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1 **Emplacement and high-temperature evolution of gabbros of the**

2 **16.5 ºN oceanic core complexes (Mid-Atlantic Ridge): insights into the**

3 **compositional variability of the lower oceanic crust**

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20 **Key Points:**

21 • Lower crust at the 16.5ºN region of the Mid-Atlantic Ridge has a bimodal
22   composition.

23 • Thermodynamic models show that this distribution is expected in regions where
24   small gabbroic bodies are intruded into mantle peridotites.

25 • Deformation related to detachment faulting commenced under high-temperature
26   conditions (>1000 ºC) as recorded by ultramylonites
27

28 **ABSTRACT**
29

30 This study reports the composition of the oceanic crust from the 16.5ºN region of the Mid-

31 Atlantic Ridge (MAR), a spreading ridge segment characterized by a complex detachment

32 fault system and three main oceanic core complexes (Southern, Central and Northern OCC).

33 Lithologies recovered from the core complexes include both greenschist facies and weathered

34 pillow basalt, diabase, peridotite and gabbro, while only weathered and fresh pillow basalt

35 was dredged from the rift valley floor. The gabbros are compositionally bimodal, with the

36 magmatic crust in the region formed by scattered intrusions of chemically primitive plutonic

37 rocks (i.e. dunites and troctolites), associated with evolved oxide-bearing gabbros. We use

38 thermodynamic models to infer that this distribution is expected in regions where small

39 gabbroic bodies are intruded into mantle peridotites. The occurrence of ephemeral magma

40 chambers located in the lithospheric mantle enables large proportions of the melt to be

41 erupted after relatively low degrees of fractionation. A large proportion of the dredged

42 gabbros reveal evidence for deformation at high-temperature conditions. In particular,

43 chemical changes in response to deformation and the occurrence of very high-temperature

44 ultramylonites (>1000 ºC) suggest the deformation related to the oceanic detachment

45 commenced at near-solidus conditions. This event was likely associated with the expulsion of

46 interstitial, evolved magmas from the crystal mush, a mechanism that enhanced the formation

47 of disconnected oxide-gabbro seams or layers often associated with crystal-plastic fabrics in

48 the host gabbros. This granulite-grade event was soon followed by hydrothermal alteration

49 revealed by the formation of amphibole-rich veins at high-temperature conditions (~900ºC).

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51

52 **1. INTRODUCTION**

53   Detachment faults, first reported as low-angle normal faults at the MARK area of the MAR

54 (Dick et al., 1981, Karson and Dick, 1983), are common at slow-spreading ridges. These long-

55 lived faults terminate at the foot of the rift valley wall, and produce a smooth surface with

56 corrugations parallel to the rifting direction (Cann et al., 1997). They expose large sections of

57 lower oceanic crust and abyssal mantle, commonly termed "megamullions" or oceanic core

58 complexes (OCCs) (Cann et al., 1997; Tucholke et al., 1998). The detachment faults operate

59 for up to several million years and exhume lower crust and mantle rocks, which are

60 continuously accreted and emplaced into the rift mountains from beneath a volcanic carapace

61 (Dick et al., 2008). Detachment faulting can account for 60–100% of the plate separation in

62 some ridge sections and may dominate a region for several millions of years (Tucholke et al.

63 1998; Smith et al., 2006, 2008; Schroeder et al., 2007; Baines et al., 2008; Escartín et al., 2008;

64 MacLeod et al., 2009). In particular, Escartín et al. (2008) and Smith et al. (2008) estimated

65 that close to 50 % of the Mid-Atlantic Ridge (MAR) from 12.5ºN and 35ºN is characterized by

66 detachment faults, indicating that a substantial portion of the seafloor in the central portion of

67 the MAR may form by such asymmetric spreading. Recently, some investigators have

68 suggested the proportion of seafloor formed by detachment faulting could be even higher, if

69 more fault surfaces are hidden under basaltic rafted blocks (e.g., Reston and Ranero, 2011).

70 Detachment faulting is now known as a major mechanism of crustal accretion at slow-

71 spreading ridges (Smith et al., 2014; Parnell-Turner et al., 2018).

72   The nucleation and growth of detachment faults is dependent on many factors, including

73 rheological differences between gabbros and serpentinized mantle peridotites, spreading rate,

74 and magma budget (e.g., Cannat et al., 2006; Ildefonse et al., 2007; Escartin et al., 2008). For

75 instance, numerical modeling suggests that detachment faults form at intermediate levels of

76 magma supply (e.g.: Cannat et al., 2006) at slow- and ultraslow-spreading ridges when the

77 ratio of magmatic accretion to tectonic extension (M) lies between 0.3 and 0.5 (Tucholke et

78 al., 2008; Olive et al., 2010; Hansen et al. 2011). This is consistent with the common

79 occurrence of gabbroic lithologies exposed along the fault surfaces at oceanic core complexes

80   (e.g.: Dick et al., 1981; 1991; Cann et al., 1997; Dick et al, 2000; Blackman et al., 2007;

81   2011; MacLeod et al., 2009). In addition, it is likely that the mechanics of detachment

82   faulting in amagmatic environments differs from magmatic ones, where the fault roots into a

83   significant crystal mush zone beneath the rift valley and synmagmatic deformation can

84   enhance fault longevity and control fault geometry (Dick et al., 2000; Natland and Dick,

85   2001).

86   OCCs offer a unique opportunity to unravel the composition of the lower crust of ridge

87   segments formed at slow spreading ridges, although these observations are pertinent to the

88   sections exposed through detachment faulting; to what extent these observations apply to the

89   lower oceanic crust at symmetrically-spreading segments is uncertain. The large

90   compositional variability of the gabbros at OCCs worldwide suggests that the lower crust in

91   these sections is not that of the classic Penrose model (Conference Participants, 1972). At the

92   660-km^2-gabbro massif at Atlantis Bank on the SW Indian Ridge ((I)ODP Holes 735B,

93   1105A and U1473) most of the gabbros recovered by seafloor sampling and drilling (up to

94   1.5 km) are olivine-gabbros interspersed with highly evolved (oxide-bearing) gabbros,

95   representing crystallization from nothing close to a primary magmatic liquid (Dick et al.,

96   2000; MacLeod et al., 2017). In contrast, the 1.4 km deep Hole U1309D at the Atlantis

97   Massif OCC, another large OCC exposed along the Mid-Atlantic Ridge, contains a much

98   broader range of lithologies, with large amounts of primitive gabbros such as troctolite, many

99   gabbro sensu-stricto (olivine <5 vol.%) and subordinate oxide-gabbros (>2% oxide)

100   (Blackman 2006, 2010). Large amounts of troctolites and olivine gabbros also constitute the

101   isolated intrusions in mantle peridotites at the Kane Megamullion, a large detachment system

102   south of the Kane Fracture Zone in the Atlantic (Dick et al., 2008), which nonetheless differ

103   from the Atlantis Massif by the absence of olivine-free gabbro but show a high amount of

104   oxide gabbro. Finally, ODP Leg 209 drilled a series of holes from 14°43' to 15°44'N

105   straddling the 15°20' Fracture Zone, providing a view of the lithostratigraphy to the south of

106   the 16.5º N core complexes. Notably, gabbro is rare, and the lower crust has a "bimodal"

107   distribution similar to that of the Kane area, with large amounts of troctolite and dunite and

108   oxide gabbro and oxide gabbronorite the most abundant lithologies. As a whole, the

109   compositional variability of gabbros recovered at OCCs suggests a variable, complex crustal

110   architecture, which emphasizes the need to characterize the lower crust at ridge segments

111   formed at different magma supplies, ridge geometries and tectonics, as part of interpreting

112   crustal petrogenesis.

113   This study aims at exploring the architecture and the composition of the lower crust

114   exposed along the 16.5ºN region of the MAR. This segment is characterized by a complex

115   detachment fault system developed in a magma-poor regime: unlike the thick gabbroic

116   sections exposed at the Atlantis massif and Atlantis Bank OCCs, oceanic core complexes

117   identified in the 16.5ºN MAR region are formed by scattered gabbroic intrusions within a

118   partly serpentinized mantle (Cannat and Casey, 1995, Cannat et al. 1997. This region is

119   characterized by a strong lateral crustal heterogeneity and an active detachment faulting

120   system (Smith et al., 2014; Parnell-Turner et al. 2016). As such, the MAR 16.5ºN is similar

121   to the Kane and the 15°20' Fracture Zones, and is an ideal case to explore the petrological

122   processes driving the formation of the lower crust in regions of low magma supply and active

123   detachment. Here we present microtextural observations, major element mineral

124   compositions and discuss the process of emplacement and early deformation histories of

125   these rocks. We use thermodynamic modeling to explore the cause of the bimodal

126   lithological variability of the lower crust in this region, proposing that a close relationship

127   between magma supply, eruption rate and deformation-driven melt extraction may explain

128   the lithological variability of the lower oceanic crust exposed through detachment faulting.

129   **2.   THE 16.5 ºN REGION OF THE MID-ATLANTIC RIDGE**

130   The study area is located between 16º and 17ºN and exposes an ~120 km stretch of the

131   slow-spreading Mid-Atlantic ridge (Fig. 1). This region encompasses two spreading segments

132   that strike ~012º, perpendicular to the inferred spreading direction (DeMets et al., 2010). The

133   spreading axis is characterized by a well-developed neovolcanic zone and by a remarkably high

134   rate of teleseismic and hydrophone-recorded seismicity (Smith et al., 2003). The area has also

135   been identified as a region of active detachment faulting, mostly concentrated on the west side

136   of the axis (Escartin et al, 2008).

137   RV Knorr Cruise 210 Leg 5 conducted bathymetric, dredging and AUV surveys focused on

138   the neovolcanic zone and western rift mountains (see Smith et al., 2014; Parnell-Turner et al.,

139   2018). Based on the occurrence of corrugated surfaces, different core complexes were

140   identified: I) the Southern Core Complex (SCC), a classic, domed, detachment fault with a

141   slope of 13º and corrugation wavelengths between 400 and 1600 m; II); the Central Core

142   Complex (CCC), characterized by two distinctive ridge-parallel, linear ridges (East and West

143   Ridges, Fig. 1), interpreted to be the breakaways of detachment faults that underwent

144   significant rotation (Macleod et al., 2009; Smith et al., 2006; Smith et al., 2008; Parnell-Turner

145   et al. 2018); and III) The Northern Core Complex (NCC). Compared to the South and Central

146   complexes, where the detachment faults intersect the valley floor with a well-developed axial

147   volcanic ridge, the NCC is flanked by a deep axial rift valley, covered by lava flows, but

148   lacking a neovolcanic ridge. This feature suggests that the formation of the NCC was

149   accompanied by limited volcanic activity compared to the Southern Core Complex.

150   During the RV Knorr Cruise 210 Leg 5 dredging program, we sampled both the footwall of

151   the detachment faults and the volcanic rocks exposed on the associated ridge axis (72 dredges

152   in total). Overall, ~5,800 kg of basalt, diabase, gabbro (s.l.), and peridotite samples were

153   dredged in 63 locations (see Table S1 of the supplementary material). However, the proportions

154   of different rock types vary significantly across the area (Fig. 2). Pillow basalt is the most

155   abundant dredged lithology at all the three core complexes. They have been collected from

156   both the crests of the complexes (i.e. detachment breakaway zone) and as the hanging wall

157   debris resting on the detachment fault footwall. Mantle peridotites (harzburgites to dunites) are

158   abundant in the CCC, whereas only gabbroic rocks were collected at the NCC. A very limited

159   amount of gabbroic lithologies (~1 kg in total) were recovered at the southern and central core

160   complexes, limited to small intrusions within mantle peridotites. The main characteristics of

161   these areas are reported in Smith et al. (2014) and Parnell-Turner et al. (2018). Here we focus

162   our analyses on the NCC and its surrounding areas, capitalizing on the abundance of gabbroic

163   lithologies there to infer lower crustal formation and evolution.

164

165   **2.1. Northern core complex and its surroundings**

166 The gabbros we studied were mainly collected at the Northern Core Complex. This complex is

167 delineated by a detachment fault, which is locally masked by several rafted blocks of volcanic

168 crust. Smith et al. (2014) suggested that these blocks were produced by short-lived steep-

169 dipping normal faults formed close to the rift axis that merged at depth with the primary

170 detachment fault (see Reston and Ranero, 2011). Each fault block is covered by volcanic

171 hummocks and flows similar to those on the valley floor. The median valley adjacent to the

172 NCC is narrow (3–4 km) and deep (~4,500 m), and is characterized by smooth, likely

173 sedimented surfaces covering flat-topped lava flows and hummocks. Although there is no

174 neovolcanic axial ridge, the occurrence of volcanic ridges and hummocks emerging from the

175 sediments at the west side of the median valley and the flows on the rift valley floor suggest

176 that this segment was volcanically active in the recent past. Several dredges were deployed at

177 the central portion of the NCC, recovered 579 kg of fresh pillow basalts from the terrace of the

178 rafted block (D52, D56, D57, D65).

