Fig. 6b](#), the
355 FeO/MnO ratio of high-Ti picrites from East Greenland is commonly high (~ 64), suggesting a major
356 pyroxenite component in the mantle source. The $100 \cdot \text{Mn}/\text{Fe}$ values of olivine have also been used to
357 estimate if olivine was in equilibrium with a pyroxenitic or lherzolititic source ([Sobolev et al., 2007](#)).
358 As shown in [Fig. 6c](#), the olivine in sample 436412 are similar to olivine compositions that were in
359 equilibrium with a pyroxenite (e.g., [Sobolev et al., 2007](#)). [Sobolev et al. \(2007\)](#) also provided an

equation to calculate the proportion of pyroxenitic material in the source, which is related to the $100 \times \text{Mn/Fe}$ ratio of the primitive olivine crystals [proportion of pyroxenitic material = $3.48 - 2.071 \times (100 \times \text{Mn/Fe})$]. This equation suggests that pyroxenite melting contributed to 57-65 % of the total melt production which is higher than the values recently calculated for primitive magmas in Iceland (30-40 %; [Shorttle et al., 2014](#)). Although pyroxenites have a high melt productivity due to their low solidus temperature, this requires a mantle source with more than 5 % of pyroxenite ([Lambart et al. 2009](#)) which is more than the average pyroxenite fraction of the upper mantle 2-5 %; [Hirschmann and Stolper, 1996](#)).

6.5 *A role of mantle metasomatism on the formation of high-Ti picrite?*

Mantle metasomatism is a common process that may change the primary mineralogy of mantle sources but also its solidus temperature ([Sarafian et al. 2017](#)). This process is well known to occur in the mantle wedge above subducting slabs ([Grove et al. 2009](#)), but it may also occur in adiabatically decompressing asthenospheric mantle ([Michael, 1988](#)), including mantle plumes ([Dixon et al., 2008](#)). Metasomatic fluids interacting with mafic rocks may lead to the formation of new, hydrous phases, such as amphibole ([Gillis and Meyer, 2001](#)) and may also create mineral assemblages enriched in Al-Fe-Ti as observed in metasomatized mantle xenoliths found in alkaline volcanic rocks ([Bailey, 1970](#)). It has been demonstrated that metasomatism especially by CO₂ rich fluids, can remarkably increase the TiO₂ content in the source (e.g., [Ripp et al., 2006](#); [Dasgupta and Hirschmann, 2007](#); [Berkési et al., 2012](#)).

The TiO₂ contents of both the natural picrites from East Greenland and the experimental liquids produced in this study are far above the fields of melt compositions produced by simple

381 partial melting of pyroxenitic or lherzolitic sources (Fig. 3). However, low-degree partial melting of
382 hydrous mantle sources produces melts that are highly enriched in alkalis and incompatible trace
383 elements (e.g. Dasgupta and Hirschmann, 2007; Walter, 1998; Wendlandt and Harris, 1979; Mysen,
384 1983; Schneider and Eggler, 1984; Hirschmann et al., 2003; Dasgupta and Hirschmann 2007).
385 Melting of mantle parcels enriched in Ti through a metasomatic process could therefore be an
386 important process for the formation of high-Ti picritic magmas.

387 Among the Ti-rich phases that are found in metasomatized mantle rocks, amphibole and
388 phlogopite are the most common. Incongruent melting of a mantle source containing these minerals
389 may therefore significantly enrich the partial melt. Estimating the exact fractions of amphibole and
390 biotite in the source of East Greenland picrites would be too speculative and is beyond the scope of
391 this study. Nevertheless, the trace element contents of East Greenland magmas can be used to discuss
392 whether the source is rather enriched in phlogopite or in amphibole. Melts in equilibrium with a
393 mantle residue containing phlogopite are expected to have relatively higher Rb/Sr (>0.1) and lower
394 Ba/Rb (<20) ratios than melt in equilibrium with an amphibole-bearing source (Furman and Graham,
395 1999). In Fig. 7, high-Ti picritic lavas from East Greenland were plotted. It can be seen that all lavas
396 have low Rb/Sr (<0.1) and the majority also have high Ba/Rb ratios. This seems to indicate that the
397 mantle is rather enriched in amphibole. However, some lavas also have low Ba/Rb ratios (< 20).
398 They may have been produced at slightly lower degrees of melting and may therefore be in chemical
399 equilibrium with a source containing both phlogopite and amphibole.