179   237 kg of gabbroic rocks were recovered in 10 of 14 dredge hauls (i.e., D49 to D55, D59,

180 D60, D63, D64, D67, D69, D70, D72) deployed along a ridge around the complex (from the

181 southern shoulder to two narrow ridges in the north). Gabbros in this locations are associated

182 with 157 kg of peridotite sampled in 8 dredges and 513 kg of fresh pillow basalts found in 6

183 dredges. Dredged gabbros were described macroscopically during the expedition. Each

184 member of the Shipboard Party was responsible for one or more aspects of the description (e.g.,

185 rock-type, textures, mineral modes and habits) to ensure consistency throughout the dredged

186 hauls. Only primary minerals were considered to define the initial modal proportion;

187 completely altered rocks are referred to as metagabbros (Table S1 of supplementary material).

188 Representative sample images are displayed in Figure 3. Preliminary onshore analyses followed

189 the shipboard observations in order to test the accuracy of the definition of rock types and

190 deformation intensity. Samples reviewed for the present study provided a good correspondence

191 with the shipboard observation, although minor interstitial phases such as orthopyroxene and

192 amphibole had been commonly underestimated (see textural description). This generally did

193 not affect the classification of the rock types, which is based on major mineral phases. The

194 proportions (wt.%) of different gabbroic rocks collected from each dredge are reported in

195 Figure 2, which shows that the lower crust exposed at NCC displays a bimodal distribution,

196 i.e., high proportions of primitive plutonic rocks (i.e., dunites and troctolites >60 wt.%)

197 associated with evolved gabbronorites to oxide gabbros (~30 wt.%). Intermediate rocks such

198 as olivine gabbros (olivine >5 vol.%; clinopyroxene >10 vol.%; oxides <2 vol.%) and gabbros

199 *sensu strictu* (olivine <5 vol.%; oxides <2 vol.%) and gabbronorite (olivine< 5vol.%;

200 orthopyroxene >5 vol.%) are very limited, as they are to the south. Interestingly, troctolites

201 were mainly collected in two dredges (D59, D60; Fig. 2) otherwise characterized by large

202 proportions of serpentinized dunites and pyroxene-bearing peridotites, and by the absence of

203 volcanics. The exposure of mantle peridotites, dunites and troctolites and rarity of basalts are

204 typical of sequences thought to represent crust-mantle transitions, such as the OPD Site 895 at

205 Hess Deep (Dick and Natland, 1996), the Adam Dome at the Kane FZ (Dick et al., 2008), the

206 breakaway area of the Godzilla Megamullion (Sanfilippo et al, 2013; 2016a), and the Uraniwa

207 Hills at the Central Indian Ridge (Sanfilippo et al., 2015; 2016b).

208

209 **3.   MAIN PETROGRAPHIC FEATURES OF MAR 16.5º N GABBROS**

210 **3.1 Igneous textures**

211 Troctolites have euhedral to subhedral olivine and plagioclase, locally showing sub-rounded

212 grain boundaries and a poikilitic texture (Fig. 4a). Clinopyroxene commonly occurs as films or

213 discrete grains interstitial to olivine and plagioclase. Large clinopyroxene oikocrysts (up to 7

214 mm) are locally present within the olivine-plagioclase matrix, commonly including euhedral to

215 irregular plagioclase and olivine chadacrysts (Fig. 4b). Clinopyroxene oikocrysts show film-

216 like apophyses interstitial to olivine and plagioclase, locally associated with orthopyroxene

217 films. In a few cases, oikocrystic clinopyroxene reaches 10 vol.% of the rock, forming

218 spatially defined Ol-gabbro domains. Pargasitic brown amphibole is common within the

219 troctolites as films or interstitial discrete grains locally reaching 0.1 mm (Fig. 4c). One

220   troctolite contains large amphibole oikocrysts up to 3 mm. These oikocrysts include rounded

221   plagioclase chadacrysts, having irregular grain boundaries taken as evidence of partial

222   dissolution (Fig. 4d). Fe-Ti oxides are locally present in troctolites, but are restricted to within

223   the interstitial brown amphibole at the rim of olivine or plagioclase.

224   Gabbros and gabbronorites consist of plagioclase and pyroxenes commonly showing a

225   subophitic texture. Olivine is absent or restricted to small grains (<0.5 mm) rounded in shape.

226   Small (<0.1 mm) irregular grains of Ti-Fe oxides and apatite are locally included in the

227   pyroxenes or at the rim of pyroxene-plagioclase crystals. Orthopyroxene in gabbronorites

228   occurs in large, subequant crystals with subhedral habit, as distinguished from intergranular

229   rims commonly found in olivine gabbros. Pargasitic brown amphibole is rare in these rocks,

230   and present as <0.05 mm-thick films around clinopyroxene.

231   Oxide gabbros (>2% oxide) have been found as i) coarse (up to 20 cm) grained massive

232   samples and as ii) medium/fine-grained centimeter-sized seams crosscutting either gabbros or

233   gabbronorites. In both cases, oxide gabbros are mostly constituted by granular plagioclase,

234   clinopyroxene and locally orthopyroxene, and abundant (up to 15 vol.%) Ti-Fe oxides. Olivine

235   is absent from all of these rocks, whereas accessory phases such as zircon and apatite are

236   commonly present. Brown pargasitic amphibole is ubiquitous within the oxide gabbros,

237   occurring as interstitial grains to coronas surrounding opaque minerals or as blebs within partly

238   exsolved clinopyroxene. Coarse-grained oxide gabbros contain irregular oxide-rich patches up

239   to 10 mm wide separating, and in some cases, surrounding aggregates of disrupted plagioclase

240   and clinopyroxene grains. The oxide-rich domains locally include small plagioclase and/or

241   apatite grains (0.1-0.5 mm), subhedral to rounded in shape. The oxide-gabbro seams are

242   typically found as medium to fine-grained, centimeter-sized layers parallel to the foliation in

243   the host rock. They have a planar aspect and well-defined sutured contacts with the host

244   gabbros, although oxide minerals can be locally dispersed in the host rock a few millimeters

245   from the contact.

246

247 **3.2 Textures indicative of high-temperature deformation**

248 Metamorphic grade and crystal-plastic fabric intensity were documented for each sample,

249 which were cut perpendicular to the foliation where possible. Approximately 40% of the

250 gabbros preserve textures indicative of shearing and recrystallization under hyper- to

251 subsolidus conditions. Deformation intensity was classified on a scale from 0 to 5 (0,

252 undeformed; 1, slightly deformed; 2, sheared; 3, protomylonitic; 4, mylonitic and 5,

253 ultramylonitic) based on the foliation, grain size and relative proportions of neoblasts and

254 porphyroclasts. The distribution of the deformation intensity for the different rock-types (Fig.

255 5) suggests that the higher grade crystal-plastic deformation (grades 3 to 5) is equally recorded

256 by primitive and evolved gabbros, while oxide gabbros and gabbronorites seem to be on

257 average more deformed than troctolites (~50% and ~30% of the total, respectively) at lower

258 grades.

259 *3.2.1-Granulite grade deformation*

260 Protomylonites and mylonites generally appear to be indistinguishable on the basis of mineral

261 proportions from their protoliths, which is not the case for ultramylonites (Fig. 6a; b).

262 Protomylonites are defined by the occurrence of a porphyroclastic texture that locally grades

263 into mylonites having alternating bands of neoblastic plagioclase, olivine and/or

264 clinopyroxene (Fig. 3). The mylonitic zones are up to decimeter wide, although we cannot

265 determinate the original thickness because undeformed portions of the same rock are not

266 recovered. The most common metamorphic assemblage includes recrystallized olivine,

267 pyroxenes and plagioclase. Depending on the protolith, spinel (in troctolites) or Ti-Fe oxides

268 (in gabbros and oxide gabbros) are also found as neoblastic phases. Olivine, pyroxenes and

269 plagioclase porphyroclasts are kinked, folded, and displaced by subgrain boundaries. Equant

270 to subequant neoblasts surround porphyroclasts of the same mineral, suggesting subgrain

271 boundary rotation and recrystallization under high-temperature conditions (Miranda and

272 John, 2010; Hansen et al., 2013). Trace amounts of pargasitic amphibole are locally found

273 within the neoblastic assemblage, as small (<0.1 mm) grains in textural equilibrium with the

274 major phases. Samples showing this metamorphic assemblage will be referred to hereafter as

275   "gabbro granulites", in analogy with the definition of similar rocks from other oceanic core

276   complexes worldwide (see Dick et al., 2000; Blackman et al., 2006; 2011; Ildefonse et al.,

277   2007; Dick et al., 2008; Miranda and John, 2010; Hansen et al., 2013; MacLeod et al., 2017).

278   *3.2.2-Ultramylonites*

279   The mylonites locally grade into 5 to 10-mm thick, very fine-grained (<0.1 mm) ultramylonitic

280   bands. These bands are always found in the median portions of the mylonites, and have sub-

281   planar shape and sharp boundaries towards the host rock (Fig. 3). Unlike the mylonites, which

282   are mineralogically indistinguishable from the magmatic protolith, the ultramylonites have a

283   distinct mineralogical composition. We found ultramylonites developed from different

284   protoliths ranging from troctolites (Fig. 6a) to oxide gabbros (Fig. 6b). These two end-

285   members will be described hereafter.

286   *i)*   Figure 6a shows a ~5-mm thick ultramylonite located within a troctolite mylonite. The host

287   protomylonite and the mylonite are formed by olivine and plagioclase porphyroclasts

288   mantled by a neoblastic assemblage of plagioclase and olivine, locally including

289   clinopyroxene and brown amphibole (Ti-pargasite). Spinel porphyroclasts are also present

290   and partly recrystallized along the main foliation planes. In contrast, olivine and spinel are

291   absent in the ultramylonite band, which is mainly formed by a very fine-grained (<0.1 mm)

292   assemblage of plagioclase and brown amphibole (Ti-pargasite) and rare clinopyroxene.

293   Both clinopyroxene and brown amphibole are locally replaced by green amphibole and

294   chlorite in a coronitic texture.

295   *ii)*   Figure 6b shows an up to ~5-mm thick ultramylonite band located within an oxide-gabbro

296   mylonite. The host mylonite is characterized by plagioclase and clinopyroxene

297   porphyroclasts immersed in a neoblastic assemblage of plagioclase, clinopyroxene and Ti-

298   Fe oxides (~5-10 vol.%). Pale to green amphibole (hornblende) statically replacing the

299   neoblastic clinopyroxene is abundant, whereas brown amphibole is absent. In contrast, the

300   ultramylonite band is characterized by a very fine-grained assemblage of plagioclase and

301   brown amphibole (Ti-pargasite) locally associated with minor clinopyroxene, and by the

302   absence Ti-Fe oxides. Similar to the ultramlylonite located in the troctolite, green

303   amphibole and chlorite locally replace the brown amphibole and the clinopyroxene.

304   Taken as a whole, despite the fact that the two ultramylonites developed from different

305   magmatic protoliths, the ultramylonites are texturally and mineralogically indistinguishable,

306   consisting of a fine-grained neoblastic assemblage of plagioclase, brown amphibole and, more

307   rarely, clinopyroxene.

308

309   **3.3 Textures indicative of high-temperature alteration**

310   Deformed gabbros locally show traces of alteration at high-temperature conditions

311   (amphibolite facies). This process is mainly indicated by *i)* the occurrence of amphibole-rich

312   veins crosscutting the foliation in the host gabbros at high angle; *ii)* the occurrence of pale-

313   brown amphibole (hornblende) statically replacing the clinopyroxene. These textures are

314   well exemplified in Figure ~~3e~~3f, which shows a sheared gabbronorite crosscut by an

315   undeformed amphibole-rich vein, at high angle with the main foliation in the host rock.

316   Hornblende in the veins shows a fibrous texture and is elongated parallel to the vein wall.

317   Plagioclase close to the contact with the vein is statically recrystallized in a fine-grained

318   assemblage. Clinopyroxene in the host gabbro is almost totally replaced by single grains or

319   aggregates of hornblende spatially locally associated with very fine-grained (<0.05 mm)

320   albitic plagioclase (Fig. 3f). Similar textures are also found where irregular veins crosscut the

321   gabbros, leading to the formation of hornblende at the expense of clinopyroxene, and fine-

322   grained plagioclase replacing the coarse plagioclase of magmatic origin. In almost all of the

323   samples, hornblende is locally rimmed or partially replaced by green amphibole and chlorite,

324   as a response to lower-temperature greenschist facies alteration.

325

326   **4.   MAJOR ELEMENT MINERAL COMPOSITIONS**

327   Major element mineral compositions of selected 16.5ºN gabbros were obtained using a JEOL

328   JXA-8200 electron microprobe (Dipartimento di Scienze della Terra, Universita`degli Studi di

329   Milano, Italy) at 15 kV accelerating voltage and 15 nA beam current. Counting time was 30 s

330   on the peak and 10 s on the backgrounds. Natural silicates were utilized as standards and data

331   reduction was carried out using the CITZAF package (Armstrong 1995). Undeformed and

332   variably deformed samples ranging from troctolites to oxide gabbros were selected for the

333   microprobe study, in order to cover the main chemical variability of the rock-types.

334   Representative analyses of olivine, plagioclase, pyroxene and amphibole of selected samples

335   are reported in supplementary material in Tables S2 to S6. Four samples were also mapped

336   using a Zeiss Sigma HD field emission gun SEM installed at Cardiff University. This machine

337   is equipped with dual 150-mm^2 Oxford Instruments X-Max^N energy dispersive silicon drift

338   detectors, which enable high count rates (>1,000,000 cps) to be obtained. The data were

339   background corrected using Oxford Instruments AZtec software prior to the production of

340   element maps.

341   Olivine in troctolites has high forsterite (Fo) contents (86 to 88 mol.%), and relatively

342   constant NiO (0.20–0.25 wt.%) and MnO (0.19–0.25 wt.%) contents. There is no

343   compositional difference between porphyroclastic and neoblastic olivine. Plagioclase anorthite

344   content varies from An_{81} in troctolites to An_{46} in oxide gabbros. Chemical analyses and

345   element maps clearly show that high-temperature deformation is associated with a decrease in

346   anorthite in plagioclase neoblasts compared to porphyroclasts in the same rock. This is well

347   illustrated in Figure 7a, where compositional maps for Na contents of a sheared troctolite are

348   shown. In particular, the lower Na (and higher Ca, not shown) of the fine-grained neoblasts

349   clearly indicates lower An contents (~78 mol.%) compared to that of the porphyroclasts (~81

350   mol.%). Similarly, an even greater variation in plagioclase anorthite component characterizes

351   the troctolite mylonite–ultramylonite association (Fig. 7b). Here, the plagioclase in the

352   ultramylonite has lower Ca (and higher Na) compared to that in the mylonite, resulting in a

353   decrease from An_{81} in the mylonites to An_{65} in the ultramylonite. A unique exception is

354   represented by the mylonite–ultramylonite association in the oxide gabbronorite (Fig. 7c),

355   where compositional maps show higher Ca and lower Na contents of the plagioclase in the

356   ultramylonite (An_{82}) compared to that in the host mylonite (46 mol.%). Finally, a decrease in

357   anorthite component is apparent in recrystallized plagioclase associated with the formation of

358   the amphibole-rich vein in Figure 7d. Here, the neoblastic plagioclase at the Amphibole-

359   vein/gabbro contact and the plagioclase in the vein has higher Na and lower Ca compared to the

360   plagioclase porphyroclasts, resulting in An_{53} for the former and ~An_{58} for the latter. Note that

361   very high Na and low Ca characterize the fine-grained plagioclase associated with veins and

362   microcracks crosscutting the magmatic phases (An_{20} to An_{25}).

363   Clinopyroxene shows a wide compositional range with Mg# ($100*Mg/\{Mg+Fe\}$)

364   varying from 89 in oikocrystic clinopyroxene in troctolites to 65 in oxide gabbros.

365   Clinopyroxene in troctolites is distinct from that in gabbros by generally higher Cr_2O_3 and

366   Al_2O_3 contents (0.4–1.5 wt.% compared to <0.2 wt.% in the gabbros), but high TiO_2 contents

367   (Fig. 8). Chemical mapping suggests that clinopyroxene neoblasts in sheared to mylonitic

368   rocks are nearly indistinguishable from the porphyroclasts in the same sample. The only

369   exception is the clinopyroxene in oxide gabbro mylonites/ultramylonite shown in Figure 7c,

370   which has a sharp change in chemical composition from the mylonite (Mg#=64–67;

371   Cr_2O_3=0.01–0.02 wt.%) to the ultramylonite (Mg#=73; Cr_2O_3=0.5 wt.%). With the exception

372   of this ultramylonite, there is a positive correlation between Mg# in clinopyroxene and

373   anorthite content in coexisting plagioclase from each sample (Fig. 9a).

374   Orthopyroxene Mg# is well correlated with Mg# of coexisting clinopyroxene (Fig. 9b).

375   Similar to clinopyroxene, orthopyroxene from our sample suite is characterized by (*i*) high Mg#

376   and Cr_2O_3 contents (87–89; 0.4–0.6 wt.%, respectively) in troctolites and progressively lower

377   (75–58; 0.01–0.04 wt.%) in gabbro/gabbronorite/oxide gabbro, and (*ii)* lack of correlation

378   between TiO_2 and Mg#. The composition of the coronitic to interstitial orthopyroxene in the

379   troctolites suggest a magmatic rather than a metamorphic origin, similar to the granular

380   orthopyroxene in the gabbronorites.

381   Amphibole is a ubiquitous phase in our samples. There is a clear chemical distinction

382   between the green to pale-brown amphibole statically replacing clinopyroxene (Type I) and the

383   brown amphibole forming poikilitic to interstitial grains in undeformed samples or found as

384   neoblasts within the granulites (Type II; Fig. 10). Type I amphibole occurs in two forms: i)

385   green amphibole with very low Al^{IV} (<0.39 atoms per formula unit, apfu), TiO_2 (<0.41 wt.%)

386   and alkalis [(K+Na)<0.12 atoms per formula unit, apfu], mainly consisting in actinolite and

387   tremolite (Fig. 10); ii) pale-brown amphibole, having higher Al^{IV} (0.83–1.29 apfu), TiO_2 (0.3–

388   1.4 wt.%) and alkalis (0.49–0.78 apfu) and mostly represented by hornblende (Fig. 10b). Pale-

389   brown hornblende also constitutes the amphibole-rich veins. These amphiboles have Al^{IV}

390   (1.24–1.41 apfu) and alkalis (0.51–0.82 apfu) partly overlapping those in the amphibole

391   coronas, but have higher TiO_2 contents (3.0–3.3 wt.%). TiO_2 contents of the hornblende in the

392   vein decrease towards the rim. Chlorine in hornblende coronas and hornblende in veins varies

393   between 0.1–0.2 wt.% (Fig. 10b, c).

394   In contrast, Type II amphibole is brown to reddish brown, and mostly consists of Ti-

395   pargasite. Within the undeformed troctolite and gabbro samples, this amphibole has high Al^{IV}

396   (1.74–1.82 apfu) and alkalis (0.61–0.79 apfu) and very high TiO_2 contents (4.2–4.9 wt.%) and

397   high Mg# (Fig 10d). Chlorine is lower than the detection limit of the microprobe analyses (~

398   0.1 wt.%) (Fig. 10b, c). Similar to the hornblende in the amphibole veins crosscutting the

399   gabbros, the large amphibole oikocryst in troctolites show a sharp decrease in Al^{IV} and TiO_2

400   from the core towards the rim. Amphibole neoblasts in mylonites and ultramylonites are

401   chemically similar to the magmatic amphibole in undeformed samples. They have high Al^{IV}

402   (1.50–1.82 apfu) and alkalis (0.62–1.80 apfu), low chlorine (<0.1 wt.%), and highly variable

403   TiO_2 contents (1.5 to 4 wt.%). The neoblastic amphiboles in the two ultramylonites are overall

404   similar to both the magmatic amphibole and the neoblastic amphibole in the mylonite. The

405   amphibole in the troctolitic ultramylonite tends to have higher Al^{IV} and TiO_2 contents and Mg#

406   than the amphibole in the oxide gabbro ultramylonite (Fig. 10b; d).

407   **5. DISCUSSION**

408   **5.1-Bimodal distribution of gabbros from the 16.5º N region of Mid-Atlantic Ridge: a**

409   **quantitative model to explain the compositional variability of the lower oceanic crust**

410   The composition of the gabbroic rocks collected from the 16.5º N region of the MAR

411   indicates a lower oceanic crust most likely formed by scattered intrusions of chemically

412   primitive gabbros (i.e. troctolites), mainly associated with evolved gabbronorites and oxide-rich

413   gabbros. Intermediate rocks such as olivine-gabbros and gabbro *sensu stricto* are restricted to

414   <3 wt.% of the ~356 kg of gabbros sampled in the region (Fig. 2). This bimodal distribution is

415   remarkably different to that of the gabbroic section drilled at Hole U1309D (Blackman et al.,

416   2006; 2011), formed in a relatively robust magmatic center along the Mid-Atlantic Ridge

417   (Godard et al., 2009). It also contrasts sharply with the Atlantis Bank gabbro massif on the SW

418   Indian Ridge, which is dominated by moderately fractionated olivine gabbro that grades into

419   subordinate highly evolved oxide gabbros (e.g.: Dick et al., 2000; MacLeod et al., 2017).

420   Motivated by the similarity of the gabbros exposed in the 16.5ºN region with those of the Kane

421   Megamullion and in the vicinity of the 15°20'Fracture Zone, hereafter we examine how magma

422   supply and melt extraction may control the overall composition of the crustal sections exposed

423   through detachment faulting. We calculate the relative amount of different gabbroic lithologies

424   (i.e. dunites, troctolites, olivine gabbros, gabbros and oxide gabbros) crystallized during a

425   fractional crystallization process using the thermodynamic program *MELTS* (Ghiorso and

426   Sacks, 1998; Ghiorso et al., 2002). In order to take into account that part of the initial liquid

427   mass is erupted as basalt on the seafloor, we performed this calculation in different steps,

428   allowing a proportion of the liquid to be subtracted at each step of the crystallization process

429   (parameters and results of the model in Tables S7, S8, S9 of the supplementary files). The

430   starting liquid (*Liq 0*) is a primary melt formed by ~10 % melting of a DMM peridotite source

431   (Mg# ~72; Kinzler and Grove, 1993). After a defined amount of differentiation (for the sake of

432   simplicity fixed at ΔMg# ~4), a precise mass of liquid *(Liq 1)* is subtracted and the remaining

433   composition of *Liq 1* is used as initial composition for the next step of differentiation. The

434   mass of liquid subtracted is calculated by scaling the melt mass in proportion to the

435   compositional distribution of the erupted basalts, which is thereby a fundamental input of our

436   model. We cannot use the observed compositional variability of basalts sampled exclusively in

437   the 16.5ºN region, considering that the genetic relationships between the basalts along axis and

438   the gabbros exposed through detachment faulting is not obvious and that basalt sampling is

439   inevitably fragmentary. It has been previously noted that the compositional distribution of

440   basalts at the MOR is globally dependent on the spreading rates (Melson et al., 1976; Sinton

441   and Detrick, 1992). In particular, basalts erupted at slower-spreading ridges tend to be more

442   primitive compared to those erupted at fast-spreading ridges (Fig. 11). The large-scale

443 compositional variability of the basalts can thereby be considered meaningful, and allows us to

444 compare the results of our model with different lower crustal sections exposed along the same

445 ridge segment. Here, we used the overall compositional distribution of the basalts erupted along

446 the 10–35ºN segment of the MAR, which comprises ~2,700 glass analyses selected from the

447 PetDB database (reference Lehnert paper), making the necessary assumption that the

448 distribution is constant for all the modeled scenarios. An explanation of the model is reported in

449 Figure 12.

450   The solid composition is calculated at each step from the total mass of crystallizing

451 phases given by the program (Fig. 13a; b). The crystallizing phases are restricted to olivine,

452 plagioclase, clinopyroxene and Fe-Ti oxide, following the fractionation trends inferred for a

453 MORB melt (see Grove et al., 1993). Minor phases such as orthopyroxene and amphibole are

454 not considered in the model because they occur as interstitial phases, likely related to

455 crystallization from intercumulus melt. This is in contrast to the primitive orthopyroxene-

456 bearing cumulates that have been sampled in different locations along the mid-ocean ridge

457 system. The orthopyroxene from these rocks has primitive compositions and may represent

458 crystallization from melts more depleted that typical MORB (e.g., Ross and Elthon 1995) either

459 derived from melting an hydrated mantle (Nonnotte et al. 2005) or from melt residual by

460 interaction with mantle peridotites (see also discussion in Gillis et al. 2014). However, it should

461 be noted that gabbronorites having granular orthopyroxene as a cumulus phase were rarely

462 sampled in this study, and that these rocks do not contain olivine and have relatively evolved

463 mineral compositions, plotting in the field of the orthopyroxene-free gabbros (Figs 8 and 9).