400 Further study should be performed to constrain the volatile content of the magma, such as the
401 systematic studies on melt inclusions hosted by magnesian olivine which could also give us a better

402 understanding the composition of primitive magma and evolution in crustal-level chambers including
403 degassing processes.

404

405 **7. Conclusions**

406 We performed new high-temperature, low- to high-pressure experiments on a high-Ti, highly
407 primitive picrite from East Greenland. Our experimental results show that the high-Ti picritic liquids
408 are multi-saturated with respect to olivine + orthopyroxene at pressures about 1 GPa. They only
409 olivine (Ol) on the liquidus at lower pressure and only orthopyroxene (Opx) at higher pressure. This
410 suggests that melting occurred at shallow depth if the multiple saturation point records the average
411 pressure–temperature conditions of the near fractional adiabatic decompression melting ([Asimow
412 and Longhi, 2004](#)). Chemical features of the picrites, especially the low Al_2O_3 contents, the steep
413 HREE patterns and the high $\text{CaO}/\text{Al}_2\text{O}_3$ ratios all suggest that the mantle source had a non-negligible
414 highly fusible pyroxenite component. The high-Ti content of the picrites probably cannot be
415 explained by simple fractional crystallization of a dry mantle source. We believe that CO_2 mantle
416 metasomatism prior to melting may have enriched the source in TiO_2 through the formation of
417 amphibole and biotite. Partial melting of such a Ti-rich source could account for the formation of the
418 high-Ti picrites magmas.

419

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425

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Figure captions

Fig. 1. Sketch geological map of East Greenland (Modified from Hansen et al., 2002).

660

661 **Fig. 2.** Phase relations in the experiments as a function of temperature and pressure. Dashed lines represent
 662 estimated liquidus. Ol = olivine; Opx = orthopyroxene; Pig = pigeonite; Aug = augite; Plag = plagioclase; Sp =
 663 spinel; Fe-Ti ox. = Fe-Ti oxide.

664

665
 666 **Fig. 3.** Variation of major elements of the experimental melts against MgO (wt.%). East Greenland lavas (Hansen et
 667 al., 2002) and experimental pyroxenite/peridotite melts (Appendix 3/4) are also plotted for comparison.

668

Fig. 4. Variation of forsterite(Fo) in olivine as a function of temperature **(a)** and Mg# of the equilibrated melt **(b)**. **(c)** Diagram of wollastonite (Wo)–enstatite (En)–forsterite (Fs) showing the compositions of experimental pyroxenes at different pressures. Experimental pyroxenes of 97SB28 (Tuff et al., 2005) are also plotted for comparison; Notably, because we took 5 wt.% CaO as criteria to discriminate pigeonite and orthopyroxene in the runs, thus less data are plotted in the orthopyroxene field. **(d)** Diagram of anorthite (An)– albite (Ab)– orthoclase (Or) showing the composition of plagioclase in two runs (NAB-15 and PIV-5).

677

678 **Fig. 5. (a)** Comparison of phase relations of the high-Ti picrite (color lines; this study) and a ferropicrite (97SB68;
 679 black lines and thick shaded line indicates the liquidus) investigated by Tuff et al. (2005). **(b)** CaO/Al₂O₃ ratio
 680 comparison between experimental results and experimental data from ferropicrite (97SB68) of Tuff et al. (2005).
 681

Fig. 6. Comparison of MgO and FeO* contents of high-Ti picrite 436412 with melts of peridotites KLB-1 (Takahashi, 1986; Takahashi et al., 1993) and PHN-1611 (Kushiro, 1996), olivine websterite HK66 (Takahashi & Kushiro, 1983; Hirose & Kushiro, 1993), quartz eclogite G2 (Pertermann & Hirschmann, 2003a, 2003b) and silica-undersaturated garnet pyroxenite MIX1G (Kogiso et al., 2003). Black circles are compositions of other Urjbergite high-Ti picrites (Hansen et al., 2002).

Fig. 7. (a) MgO (wt.%) versus CaO (wt.%) illustrating the experimental division between peridotite- and pyroxenite-derived melts (Herzberg and Asimow, 2008). **(b)** Bulk-rock geochemical data compilation for the East Greenland flood basalts are showing the involvement of a pyroxenitic source. Data source: *Etendeka*: Ewart et al. (1998, 2004), Marsh et al. (2001), Marsh and Milner (2007), Trumbull et al. (2007); *Siberian Trap*: Fedorenko and Czamanske (1997), Zolotukhin et al. (1991), Wooden et al. (1993), Arndt et al. (1995), Sobolev et al. (2009). *Karoo*: Luttinen and Furnes (2000). *Emeishan* data were from Xu et al. (2001); *East Greenland*: Hansen et al. (2002). **(c)** Olivine chemistry for East Greenland picrites, $100 \times \text{Mn/Fe}$ versus Ni (ppm) illustrates fields for olivine in equilibrium with peridotite and pyroxenite melts (Sobolev et al., 2007). Data of MORB, Hawaii, komatiite and Siberian Trap are from Sobolev et al. (2007). **(d)** Rb/Sr versus Ba/Rb diagram suggest the presence of amphibole in the mantle source for the high-Ti picritic lava.

699 **Table 1.** Compositions (in wt. %) starting material (436412) and a high-MgO ferropicrite (sample
700 97SB68) used in [Tuff et al. \(2005\)](#) for comparison. FeO*=total FeO
701

Sample	SiO ₂	TiO ₂	Al ₂ O ₃	FeO*	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Ni (ppm)	LOI	Total	CaO/Al ₂ O ₃
436412	45.67	5.43	7.92	11.14	0.17	12.88	9.00	1.77	0.83	0.62	461	2.3	99.27	1.14
97SB68	46.91	1.75	9.09	14.90	0.17	14.94	8.08	1.94	0.68	0.19	629	-	100.31	0.89

Table 2. Electron microprobe analyses of run products.