464 For the sake of simplicity, we hence grouped the gabbronorites with the gabbros *sensu stricto*

465 in our model. The total mass of initial liquid subtracted by eruption (M_e) with respect to that

466 crystallized (M_c) is referred to as M_e/M_c. At M_e/M_c =0 the melt is totally retained in the lower

467 crust and the mass of gabbro crystallized corresponds to that of pure fractional crystallization.

468 At high Me/Mc the melt is preferentially erupted and the relative mass of crystallized gabbros

469 decreases.

470   The proportion (mass %) of gabbros crystallized at different M_e/M_c (0 to ~7:3) is

471   shown in Figure 13, where the chemical and modal compositions of the crystallizing

472   assemblages and the distribution of the resulting erupted basalts are shown at steps of ΔMg# 4.

473   Remarkably, the olivine and clinopyroxene crystallized from the computed melt fits the

474   composition of the natural samples from this study very well. As was expected, this model

475   shows that at increasing M_e/M_c, a large proportion of the melt is erupted on the seafloor during

476   the first phase of crystallization. Hence, the crystallizing assemblage tends to be progressively

477   more primitive, i.e.: olivine ± plagioclase assemblages are preferentially formed until the melt

478   is exhausted (Fig. 13e). At the same time, at high M_e/M_c ratios most of erupted melts are

479   restricted to basalts with very primitive compositions.

480   This simple scenario is highly improbable, however, as it implies that the liquid mass

481   is entirely extracted from the cumulate at each step of differentiation. Studies of oceanic

482   gabbros indicate that variable amounts of interstitial melt are retained within the crystal mush

483   (e.g.: Meyer et al., 1989; Natland and Dick, 2001). Hence, we assume that a proportion of

484   the initial melt is retained at each step of the crystallization process, following a separate

485   liquid line of descent (see Fig. 12). We must consider that the interstitial melts can crystallize

486   within the cumulus matrix at the closure of the magmatic system (e.g.: Tribuzio et al., 2000;

487   Borghini and Rampone, 2007) or can migrate through the crystal mush driven by buoyancy

488   differences or by compaction (e.g.: Dick et al., 2000; Natland and Dick, 2001). In both cases,

489   textural and geochemical evidence in gabbros suggest that these melts tend to partly

490   assimilate the crystal matrix, causing geochemical modification ascribable to assimilation–

491   fractional-crystallization (AFC) processes (e.g.: Coogan et al., 2000; Borghini and Rampone,

492   2007; Gao et al., 2007; Lissenberg and Dick, 2008; Drouin et al., 2009; Sanfilippo et al.,

493   2013; Lissenberg and MacLeod, 2016). The chemical response to these processes has been

494   modelled following equations from DePaolo (1981). These equations replicate the variations

495   in trace element compositions of the reacting melts and of the phases crystallized from it (see

496   Coogan et al., 2000; Gao et al., 2007; Borghini and Rampone, 2007; Lissenberg et al., 2013;

497   Sanfilippo et al., 2015; Lissenberg and MacLeod, 2016). They have been successfully used to

498   test AFC hypotheses in well-constrained melt-rock reaction scenarios, such as the conversion

499 of dunite to troctolite (e.g., Rampone et al., 2016 and reference therein) or troctolite to

500 olivine-gabbro (Lissenberg and Dick, 2008). When applied to the scale of the entire oceanic

501 crust, a process of assimilation would introduce a number of independent variables (i.e., the

502 modal amount of mineral phases involved; the chemical compositions of the reactants; the

503 mass assimilate/mass crystallized ratios; see discussion in Lissenberg et al., 2013). Such a

504 model could not be reproduced by the *MELT* model used here.

505   As a simplification, at each step we subtracted a certain amount of melt proportional

506 to the melt density (Fig. 13d), considering that a maximum of 30% of liquid can initially be

507 trapped in the solid framework before migrating upwards due to compaction by density

508 driven flow (Morse, 1979; Meyer et al., 1989). Then we calculate the gabbros crystallized

509 from the total mass of trapped melts (Fig. 12). The mass of gabbro resulting from this

510 calculation was then added to that of the gabbros previously obtained. Note that owing to an

511 increase in melt density with the chemical evolution, our model supposes that the amount of

512 melt potentially trapped in the crystal framework increases at decreasing liquid Mg#. This

513 causes a dependence of the amount of trapped melt on the M_e/M_c ratio, with a more extensive

514 melt entrapment occurring at low M_e/M_c ratios (Fig. 13e).

515   As expected, this calculation shows that the relative amount of evolved lithologies

516 (mainly oxide-gabbro) increases in all the modelled scenarios (Fig. 13e). In particular, at high

517 M_e/M_c ratios, most of the initial liquid is erupted on the seafloor before the melts reach

518 saturation in clinopyroxene, thereby preventing the formation of moderately evolved gabbros

519 (olivine-gabbros and gabbros *sensu stricto*). However, at the same time, the occurrence of

520 trapped liquid enables the local formation of evolved lithologies as this melt migrates due to

521 upward compaction and lateral density driven flow (gabbronorites to oxide gabbros). As a

522 consequence, at high *Me/Mc* ratios the model is able to produce a bimodal distribution of the

523 gabbros, characterized by large proportions of primitive lithologies and lack of intermediate

524 compositions.

525   It is noteworthy, however, that the model fails to reproduce the large amount of oxide

526 gabbros present in the 16.5 ºN region. There are several factors that can account for this. Most

527   oxide gabbros appear to be formed by interstitial melts expelled from the cumulate matrix and

528   by upward compaction and/or high temperature deformation into narrow zones of the crystal

529   mush often associated with shear zones along the detachment fault surface (Dick et al., 2000;

530   Natland and Dick, 2001; Macleod et al., 2017). Moreover, it has been recently observed that in

531   addition to an upward increase in oxide gabbros at the Atlantis Bank gabbro massif, that the

532   body is laterally zoned to more evolved cumulates that comprise the entire section from the

533   base of the crust to the dike-gabbro transition at the distal end of the body near the bounding

534   Atlantis II Transform (MacLeod et al., 2017). This shows that density driven lateral flow by

535   either direct intrusion or permeable flow plays a major role in the architecture of the lower

536   oceanic crust. Thus, as seen at the Kane Megamullion (Dick et al., 2008) and Atlantis Bank

537   (MacLeod et al., 2017), while dunite and troctolitic dikes are rare in mantle massifs adjacent to

538   major magmatic centers at slow spreading ridges, oxide gabbro dikes are widely distributed

539   lithology. Thus, the amount of oxide gabbros can be biased by dredging and drilling in areas of

540   low magma flux, with highly centralized magmatic centers and intervening relatively

541   amagmatic regions, as from 13° to 17°N on the MAR (e.g.: Mendel et al., 1997, Sauter et al.,

542   2001, Standish et al., 2008). Finally, we emphasize that any AFC process would tend to

543   increase the incompatible element composition of the interstitial melt (see Lissenberg and

544   MacLeod, 2016 and reference therein), producing melts anomalously enriched in incompatible

545   elements compared to a melt evolving through fractional crystallization. This suggests that a

546   melt produced by AFC would potentially enriched the melt in Ti, bringing the saturation in Fe-

547   Ti oxides forward, and producing higher amounts of oxide-bearing lithologies. This is another

548   possible reason why our model seems to underestimate the amount of oxide-gabbros compared

549   to the natural occurrences.

550   Although fairly simplistic, in combination with insight provided by field mapping into

551   ocean crust architecture, this model provides insight into why a bimodal distribution of the

552   lower oceanic crust may be expected in regions where a high proportion of liquid produced in

553   the mantle is erupted on the seafloor, with a minor proportion of intrusive rocks. This scenario

554   recalls the "plum-pudding model" suggested for slow-spreading ridges (Cannat, 1993, Cannat

555   et al., 1995; 1997), which proposes that the lower oceanic crust might consist of small gabbroic

556   plutons scattered within the lithospheric mantle (see also Kelemen et al., 2004; Dick et al.

557   2008; 2010).

558   The outcomes of our model are consistent with the compositional heterogeneity of the

559   lower oceanic crust along the MAR (Fig. 13). For instance, a lower oceanic crust mainly

560   consisting of primitive gabbros and highly evolved lithologies is expected at regions

561   characterized by scattered injections intruded within mantle peridotites, such as the region

562   considered here or the 15º20'N fracture zone drilled during Leg 209 (Kelemen et al., 2004).

563   On the other hand, a lower oceanic crust mostly composed of moderately evolved gabbros

564   associated with primitive lithologies (troctolites and dunites) and minor oxide gabbro is

565   expected in areas of a thick crust and moderate melt supply. This can be the case for relatively

566   robust magmatic segments as for instance the Atlantis Massif OCC (Blackman et al., 2007;

567   2011; Ildefonse et al., 2007; Godard et al., 2009; Drouin et al., 2009; 2010) or the crustal

568   transect exposed at the Adam Dome of the Kane megamullion (Dick et al., 2008) (Fig. 12d).

569

570   **5.2 Emplacement and crystal-plastic deformation of the NCC gabbros**

571   Clinopyroxene in the 16°30'N gabbros are characterized by different combinations of

572   Mg# and TiO_2 contents (Fig. 8). In particular, they show two distinct chemical trends in

573   troctolites (Trend 1) and gabbros/gabbronorites/oxide gabbros (Trend 2). The

574   clinopyroxenes following Trend 1 show a sharp increase in TiO_2 at nearly constant Mg#;

575   whereas those following Trend 2 display a gradual increase in TiO_2 at decreasing Mg#

576   followed by (from Mg# ~70) a renewed decrease in TiO_2. These chemical trends are typical

577   for clinopyroxene from abyssal gabbros (see Coogan, 2007) and respectively explained as

578   (Trend 1) evidence for magmatic metasomatism (i.e., melt–rock reactions) in a troctolitic

579   matrix (see also Lissenberg and Dick, 2008; Suhr et al., 2008; Drouin et al., 2009; Sanfilippo

580   and Tribuzio, 2013; Sanfilippo et al., 2013; 2015a; b; Lissenberg and MacLeod, 2016);

581   (Trend 2) evidence for progressive magmatic differentiation followed by saturation in Fe-Ti

582   oxides to form the oxide gabbros. The gradual decrease in clinopyroxene Mg# versus

583   plagioclase An contents, and the variability of the Cpx suggest that the

584   troctolite/gabbro/oxide-gabbro series crystallized from cogenetic melts evolving through

585   fractionation of Ol, Pl, Cpx and, finally, Ti-Fe oxides. Despite this, it seems obvious that the

586   crystallization of Cpx in the troctolites also involved interactions between the olivine and

587   plagioclase matrix and an evolved interstitial melts. This is consistent with the occurrence of

588   large brown amphibole occurring as interstitial to oikocrystic grains within the troctolites

589   (Fig. 4c;d). This amphibole locally shows poikilitic textures with inclusions of irregular

590   plagioclase chadacryst and small Ti-Fe oxides, and is characterized by high TiO_2, (K+Na)

591   and Al, and negligible Cl contents (Fig. 10). The poikilitic texture and the chemical

592   compositions are reconcilable with a magmatic origin, indicating the crystallization of late-

593   stage melts enriched in water and incompatible elements (see also Tribuzio et al., 2000;

594   Coogan et al., 2001; Borghini and Rampone, 2007). Hence, in agreement with the extensive

595   literature data for abyssal gabbros (e.g., Lissenberg et al., 2008; Suhr et al., 2008; Drouin et

596   al., 2009; Renna and Tribuzio, 2010; Sanfilippo et al., 2013; 2015; Lissenberg and MacLeod,

597   2016; Rampone et al., 2016), the enrichments in TiO_2 in the high-Mg# clinopyroxene and the

598   occurrence of magmatic amphibole with poikilitic texture testify that the 16.5ºN troctolites

599   underwent a process of magmatic metasomatism by an interstitial, evolved melt. Below, we

600   show that evolved melts were also locally present at the onset of the deformation related to

601   the detachment faulting.