Run no.	Phase ^a	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	NiO	Cr ₂ O ₃	Total	An/Fo/Mg#	Kd
NAB-13	melt (20) ^b	47.95 (0.27) ^c	5.39 (0.17)	8.29 (0.09)	12.69 (0.26)	0.17 (0.06)	13.35 (0.14)	9.00 (0.16)	1.12 (0.08)	0.56 (0.03)	0.32 (0.04)	0.03 (0.03)	0.11 (0.06)	98.98	65	
NAB-11	melt (20)	47.53 (0.48)	5.50 (0.17)	8.26 (0.19)	12.66 (0.32)	0.18 (0.05)	12.29 (0.24)	9.14 (0.20)	1.70 (0.08)	0.74 (0.04)	0.56 (0.06)	0.03 (0.03)	0.12 (0.05)	98.72	64	
	ol (12)	39.96 (0.42)	0.09 (0.03)	0.05 (0.05)	12.62 (0.22)	0.15 (0.05)	45.83 (0.47)	0.28 (0.03)	0.01 (0.01)		0.05 (0.05)	0.33 (0.11)	0.10 (0.14)	99.47	87	0.27
NAB-10	melt (20)	49.45 (0.23)	5.71 (0.17)	9.21 (0.11)	13.00 (0.31)	0.18 (0.04)	10.90 (0.16)	9.54 (0.16)	1.82 (0.07)	0.78 (0.05)	0.63 (0.04)	0.03 (0.03)	0.12 (0.06)	101.39	60	
	ol (9)	38.31 (0.32)	0.13 (0.13)	0.09 (0.15)	14.67 (0.17)	0.21 (0.04)	42.68 (0.56)	0.35 (0.21)	0.02 (0.03)		0.06 (0.03)	0.32 (0.12)	0.15 (0.17)	96.99	84	0.31
NAB-9	melt (20)	48.29 (0.24)	5.81 (0.19)	8.93 (0.11)	12.93 (0.33)	0.20 (0.07)	10.30 (0.14)	9.75 (0.12)	1.71 (0.08)	0.81 (0.04)	0.64 (0.04)	0.02 (0.03)	0.13 (0.06)	99.53	59	
	ol (9)	40.29 (0.44)	0.08 (0.04)	0.06 (0.06)	14.46 (0.29)	0.21 (0.07)	45.18 (0.41)	0.30 (0.04)	0.03 (0.05)		0.04 (0.03)	0.40 (0.14)	0.10 (0.15)	101.14	85	0.26
NAB-8	melt (20)	48.33 (0.22)	5.89 (0.15)	9.25 (0.11)	12.60 (0.25)	0.17 (0.06)	8.45 (0.15)	10.25 (0.17)	1.86 (0.08)	0.87 (0.05)	0.69 (0.04)	0.03 (0.03)	0.08 (0.03)	98.44	55	
	ol (10)	39.07 (0.31)	0.12 (0.09)	0.08 (0.08)	16.18 (0.33)	0.20 (0.06)	43.15 (0.37)	0.30 (0.05)	0.02 (0.03)		0.04 (0.04)	0.29 (0.09)	0.22 (0.30)	99.68	83	0.26
	sp (3)	0.00 (0.00)	11.59 (1.49)	7.52 (0.36)	32.39 (0.99)	0.26 (0.06)	11.28 (0.45)					0.32 (0.16)	32.56 (1.96)	95.92		
NAB-2	melt (20)	48.74 (0.26)	6.27 (0.22)	9.85 (0.16)	12.29 (0.25)	0.21 (0.05)	7.25 (0.12)	10.58 (0.18)	2.09 (0.07)	0.92 (0.05)	0.78 (0.06)	0.03 (0.04)	0.08 (0.05)	99.10	51	
	ol (8)	37.64 (0.59)	0.31 (0.26)	0.29 (0.26)	17.41 (0.95)	0.22 (0.05)	40.29 (0.44)	0.41 (0.15)	0.03 (0.02)		0.06 (0.02)	0.44 (0.17)	0.16 (0.12)	97.25	81	0.28
	sp (2)	0.00 (0.00)	12.06 (4.44)	6.88 (0.68)	33.45 (2.15)	0.15 (0.03)	10.13 (0.26)					0.38 (0.10)	32.63 (4.83)	95.69		
NAB-5	melt (20)	48.80 (0.33)	6.41 (0.14)	10.08 (0.09)	12.15 (0.30)	0.19 (0.05)	6.76 (0.12)	10.43 (0.18)	2.09 (0.07)	0.97 (0.05)	0.78 (0.05)	0.03 (0.04)	0.07 (0.05)	98.76	50	
	ol (9)	38.42 (0.37)	0.15 (0.07)	0.06 (0.06)	18.57 (0.43)	0.20 (0.07)	41.51 (0.60)	0.36 (0.04)	0.02 (0.02)		0.04 (0.04)	0.30 (0.10)	0.12 (0.12)	99.74	80	0.26
	aug (18)	51.05 (0.63)	1.45 (0.21)	1.84 (0.41)	7.56 (0.66)	0.17 (0.07)	16.68 (0.62)	19.15 (0.61)	0.27 (0.05)		0.25 (0.03)	0.07 (0.05)	0.38 (0.26)	98.86	80	
	sp (4)	0.00 (0.00)	11.63 (2.24)	6.77 (0.40)	34.75 (2.24)	0.26 (0.04)	9.75 (0.08)					0.42 (0.16)	33.39 (3.46)	96.96		
NAB-3	melt (20)	49.06 (0.30)	6.82 (0.15)	10.45 (0.15)	12.13 (0.25)	0.18 (0.06)	6.82 (0.10)	10.03 (0.19)	2.33 (0.09)	0.99 (0.06)	0.82 (0.05)	0.03 (0.03)	0.07 (0.05)	99.72	50	
	ol (4)	38.87 (0.60)	0.26 (0.11)	0.27 (0.26)	19.10 (0.80)	0.25 (0.02)	41.25 (0.51)	0.36 (0.06)	0.06 (0.12)		0.02 (0.01)	0.29 (0.05)	0.17 (0.14)	100.89	80	0.27
	aug (20)	50.78 (0.83)	1.71 (0.38)	2.35 (0.61)	7.90 (0.40)	0.15 (0.08)	16.45 (0.56)	19.38 (0.57)	0.31 (0.05)		0.25 (0.05)	0.04 (0.04)	0.32 (0.13)	99.63	79	
	sp (1)	0.00	14.87	6.10	38.34	0.18	8.94					0.35	26.71	95.49		
NAB-1	melt (20)	48.65 (0.32)	6.56 (0.19)	11.35 (0.22)	12.07 (0.30)	0.17 (0.07)	5.72 (0.13)	9.47 (0.14)	2.51 (0.12)	1.13 (0.05)	0.88 (0.06)	0.03 (0.02)	0.04 (0.03)	98.58	46	
	ol (9)	38.65 (0.43)	0.16 (0.12)	0.17 (0.16)	20.72 (0.59)	0.25 (0.06)	39.78 (0.75)	0.30 (0.06)	0.01 (0.01)		0.05 (0.03)	0.35 (0.11)	0.05 (0.04)	100.50	78	0.26
	aug (21)	50.41 (0.82)	1.55 (0.29)	2.20 (0.64)	7.67 (0.80)	0.19 (0.06)	16.39 (0.67)	19.23 (0.08)	0.30 (0.05)		0.24 (0.03)	0.05 (0.04)	0.26 (0.20)	98.48	79	
	psb (1)	0.10	65.18	1.89	19.01	0.11	8.49					0.00	2.21	96.98		
	sp (1)	0.00	11.75	6.62	36.71	0.30	8.60					0.24	31.48	95.71		
NAB-16	melt (9)	49.56 (0.46)	5.31 (0.17)	12.49 (0.21)	11.91 (0.31)	0.15 (0.03)	4.98 (0.16)	8.75 (0.15)	2.69 (0.11)	1.25 (0.07)	1.07 (0.06)	0.03 (0.03)	0.04 (0.03)	98.22	43	
	ol (4)	37.73 (0.36)	0.12 (0.04)	0.02 (0.03)	21.55 (0.28)	0.28 (0.04)	39.16 (0.38)	0.31 (0.04)	0.03 (0.03)		0.03 (0.03)	0.29 (0.04)	0.02 (0.03)	99.56	77	0.24
	aug (7)	51.33 (0.72)	1.80 (0.26)	2.20 (0.23)	7.47 (0.48)	0.16 (0.09)	16.86 (0.45)	19.57 (0.50)	0.31 (0.06)		0.22 (0.02)	0.05 (0.08)	0.33 (0.08)	100.31	80	
	psb (5)	0.10 (0.03)	63.74 (0.08)	1.79 (0.04)	23.37 (0.41)	0.10 (0.07)	7.26 (0.15)					0.08 (0.15)	1.42 (0.12)	97.86		