602   A large proportion of the 16.5ºN gabbros show textures associated with a process of

603   deformation under high-temperature conditions (Fig. 3). In particular, most samples have

604   olivine, plagioclase, and clinopyroxene recrystallized in polygonal aggregates that are often

605   associated with brown amphibole and Ti-Fe oxides. The amphibole neoblasts in all deformed

606   troctolites (sheared, mylonites, and ultramylonites) are compositionally identical to the

607   magmatic amphibole in the undeformed samples. In addition, the occurrence of deformed

608   amphibole porphyroclasts in a sheared troctolite and the textural equilibrium with the other

609   neoblastic phases (olivine, plagioclase and clinopyroxene) clearly indicate that the formation

610   of pargasitic amphibole in this sample predated the "anhydrous" granulite-facies deformation

611   event.

612   The edenite-richterite amphibole-plagioclase thermometer of Holland and Blundy

613   (1994) on adjacent grains of amphibole and plagioclase was selected to infer the temperature

614   of recrystallization (Fig. 14). In one mylonitic troctolite, the amphibole-plagioclase

615   thermometry was also confirmed by two-pyroxene thermometry (Lindsley and Frost, 1992),

616   which yielded similar temperature estimates (see Fig. 14). Independent of rock-type, the

617   amphibole-plagioclase thermometer provided high temperature estimates for the granulite-

618   grade deformation, varying between 930 ºC and ~1010 ºC. These estimates fall in the range

619   of temperature estimated by Mehl and Hirth (2008), who showed that granulite-facies

620   deformation in gabbros from Hole 735B occurred at temperatures as high as 946 ±70 ºC.

621   These data suggest that the crystal-plastic deformation, which caused dynamic

622   recrystallization and formation of neoblasts of olivine + clinopyroxene ± brown amphibole in

623   the troctolites took place at hyper-solidus conditions (close to the solidus temperature of the

624   magmatic system). The high equilibration temperatures (1004 ± 20 ºC; Fig. 14) estimated on

625   the basis of the coexistence of clinopyroxene–orthopyroxene in one troctolite mylonite

626   confirm this hypothesis. We must note that these temperatures are very close to the stability

627   limit of pargasitic amphibole at low pressures, which is considered to be at temperatures

628   <1000 ºC (Koepke et al., 2004; 2005; 2007). Hence, one possibility is that the temperature

629   estimates for the mylonites give the crystallization temperature of an amphibole formed by a

630   melt present during the deformation process. The chemical modifications in the neoblastic

631   plagioclase of the sheared troctolite in Figure 7a, which has lower anorthite compared to that

632   of the porphyroclasts (An_{78} and An_{81} respectively) may also suggests that the granulite facies

633   deformation in the mylonites might have occurred in the presence of interstitial melt.

634   Although this idea would need to be better constrained by a more thorough chemical

635   characterization of the porphyroclastic and neoblastic minerals in the mylonites, the

636   occurrence of melts at the onset granulite-grade deformation is consistent with the

637   composition of the two ultramylonites shown in Figure 6. Both cases show a clear variation

638   in the neoblastic assemblage crystallized during the strain localization from the proto-

639   mylonite to the ultramylonite zone. In particular, although they are hosted in mineralogically

640   different mylonites, both ultramylonites are formed by anorthitic plagioclase (An_{65} to An_{82}),

641   Ti-rich pargasitic amphibole (TiO_2 1.7–3.7 wt.%), and minor clinopyroxene (Mg# 75 to 85).

642   In contrast to the host rocks (oxide gabbro and troctolite), they do not contain olivine, spinel

643   or Ti-Fe oxides (Figs 6; 7). Felsic rocks containing plagioclase and amphibole are often

644   found in abyssal and ophiolitic gabbros as crosscutting veins or discrete bodies associated

645   with oxide-rich gabbros (Dick et al., 2000; Koepke et al., 2004; 2007; Tribuzio et al., 2014;

646   MacLeod et al., 2017). These rocks are interpreted as the product of crystallization of silica-

647   rich melts produced by extreme fractional crystallization of MOR-type basalt and/or partial

648   melting of hydrothermally altered gabbro (see discussion in Tribuzio et al., 2014). Hence,

649   one can argue that the ultramylonites represent the recrystallization of felsic material

650   originally located within the host rock and mylonitized during the deformation event.

651   However, this is at odds with the high anorthite in plagioclase (An_{83}-An_{65}) and Mg# in

652   clinopyroxene and amphibole (Mg# 75–85 and 67–80 respectively) in the ultramylonites. In

653   addition, the composition of the amphibole in the ultramylonite formed from a troctolite

654   progenitor is similar to that of the amphibole in the host rock (Figure 10b). The temperature

655   estimates for the amphibole-plagioclase couples in the ultramylonite furnish very high values

656   ranging from 930 ºC and ~1020 ºC, comparable to those in the host mylonites (Fig. 14).

657   Based on these data, it is plausible that the ultramylonite layers may represent melt-rich

658   zones developed during the same deformation event as the host granulite-facies mylonites.

659   The peculiar mineralogical and chemical compositions of these layers indicate that the melt

660   had a hybrid composition, i.e., high Mg/Fe and Ca/Na ratios, but rich in silica, incompatible

661   elements and water. The ultramylonite hosted in an oxide-gabbro, has plagioclase showing a

662   strong increase in anorthite component (~82 mol.%) compared to that in the host mylonite

663   (~46 mol.%). The high anorthite is also ascribable to the occurrence of a water-bearing melt

664   during the crystallization of the neoblastic plagioclase (Berndt et al., 2005). Further

665   investigation will shed light on the nature of this hybrid melt and the mechanism that allowed

666   the strain localization to form the ultramylonite.

667   In summary, our data suggest that the granulite facies deformation event recorded by

668   the gabbros occurred shortly after the crystallization of the magmatic phases. This event my

669   have possibly enhanced the expulsion of the liquid from the crystal framework, allowing the

670   formation of oxide-rich seams within the gabbros and/or discrete oxide-rich horizons often

671   associated with high-temperature deformation. A similar process has been proposed by

672   Natland and Dick (2001) to explain the close association of oxide-rich layers and granulite-

673   grade mylonites in gabbros from Hole 735B at Atlantis Bank. We note though that

674   thermometry estimates of the mylonites analyzed by Miranda and John (2010) from Hole

675   735B yielded temperatures <910ºC generally lower than those reported in the present study.

676   Nonetheless, the occurrence of high-temperature ultramylonites likely formed in the presence

677   of melt suggests a complex interplay between deformation and melt percolation that has also

678   been proposed for mylonitic rocks drilled at the inside-corner high of the MAR at 23°N

679   (Agar and Lloyd, 1997).

680

681   **5.3 Hydrothermal alteration**

682   Distinctive black hornblende-rich veins locally crosscut the gabbros at high angle to the high-

683   temperature metamorphic foliation of the host rock. These vein-filled fractures represent

684   formation during brittle deformation that is also related to the crystallization of hornblende

685   pseudomorphs after the magmatic clinopyroxene in the host rock, indicating the host gabbros

686   enjoyed migration of hydrous fluids from the vein into the gabbro. This process also accounts

687   for the formation of secondary plagioclase (albitic in composition) crystallizing within

688   microcracks connecting the different amphibole pseudomorphs.

689   Both the vein and clinopyroxene pseudomorph hornblende in the gabbros have

690   appreciable Cl contents (0.1–0.2 wt.%) similar to that in hydrothermal amphiboles in other

691   abyssal gabbros (Coogan et al., 2001; Cortesogno et al., 2004; Tribuzio et al., 2016). This

692   suggests that both amphibole types formed by interaction with water-rich fluids likely derived

693   from seawater (see also Alt and Bach 2006). However, the amphibole from the veins has lower

694   Mg# (70–77) and higher TiO_2 contents (3.0–3.3 wt.%) compared to the coronitic hornblende

695   adjacent to the veins in the gabbro (Mg# = 75–81; TiO_2 <1.5 wt.%), approaching the Ti and Al

696   of the magmatic pargasite in the troctolites (Fig. 10). The chemical map in Figure 7d shows

697   that the hornblende statically replacing the pyroxene is associated with fine-grained plagioclase

698   having very low anorthite contents (20–25 mol.%) and is typically found in microcracks.

699   Temperature estimates based on plagioclase-amphibole pairs yield amphibolite facies

700   temperatures of ~700 ºC (Fig. 14). This suggests that the hornblende replacing the

701   clinopyroxene likely formed by interaction of the magmatic phases with seawater-derived

702   fluids (Coogan et al., 2001), which caused the recrystallization of Na-rich plagioclase within

703   cracks and microveins (see also Tribuzio et al., 2016). On the other hand, the high Ti and Al in

704   the hornblende from the veins suggest that magmatic components were likely still present

705   within the fluids forming these veins. This is in agreement with the high temperature (920–970

706   ºC) indicated by the plagioclase-amphibole thermometers on a euhedral plagioclase contained

707   in the vein. The occurrence of magmatic components in the hornblende-rich veins was also

708   inferred for the amphibole-rich veins crosscutting gabbro at Atlantis Bank (Alt and Back, 2006)

709   and in the Ligurian ophiolites (Tribuzio et al., 2014) and related to fluids exsolved from highly

710   evolved melts residual from the magmatic differentiation process.

711   In summary, the microtextures and the chemical variability of the amphibole forming

712   the hydrothermal hornblende offer a snapshot to the retrograde evolution of this portion of the

713   lower crust exhumed by a detachment fault. In line with previous studies on abyssal and

714   ophiolitic gabbros, we propose that a magmatic component was still present during the first

715   stages of hydrothermal alteration at the ductile-brittle transition. This stage was followed by

716   further cooling, by the ingress of seawater along cracks and by the static replacement of the

717   magmatic phases by green hornblende and albitic plagioclase.

718

719   **6.   CONCLUSIONS**

720   The lower oceanic crust exposed in the 16.5 ºN region of the Mid-Atlantic Ridge formed in a

721   detachment fault system characterized by a moderate to low magma supply. Dredge hauls

722   indicate a bimodal composition of the lower crust that is formed by a high proportion of

723   dunites and troctolites associated with evolved gabbronorites and oxide gabbros, and a lack

724   of intermediate rock types. Using a thermodynamic model, we infer that a similar bimodal

725   distribution of the lower oceanic crust is expected in regions where gabbroic intrusions are

726   scattered within partly serpentinized mantle peridotite. The formation of these "ephemeral"

727   magma chambers within the mantle lithosphere causes a high eruption rate, which in turn

728   allows most of the melt to be erupted after moderate degrees of fractionation. This idea is

729   also consistent with the evidence that small amounts of melt were present at the onset of the

730   high-temperature deformation related to the oceanic detachment. The development of

731   porphyroclastic to mylonitic fabrics occurred at near-solidus conditions (~1000 ºC) and was

732   likely associated with the expulsion of interstitial, evolved melts from the crystal matrix.

733   This mechanism enhanced the formation of disconnected oxide-gabbro seams or layers often

734   associated with crystal-plastic fabrics in the host gabbros. The high-temperature deformation

735   was followed by hydrothermal alteration marked by the formation of amphibole-rich veins at

736   high-temperature conditions (~900 ºC). This study suggests that the compositional variability

737   of the lower oceanic crust formed at the 16.5 ºN region of the Mid-Atlantic Ridge is the

738   consequence of the complex interplay between low melt supply and high-temperature

739   deformation representing the deep expression of the detachment fault.