NAB-15	sp (2)	0.00 (0.00)	12.29 (2.74)	6.50 (0.31)	47.05 (3.58)	0.30 (0.12)	7.67 (0.47)					0.29 (0.23)	23.74 (6.05)	97.84		
	melt (7)	52.39 (0.54)	3.63 (0.19)	13.04 (0.31)	10.90 (0.41)	0.16 (0.08)	4.25 (0.32)	7.51 (0.24)	2.92 (0.25)	1.74 (0.10)	1.44 (0.10)	0.02 (0.02)	0.03 (0.03)	98.02	41	
	ol (5)	37.47 (0.14)	0.09 (0.02)	0.04 (0.02)	22.35 (0.38)	0.33 (0.04)	38.42 (0.20)	0.24 (0.02)	0.01 (0.00)		0.05 (0.03)	0.29 (0.06)	0.02 (0.02)	99.31	76	0.24
	aug (6)	51.70 (0.47)	1.48 (0.07)	1.60 (0.18)	7.50 (0.91)	0.18 (0.07)	17.32 (0.54)	19.12 (0.52)	0.30 (0.06)		0.25 (0.04)	0.03 (0.04)	0.28 (0.19)	99.76	81	
	plag (3)	53.87 (0.48)	0.31 (0.09)	27.29 (0.32)	1.51 (0.09)	0.05 (0.06)	0.31 (0.18)	10.70 (0.23)	4.51 (0.14)	0.48 (0.09)	0.24 (0.10)	0.02 (0.03)	0.03 (0.03)	99.32	42	
PIV-8	psb (4)	0.10 (0.04)	60.81 (0.66)	1.50 (0.12)	28.57 (0.57)	0.11 (0.03)	6.11 (0.19)					0.06 (0.09)	0.80 (0.13)	98.07		
	melt (10)	47.19 (0.21)	5.59 (0.17)	8.21 (0.11)	12.54 (0.31)	0.17 (0.08)	12.86 (0.14)	8.86 (0.16)	1.85 (0.10)	0.76 (0.04)	0.66 (0.06)	0.02 (0.05)	0.10 (0.06)	98.83	65	
PIV-10	melt (10)	46.57 (0.15)	5.68 (0.13)	8.78 (0.12)	13.14 (0.31)	0.16 (0.07)	11.50 (0.17)	9.43 (0.18)	1.95 (0.04)	0.82 (0.04)	0.67 (0.03)	0.02 (0.03)	0.12 (0.06)	98.85	61	
	ol (3)	39.44 (0.28)	0.11 (0.01)	0.03 (0.03)	15.51 (0.17)	0.22 (0.05)	44.84 (0.08)	0.27 (0.06)	0.00 (0.00)		0.07 (0.03)	0.04 (0.04)	0.07 (0.03)	100.60	84	0.30
	opx (5)	55.65 (0.63)	0.61 (0.07)	1.18 (0.17)	9.37 (0.41)	0.15 (0.06)	31.61 (0.49)	2.05 (0.13)	0.06 (0.01)		0.02 (0.01)	0.05 (0.07)	0.27 (0.06)	101.02	86	
PIV-2	melt (10)	46.43 (0.13)	6.05 (0.17)	9.09 (0.09)	12.65 (0.42)	0.20 (0.07)	11.20 (0.16)	9.64 (0.15)	1.93 (0.05)	0.83 (0.04)	0.74 (0.04)	0.01 (0.02)	0.09 (0.06)	98.86	61	
	ol (4)	39.68 (0.35)	0.11 (0.04)	0.06 (0.01)	16.17 (0.08)	0.18 (0.03)	44.89 (0.33)	0.27 (0.02)	0.01 (0.01)		0.04 (0.02)	0.04 (0.03)	0.10 (0.06)	101.55	83	0.32
	opx (5)	55.81 (0.41)	0.65 (0.04)	1.43 (0.04)	9.65 (0.17)	0.20 (0.05)	31.02 (0.37)	2.27 (0.14)	0.07 (0.03)		0.03 (0.02)	0.06 (0.11)	0.35 (0.03)	101.54	85	
PIV-1	melt (10)	45.30 (0.28)	7.11 (0.13)	10.41 (0.19)	13.02 (0.14)	0.19 (0.05)	8.69 (0.38)	9.49 (0.23)	2.28 (0.10)	1.04 (0.05)	0.88 (0.06)	0.01 (0.02)	0.06 (0.05)	98.47	55	
	ol (3)	39.24 (0.19)	0.10 (0.03)	0.05 (0.01)	17.87 (0.93)	0.20 (0.03)	42.52 (0.92)	0.27 (0.04)	0.02 (0.01)		0.02 (0.02)	0.07 (0.06)	0.09 (0.03)	100.44	81	0.28
	opx (5)	54.05 (0.65)	0.82 (0.18)	1.73 (0.45)	10.72 (0.11)	0.30 (0.15)	29.50 (1.07)	2.48 (0.82)	0.18 (0.02)		0.03 (0.01)	0.16 (0.09)	0.40 (0.09)	100.38	83	