740

758

759 **FIGURE CAPTIONS**
760
761 **Figure 1.** Bathymetric map of the Mid-Atlantic Ridge, 16.5 ºN area studied by cruise, R/V
762 Knorr, KNR210-5 (Smith et al., 2014). The location of Southern, Central and Northern Core
763 complexes are also indicated.
764
765 **Figure 2. (A)** Location of the dredges collected during the cruise, R/V Knorr, KNR210-5
766 (Smith et al., 2014). Large pie diagrams refer to the dredges where gabbros or peridotites were
767 collected, dredge numbers are indicated. **(B)** Compositional variability of all lithologies and
768 plutonic rocks from Southern , Central (CC) and Northern Core Complexes (NCC).
769
770 **Figure 3.** Photographs of cut hand specimens of six different plutonic rocks recovered by
771 dredging during R/V Knorr cruise 210-5. Some of the samples show high-temperature crystal-
772 plastic deformation with protomylonitic, mylonitic and ultramylonitic fabrics. In one case, the
773 high-temperature deformation fabrics are cross-cut by lower-temperature brittle fractures and
774 amphibole veins. CPF refers to crystal-plastic fabric as defined in the text.
775
776 **Figure 4.** Microphotographs of selected samples. **(A)** Olivine oikocrysts showing rounded
777 grain boudaries against plagioclase in an undeformed troctolite; **(B)** Large clinopyroxene
778 oikocryst including subrounded olivine and plagioclase chadacryst in a troctolite; Ti-pargasitic
779 amphibole in a sheared troctolite (CPF 1) showing interstitial **(C)** and oikocrystic **(D)** habitus;
780 **(E)** Sheared gabbronorite cut by amphibole vein with diffuse boundaries. Clinopyroxene
781 locally shows partial replacement by amphibole (CPF 2); **(F)** Hornblende corona statically
782 replacing clinopyroxene in the vicinity of the amphibole vain in **(E)**. Ol, olivine; Pl,
783 plagioclase; Cpx; clinopyroxene; Ti-Prg, Ti-pargasite; Hbl, horneblende.
784
785 **Figure 5**. Histogram of deformation intensity of plutonic rocks collected during R/V Knorr
786 cruise 210-5 (see text for detail).
787
788 **Figure 6.** Macroscopic occurrence and microphotographs of the amphibole-plagioclase
789 ultramylonites and host mylonites. (**A**) Troctolite showing a gradual increase in deformation
790 intensity from protogranular to mylonite to ultramylonite (UM). (**B**) Scanned thin-section
791 image of the ultramylonite band crosscutting an oxide-gabbro mylonite. Note that the texture
792 and the mineralogical composition of the two ultramylonites are almost identical, although they
793 are developed in highly different magmatic protoliths.
794
795 **Figure 7.** FEG-SEM compositional maps of selected samples. (**A**) Na, Ca and Mg map of a
796 sheared troctolite clearly showing an increase in Na and decrease in Ca in the neoblastic
797 plagioclase compared to the coexisting porphyroclasts. (**B**) Cr, Ca, Mg, S and Ti maps of the
798 troctolite mylonite including a Amp-Pl ultramylonitic band. The maps show a decrease in Ca in
799 plagioclase in the ultramylonite compared to the mylonite and the distribution of Ti-bearing
800 phases (i.e., amphibole) and Cr-bearing phases (i.e., spinel) between the two lithologies. (**C**) Na
801 and Ti map of the oxide-gabbro mylonite including a amphibole-plagioclase ultramylonitic
802 band. The maps show a sharp decrease in Na in plagioclase in the ultramylonite compared to
803 the mylonite and the lack of Ti-Fe oxides in the ultramylonite. (**D**) Detail of a Na, Mg and Fe
804 map of a gabbro crosscut by an amphibole vein. Note that hornblende replacement is
805 widespread in the vicinity of the amphibole vein, and that Na-rich plagioclase is associated
806 with micro-cracks and late fractures.
807
808
809 **Figure 8.** Variations in TiO_2 (**A**) and Cr_2O_3 (**B**) versus $Mg/(Mg+Fe^{2+})$ (mol.%) in
810 clinopyroxenes from the 16.5ºN gabbros. Symbols are grouped per rock-type and indicate
811 single spot analyses (core and rim). The composition of clinopyroxenes from gabbros from
812 other lower crustal sections exposed along the Mid-Atlantic Ridge are also depicted (Ross and
813 Elthon, 1993; Nonnotte et al., 2005; Coogan et al., 2002; Suhr et al., 2008; Drouin et al., 2009;
814 Lissenberg and Dick, 2008; Dick et al., 2010).

815
816 **Figure 9.** Covariations in clinopyroxene $Mg/(Mg+Fe^{2+})$ (mol.%) versus $Ca/(Ca+Na)$ (mol.%)
817 in plagioclase and $Mg/(Mg+Fe^{2+})$ (mol.%) in orthopyroxene coexisting in selected gabbros.
818 Dashed arrows indicate the chemical variation in clinopyroxene and plagioclase in the
819 ultramylonite compared to the host oxide-gabbro mylonite. Data are averaged per sample, error
820 bars represent one standard deviation of the mean value.
821
822 **Figure 10.** Compositions of amphibole from selected gabbros. (**A**) Variations in alkali (Na+K)
823 and TiO_2 (wt.%) versus Al^{IV} (apfu) in amphibole from selected gabbros. (**B**) Same plot as A
824 including only core analyses of pargasite and brown horneblende. (**C**) Chlorine (wt.%) versus
825 Al^{IV} (apfu) in amphibole; detection limit of EPMA analyses is also indicated. (**D**) Variations in
826 TiO_2 (wt.%) versus $Mg/(Mg+Fe^{2+})$ (mol.%) in pargasite and brown-hornblende (core analyses).
827 Each symbol represents a single spot analysis.
828
829 **Figure 11.** (**A**) Compositional distribution of basalts from Gakkel Ridge (~500 glass analyses);
830 Mid-Atlantic Ridge 10–35 ºN (~1300 glass analyses); East Pacific Rise 9–13 ºE and East
831 Pacific Rise 9–13 ºN (~230 glass analyses) and East Pacific Rise 14–20 ºN (~1050 glass
832 analyses). Data compiled from PetDB (Lehnert et al., 2000). (**B**) Average compositions and
833 standard deviation of Na_8 (Klein and Langmuir, 1987) versus $Mg/(Mg+Fe)$ of the basalts in
834 (A). The average full spreading rates of the regions are also indicated (see also Sinton and
835 Detrick, 1992 and Coogan, 2007).
836
837 **Figure 12.** Visual explanation of the quantitative model. M1 is the primary melt with
838 $Mg/(Mg+Fe)$ ~72 and mass 100g. M2 to M13 are melts produced by the model at steps of
839 Mg#~4. The mass of melt at each step is calculated as indicated by the italic numbers. B1 to
840 B12 are melt extracted at each step of the model. Their mass is scaled to reproduce the actual
841 compositional variability of melts erupted at the MAR (10-35ºN). G1 to G13 are gabbros
842 produced by the model, and forming the "cumulate". T1 to T12 are melts trapped at each step
843 and subtracted form the total melt mass. T_{tot} is the sum of T1 to T12 and is used to calculate the
844 proportion of gabbros crystallized from trapped melts ("trapped"). The "Total" lower crust
845 composition is given by the sum of gabbros crystallized as "cumulate" + "trapped" (see text for
846 details).
847
848 **Figure 13.** Liquid $Mg/(Mg+Fe)$ (mol.%) compositions versus (**A**) olivine Fo (black line and
849 red circles) and clinopyroxene $Mg/(Mg+Fe)$ (blue line and blue circles); (**B**) mass of gabbros
850 (%) crystallized at each step of the thermodynamic model at different mass erupted /mass
851 crystallized ratios (Me/Mc); (**C**) trapped melt (%) and (**D**) liquid density calculated at each step
852 of the crystallization process. (**E**) Compositional variability of gabbros resulted from the
853 thermodynamic model. The origin represents the mass of gabbros neglecting eruption, on the x-
854 axis are depicted the model outcomes for different Me/Mc ratios if melt entrapment is ignored.
855 The effect of melt entrapment is shown in the pie diagrams displayed vertically. The vertical
856 displacement of each diagram depends on the total amount of trapped melt, as represented on
857 the y-axis. The grey dashed line connects model results for different trapped melt compositions
858 (see text for details). The compositional variability of the gabbros dredged at the MAR 16.5ºN
859 region, and those from the hole U1309 at the Atlantis Massif (Blackman et al., 2011) are also
860 reported. (**F**) Compositional variability of the basalts erupted at different Me/Mc ratios
861 compared to the basalts from the MAR 10–35ºN segment.
862
863 **Figure 14.** Temperature estimates calculated using the edenite-richterite amphibole-plagioclase
864 thermometer (Holland & Blundy,1994) assuming a pressure of 200 MPa. Mean values and
865 standard deviations are given for individual textural amphibole-plagioclase types for each
866 sample.
867
868

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Figure 1.
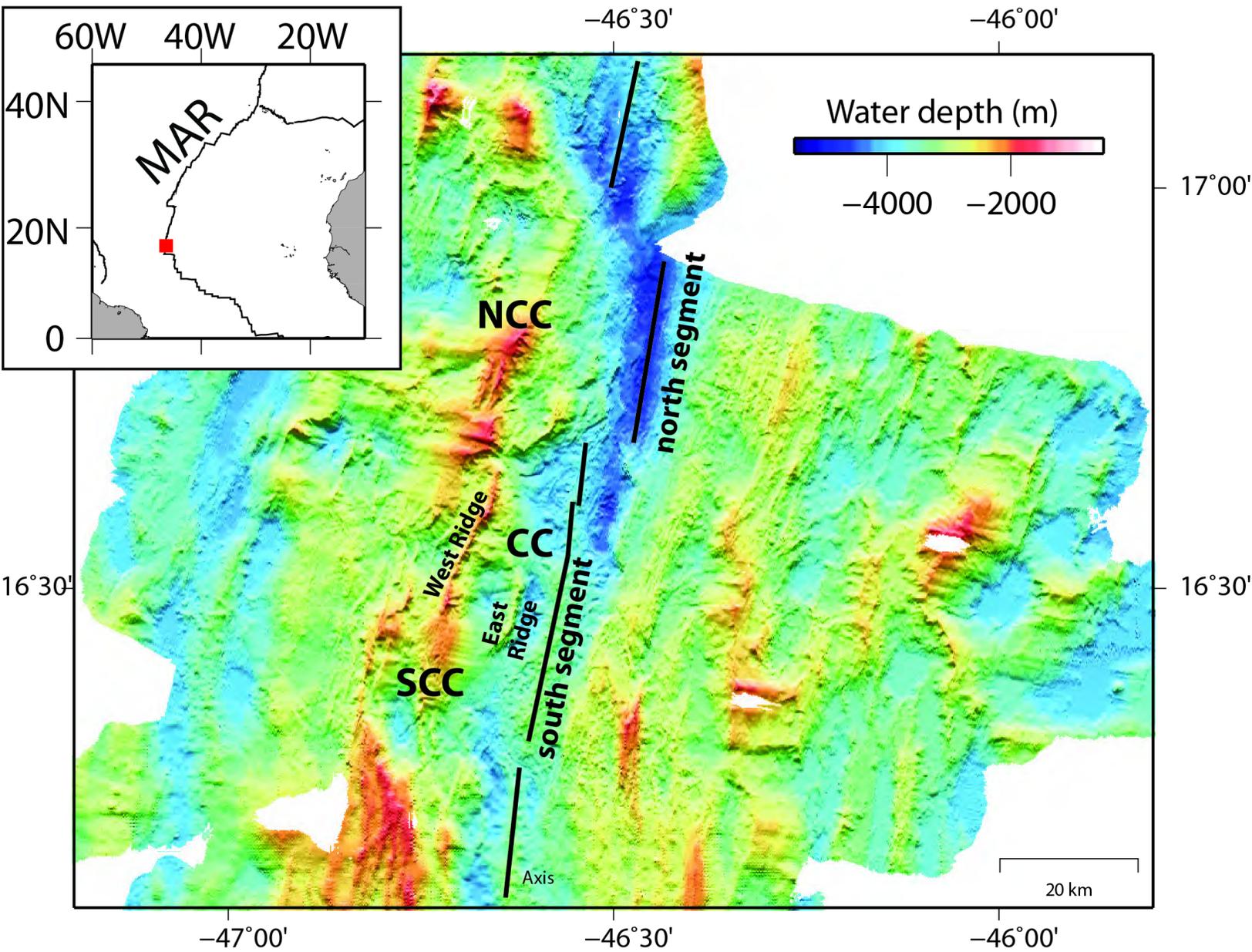
Figure 2.

A.
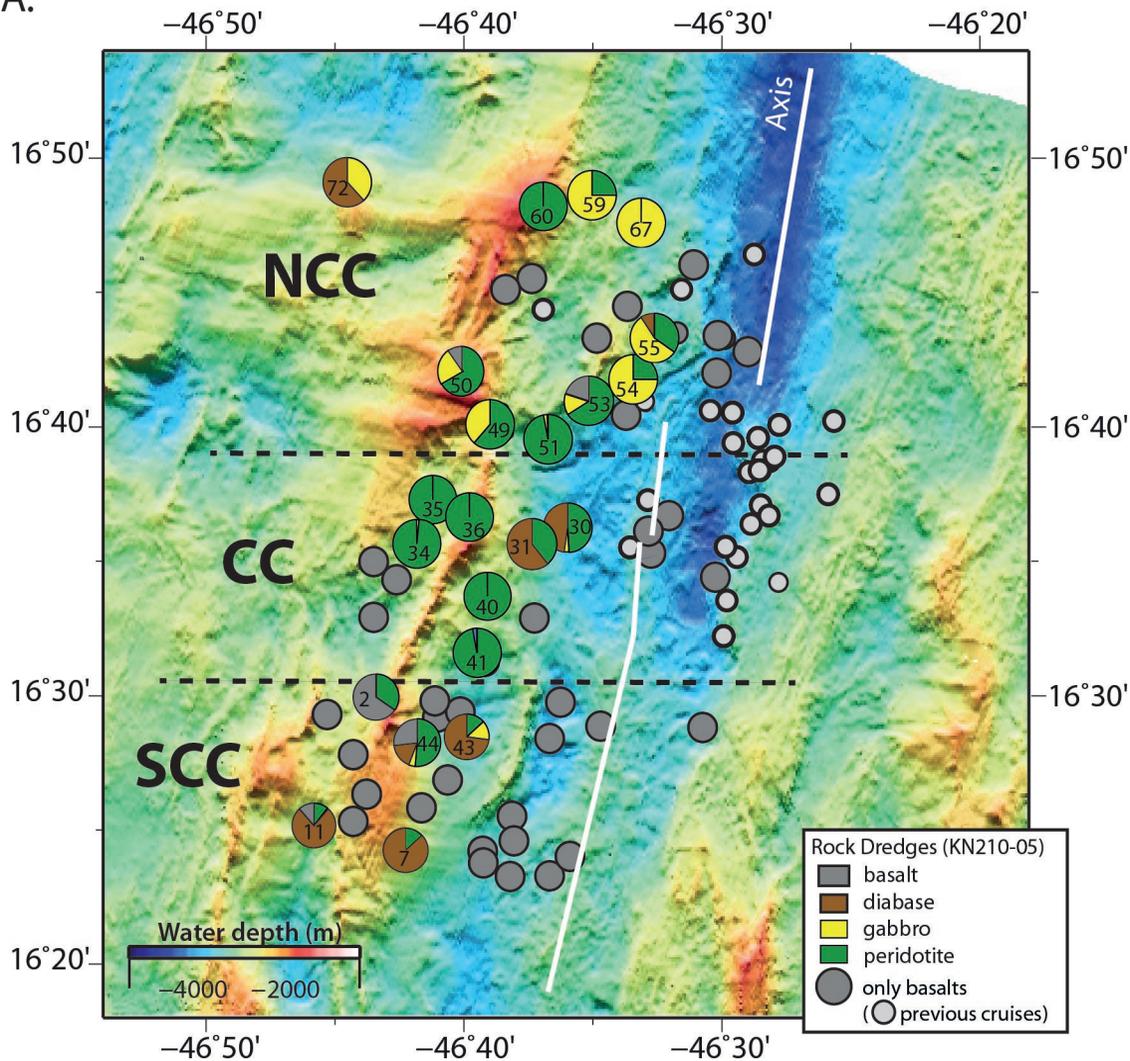
B.

Nothern Core Complex (NCC)

Tot. 1033 kg

Tot. 241 kg

Central Core Complex (CC)

Tot. 953kg

Tot. 6.00 kg

Southern Core Complex (SCC)

Tot. 2620 kg

Tot. 1.25 kg

Rock Dredges (KN210-05)
- basalt
- diabase
- gabbro
- peridotite
- only basalts
  (○ previous cruises)

Water depth (m)
−4000   −2000
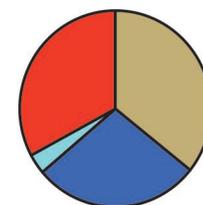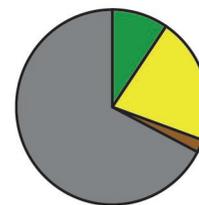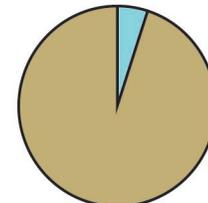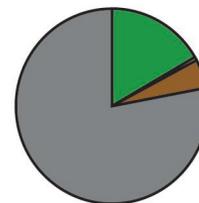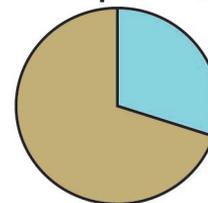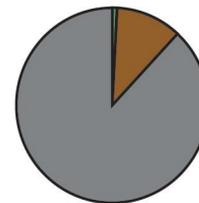
Basalt
Diabase
Gabbro
Peridotite

Dunite
Troctolite
Olivine gabbro
Gabbros/norite
Fe-Ti oxide gabbro

Axis

NCC

CC

SCC

Figure 3.

A — Dunite

B — Troctolite — CPF 1

C — Knr210-5-Dr. 60-27 — Troctolite — CPF 3 — CPF 4 — CPF 5 — Ultramylonite

D — Knr210-5-Dr. 55-44 — Ultra-mylonite — CPF 5 — CPF 4 — CPF 3 — Oxide-gabbro

E — CPF 2 — Oxide-gabbro (deformed) — CPF 0 — Gabbronorite

F — Knr210-5-Dr. 49-18 — CPF 2 — Gabbro — Amp vein

Figure 4.

A — Ol oik, Pl, 1 mm

B — Cpx oik, Ol, Pl, 0.5 mm

C — Pl, Ti-prg, Ol, 0.5 mm

D — Ti-prg, Pl, Pl, 0.5 mm

E — Amp-vein, Pl, Cpx, 5 mm

F — Pl, Hbl, Cpx, 1 mm

Figure 5.

Figure 6.

a) Troctolite / Amp-Pl Ultramylonite

Protomylonite

Mylonite

UM

Ol+Pl ± Cpx ± Ti-Prg

Ol+Pl ± Cpx ± Ti-Prg

Pl + Ti-Prg (± Cpx)

Pl + Ti-Prg (± Cpx)

b) Ox-gabbro / Amp-Pl Ultramylonite

Mylonite

UM

Pl ± Cpx ± Ti-Fe Ox

Pl

Ti-Fe Ox

Hbl

Cpx

Pl + Ti-Prg ± Cpx
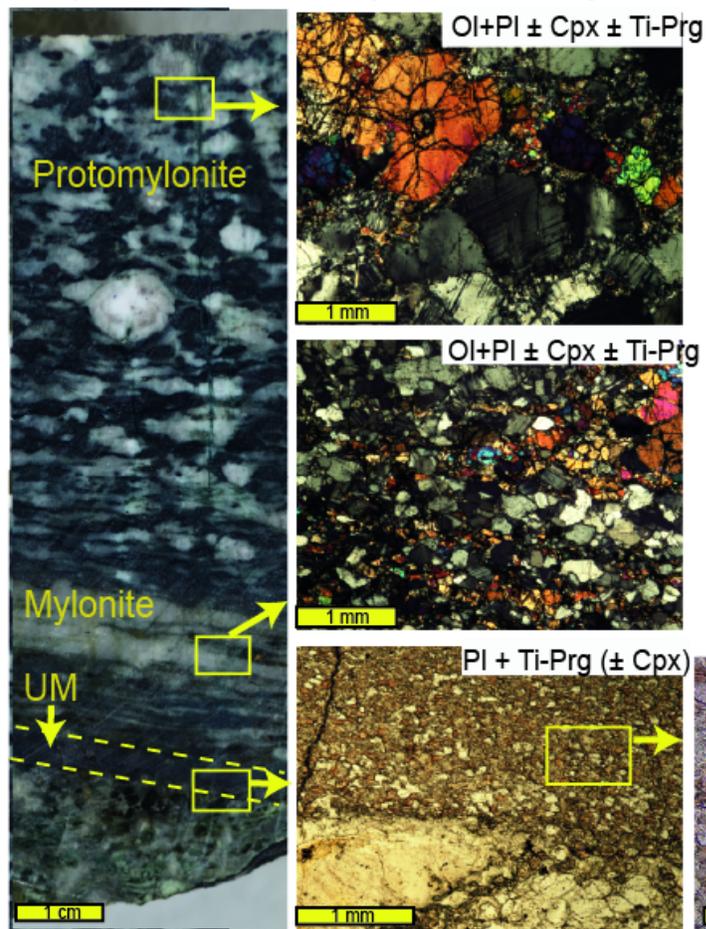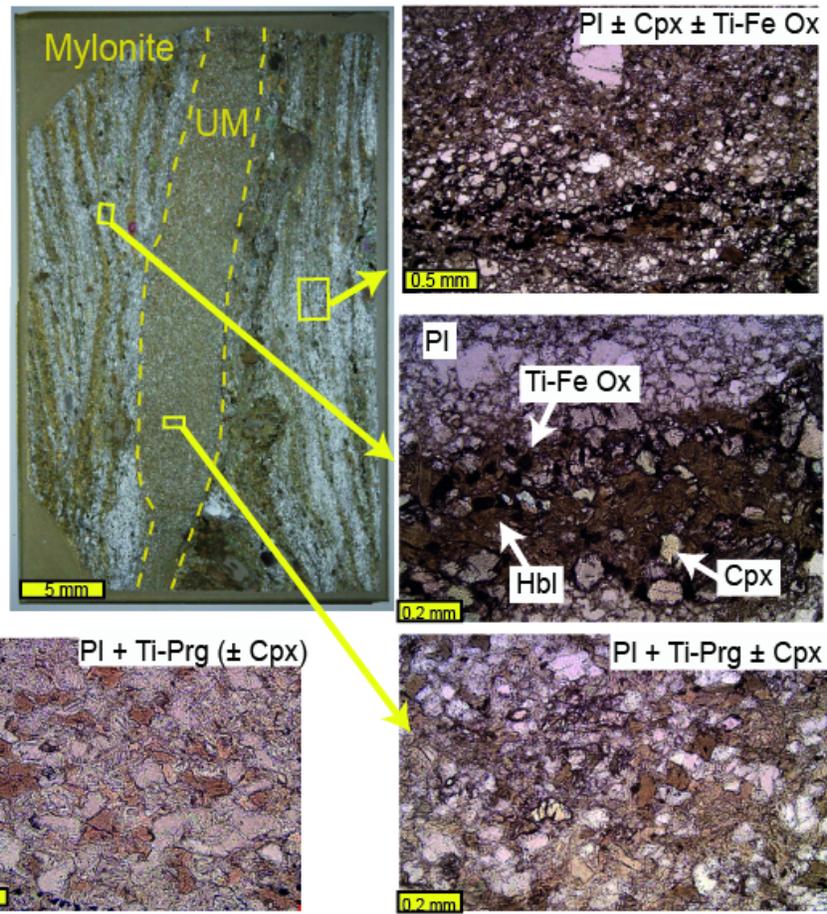
Figure 7.

a) Sheared troctolite

Na Ca Mg

Pl neo (An~78)

Pl porp (An~81)

Cpx

Ol

5 mm

b) Troctolite / Amp-Pl Ultramylonite

Cr Ca Mg S

Ti

Troctolite mylonite
(An ~81)

Amp-Pl UM (An ~65)

5 mm

c) Ox-gabbro / Amp-Pl Ultramylonite

Ox-gabbro Mylonite
(An ~46)

Amp-Pl UM (An ~82)