	pig (3)	53.36 (0.59)	1.08 (0.11)	2.22 (0.35)	9.72 (0.50)	0.16 (0.05)	23.04 (0.33)	10.38 (1.20)	0.29 (0.02)		0.13 (0.03)	0.00 (0.00)	0.47 (0.05)	100.84	81	
PIV-3	melt (10)	45.64 (0.26)	6.79 (0.18)	9.99 (0.09)	13.15 (0.29)	0.19 (0.04)	9.30 (0.14)	9.88 (0.06)	2.25 (0.08)	0.95 (0.04)	0.82 (0.04)	0.01 (0.01)	0.10 (0.04)	99.08	56	
	ol (4)	39.36 (0.29)	0.16 (0.07)	0.04 (0.01)	17.99 (0.26)	0.21 (0.08)	43.00 (0.27)	0.35 (0.05)	0.04 (0.06)		0.05 (0.04)	0.03 (0.06)	0.08 (0.06)	101.32	81	0.30
	pig (6)	54.49 (0.73)	0.68 (0.11)	1.33 (0.33)	10.66 (0.27)	0.22 (0.04)	27.25 (0.61)	5.43 (0.58)	0.22 (0.06)		0.04 (0.03)	0.07 (0.07)	0.31 (0.07)	100.71	82	
PIV-4	melt (10)	44.45 (0.28)	7.95 (0.12)	11.41 (0.14)	13.03 (0.30)	0.18 (0.05)	7.26 (0.14)	9.01 (0.14)	2.60 (0.10)	1.15 (0.04)	0.99 (0.05)	0.01 (0.02)	0.04 (0.04)	98.08	50	
	ol (4)	38.65 (0.06)	0.14 (0.05)	0.06 (0.02)	21.40 (0.46)	0.23 (0.03)	39.98 (0.11)	0.34 (0.07)	0.04 (0.06)		0.04 (0.05)	0.14 (0.06)	0.01 (0.01)	101.03	77	0.30
	opx (4)	53.45 (0.12)	1.14 (0.11)	2.82 (0.09)	12.51 (0.48)	0.23 (0.06)	27.42 (0.26)	2.56 (0.19)	0.10 (0.02)		0.03 (0.03)	0.08 (0.09)	0.32 (0.09)	100.65	80	
	pig (4)	51.72 (0.62)	1.38 (0.19)	3.15 (0.56)	11.16 (0.12)	0.19 (0.06)	22.28 (0.29)	9.05 (0.24)	0.32 (0.02)		0.09 (0.00)	0.05 (0.04)	0.34 (0.04)	99.75	78	
PIV-5	aug (4)	51.24 (0.89)	1.41 (0.10)	1.66 (0.34)	10.16 (0.90)	0.24 (0.07)	17.33 (0.99)	17.02 (0.60)	0.29 (0.07)		0.25 (0.06)	0.13 (0.13)	0.09 (0.06)	99.83	75	
	melt (5)	47.94 (1.13)	5.16 (0.27)	14.30 (0.54)	11.35 (0.31)	0.15 (0.03)	4.75 (0.47)	6.68 (0.16)	3.03 (0.17)	2.42 (0.10)	2.12 (0.19)	0.03 (0.03)	0.05 (0.06)	97.98	43	
	ol (4)	37.52 (0.51)	0.15 (0.03)	0.07 (0.10)	26.00 (0.52)	0.25 (0.04)	35.54 (0.53)	0.30 (0.04)	0.01 (0.01)		0.04 (0.04)	0.24 (0.13)	0.01 (0.01)	100.13	71	0.31
	aug (4)	51.00 (0.93)	1.58 (0.36)	2.04 (0.55)	8.93 (1.57)	0.22 (0.05)	16.92 (0.67)	18.10 (1.37)	0.29 (0.06)		0.20 (0.05)	0.05 (0.06)	0.21 (0.14)	99.54	77	
	pig (1)	52.24	0.99	2.44	14.65	0.27	21.92	6.66	0.29		0.08	0.24	0.07	99.84	73	
	plag (4)	57.23 (0.42)	0.34 (0.14)	25.73 (0.66)	1.33 (0.53)	0.00 (0.01)	0.42 (0.52)	7.83 (0.28)	5.83 (0.08)	1.45 (0.16)	0.16 (0.07)	0.00 (0.00)	0.04 (0.04)	100.36	55	
	ilm (3)	0.07 (0.04)	54.00 (0.85)	0.59 (0.03)	35.56 (0.16)	0.35 (0.08)	7.95 (0.25)					0.20 (0.09)	0.51 (0.10)	99.23		
PIV-12	melt (10)	46.83 (0.26)	5.67 (0.15)	8.17 (0.08)	12.67 (0.31)	0.20 (0.06)	12.97 (0.19)	8.85 (0.11)	1.80 (0.09)	0.75 (0.04)	0.63 (0.06)	0.02 (0.03)	0.12 (0.04)	98.69	65	

PIV-9	melt (10)	45.47 (0.17)	6.14 (0.12)	9.26 (0.12)	13.53 (0.26)	0.15 (0.09)	10.75 (0.13)	9.67 (0.18)	2.03 (0.10)	0.90 (0.04)	0.71 (0.03)	0.02 (0.03)	0.09 (0.06)	98.73	59
	opx (5)	55.08 (0.34)	0.61 (0.16)	1.92 (0.29)	10.11 (0.33)	0.18 (0.05)	30.20 (0.14)	2.56 (0.14)	0.12 (0.01)		0.06 (0.04)	0.05 (0.06)	0.30 (0.09)	101.20	84
PIV-11	melt (10)	44.26 (0.40)	6.93 (0.19)	10.18 (0.15)	13.77 (0.38)	0.19 (0.05)	9.00 (0.14)	9.07 (0.21)	2.73 (0.09)	1.04 (0.05)	0.79 (0.07)	0.05 (0.05)	0.05 (0.03)	98.05	54
	pig (5)	52.09 (0.52)	1.06 (0.10)	3.02 (0.20)	9.91 (0.33)	0.19 (0.03)	22.97 (0.74)	9.38 (0.73)	0.47 (0.04)		0.11 (0.02)	0.05 (0.07)	0.36 (0.06)	99.60	81
	opx (5)	54.07 (0.50)	0.71 (0.10)	2.14 (0.49)	10.78 (0.25)	0.14 (0.08)	29.03 (0.41)	2.61 (0.20)	0.16 (0.04)		0.04 (0.04)	0.10 (0.14)	0.28 (0.06)	100.06	83
PIV-6	melt (10)	43.23 (0.28)	8.46 (0.20)	11.25 (0.12)	14.64 (0.41)	0.18 (0.04)	7.55 (0.25)	8.42 (0.16)	2.60 (0.29)	1.26 (0.05)	1.00 (0.05)	0.03 (0.03)	0.05 (0.04)	98.66	48
	pig (6)	52.19 (0.82)	1.30 (0.19)	4.03 (0.62)	10.69 (0.82)	0.15 (0.09)	20.93 (1.11)	10.14 (1.50)	0.62 (0.12)		0.14 (0.05)	0.13 (0.11)	0.27 (0.05)	100.59	78
	opx (2)	52.64 (0.49)	0.94 (0.04)	3.23 (0.35)	13.09 (0.22)	0.13 (0.10)	27.68 (0.25)	2.55 (0.24)	0.15 (0.00)		0.04 (0.00)	0.09 (0.13)	0.21 (0.13)	100.75	79
PIV-7	melt (10)	43.15 (0.20)	9.39 (0.22)	12.26 (0.14)	14.53 (0.26)	0.15 (0.05)	5.81 (0.22)	7.35 (0.11)	3.23 (0.09)	1.80 (0.09)	1.29 (0.05)	0.04 (0.08)	0.02 (0.03)	99.00	42
	pig (6)	50.11 (0.39)	2.08 (0.21)	5.99 (0.58)	11.66 (0.58)	0.23 (0.03)	17.88 (0.59)	11.20 (0.58)	0.90 (0.07)		0.19 (0.13)	0.11 (0.10)	0.16 (0.05)	100.50	73
	opx (5)	51.90 (0.59)	1.19 (0.26)	3.50 (0.69)	14.62 (0.43)	0.18 (0.07)	25.90 (0.54)	2.50 (0.35)	0.19 (0.05)		0.04 (0.05)	0.04 (0.07)	0.12 (0.03)	100.17	76
	ilm (4)	0.08 (0.06)	54.54 (0.78)	0.85 (0.01)	34.63 (0.78)	0.21 (0.02)	8.81 (0.27)					0.02 (0.02)	0.43 (0.04)	99.58	
PIV-13	melt (7)	44.46 (0.26)	7.90 (0.09)	10.72 (0.08)	13.41 (0.22)	0.19 (0.03)	8.43 (0.08)	9.45 (0.09)	2.40 (0.05)	1.17 (0.02)	0.87 (0.06)	n/a	n/a	99.00	53
	pig (16)	52.73 (0.82)	1.15 (0.20)	2.68 (0.40)	10.50 (0.89)	0.21 (0.02)	22.62 (1.31)	8.98 (2.05)	0.36 (0.06)		0.00 (0.00)	0.02 (0.02)	0.32 (0.05)	99.51	79
PIV-14	melt (7)	45.99 (0.09)	6.65 (0.08)	9.42 (0.04)	12.70 (0.14)	0.19 (0.03)	10.29 (0.06)	9.99 (0.06)	2.06 (0.04)	0.96 (0.02)	0.69 (0.04)	n/a	n/a	98.94	59
	ol (1)	38.74	0.08	0.03	16.04	0.21	41.95	0.36	0.02		n/a	n/a	0.09	97.52	82
	opx (16)	54.05 (0.78)	0.75 (0.31)	1.69 (0.54)	10.17 (0.30)	0.18 (0.02)	28.32 (1.35)	3.29 (0.95)	0.13 (0.07)		0.00 (0.00)	0.00 (0.01)	0.29 (0.04)	98.77	83

^a Abbreviation: ol-olivine; sp-spinel; aug-augite; opx-orthopyroxene; pig-pigeonite; plag-plagioclase; ilm-ilmenite; psb- pseudobrookite

^b Number of microprobe analyses

^c One standard deviation of replicate analyses in terms of least unit cited

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Appendix

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Appendix 1. Experimental conditions and phase assemblages.

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Appendix 2. Variation in Mg# of Aug, Pig and Opx as a function of temperature (°C) (a, b, c,

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respectively) and Mg# of melt (c, d, e, respectively). Mg# is calculated using the total FeO content of

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FeO*.

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Appendix 3. Composition of pyroxenite partial melts in the experiments.

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Appendix 4. Compositions of experimental peridotite partial melts.