Na Ti

5 mm
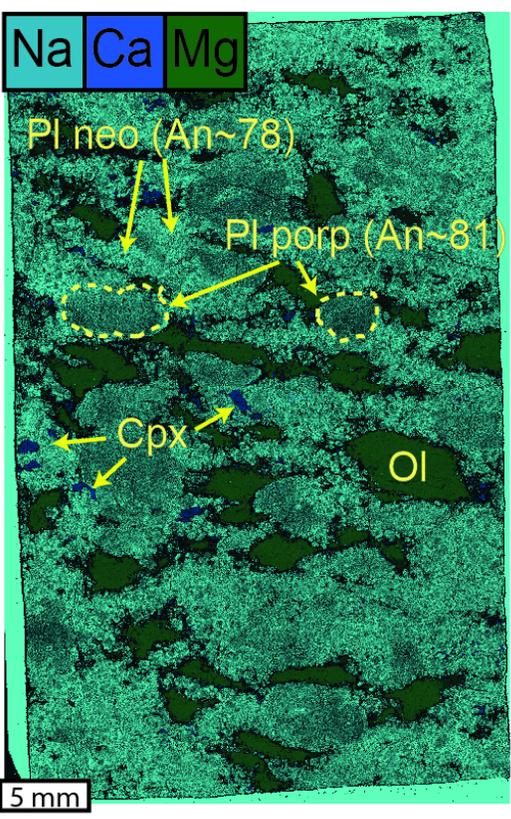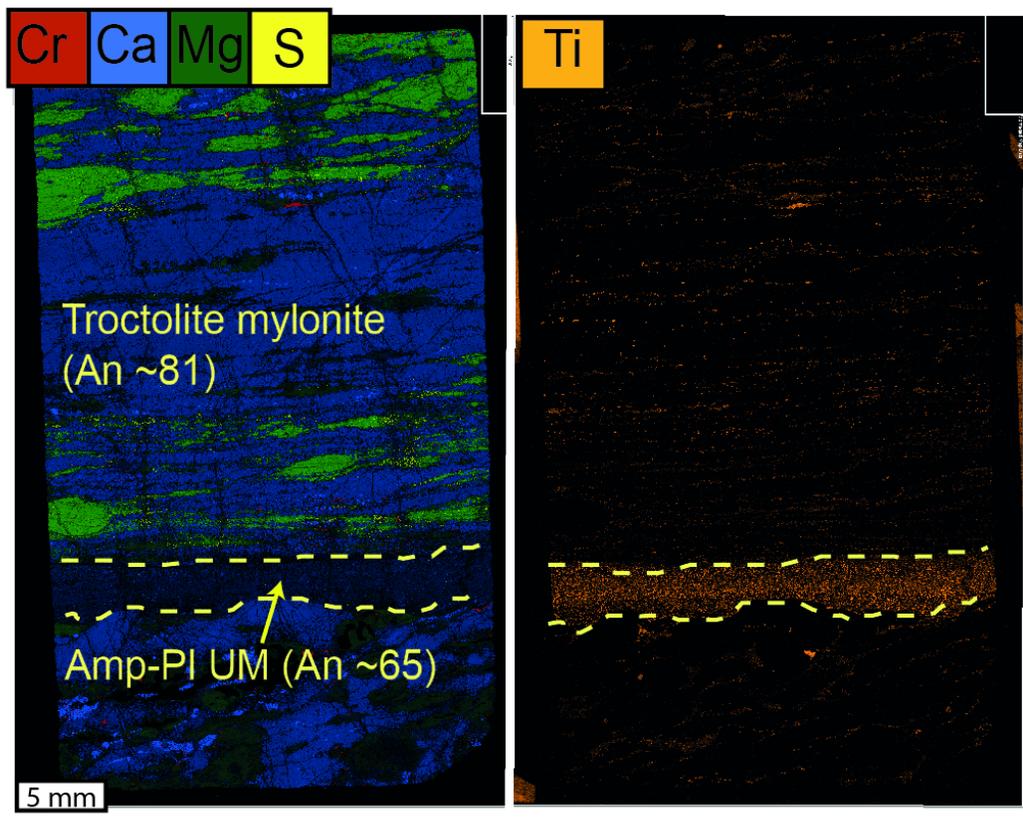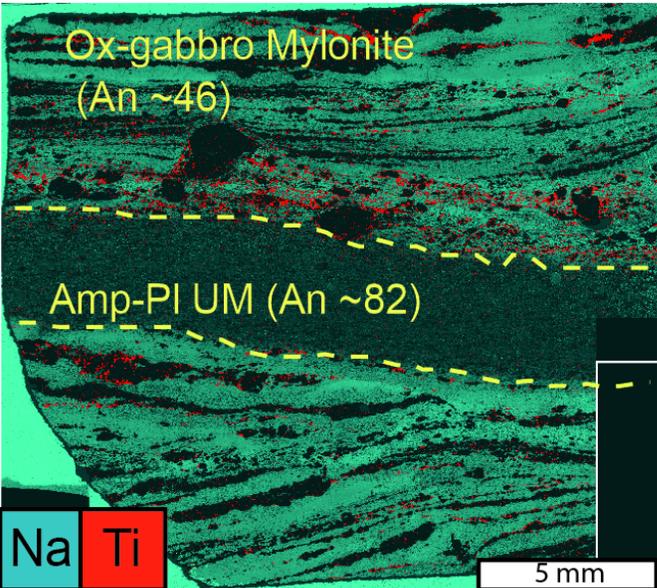
d) Amp-vein on gabbronorite

Na Mg Fe

Amp-vein

vein Pl (An 53)

magmatic Pl (An ~58)

Hbl

Opx

secondary Pl (An 25)

5 mm

Figure 8.

A. Clinopyroxene

Wt.% TiO_2

B.

Wt.% Cr_2O_3

Troctolites
Gabbros
Oxide-gabbros
Gabbronorites
Mid Atlantic Ridge
(other sections)

Mg/(Mg+Fe^{2+}) mol%

Figure 9.

A. (y-axis) Plagioclase Ca/(Ca+Na) mol%

B. (y-axis) Orthopyroxene Mg/(Mg+Fe²⁺) mol%

(x-axis) Clinopyroxene Mg/(Mg+Fe²⁺) mol%

Ultramylonite

Troctolites
- 60-27 porph
- 60-30 porph
- 59-15 core
- 60-27 neob My

Gabbro (Amp-vein)
- 49-18 relict (Amp)

Gabbronorites
- 67-1a core
- 49-22 core

Ox-gabbros
- 55-34 porph
- 55-34 neob
- 55-44b porph
- 55-44b neob My
- 55-44b neo Umy

Figure 10.

a) All amphiboles
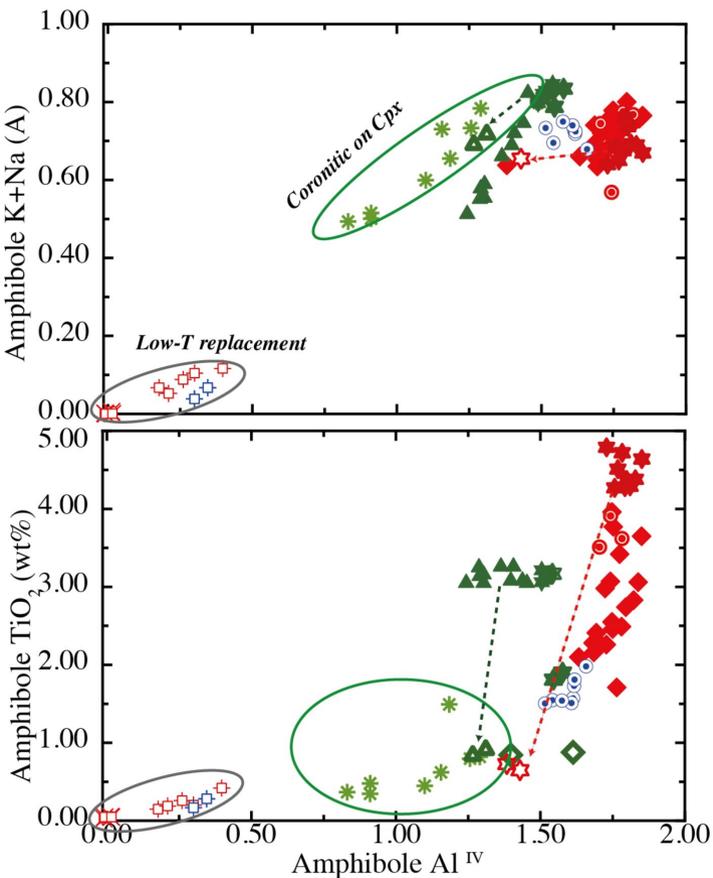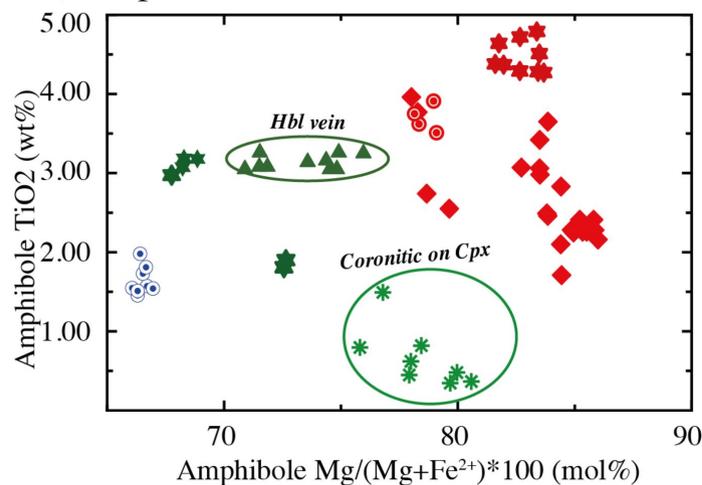
b) Amphibole core
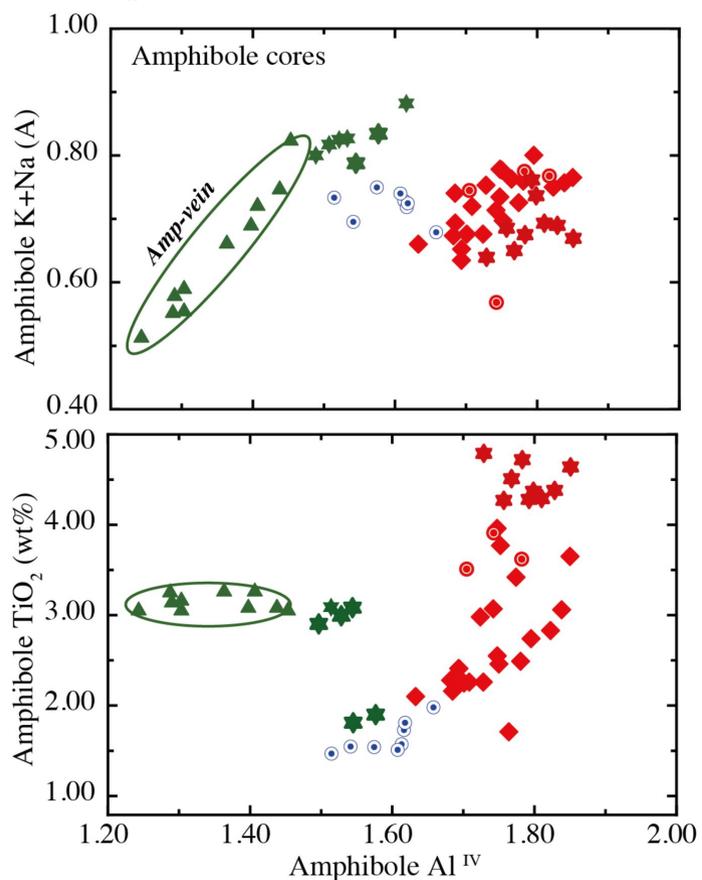
c) Amphibole core
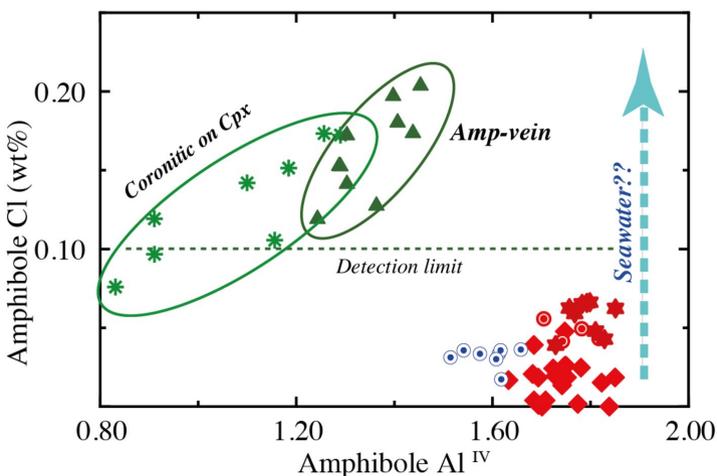
d) Amphibole core

Troctolites
★ Magmatic core
☆ Magmatic rim
◆ Neob My
◉ Neob UM

Gabbro (Amp-vein)
★ Interstitial
▲ Hbl Vein Core
△ Hbl Vein rim
✳ Replacing Cpx

Ox-gabbro
◉ Neob UM

Low-T replacement
⊠ Replacing Ol (Troct)
⊞ Replacing Amp (Troct)
⊞ Replacing Amp (Gabbro)

Figure 11.

A.
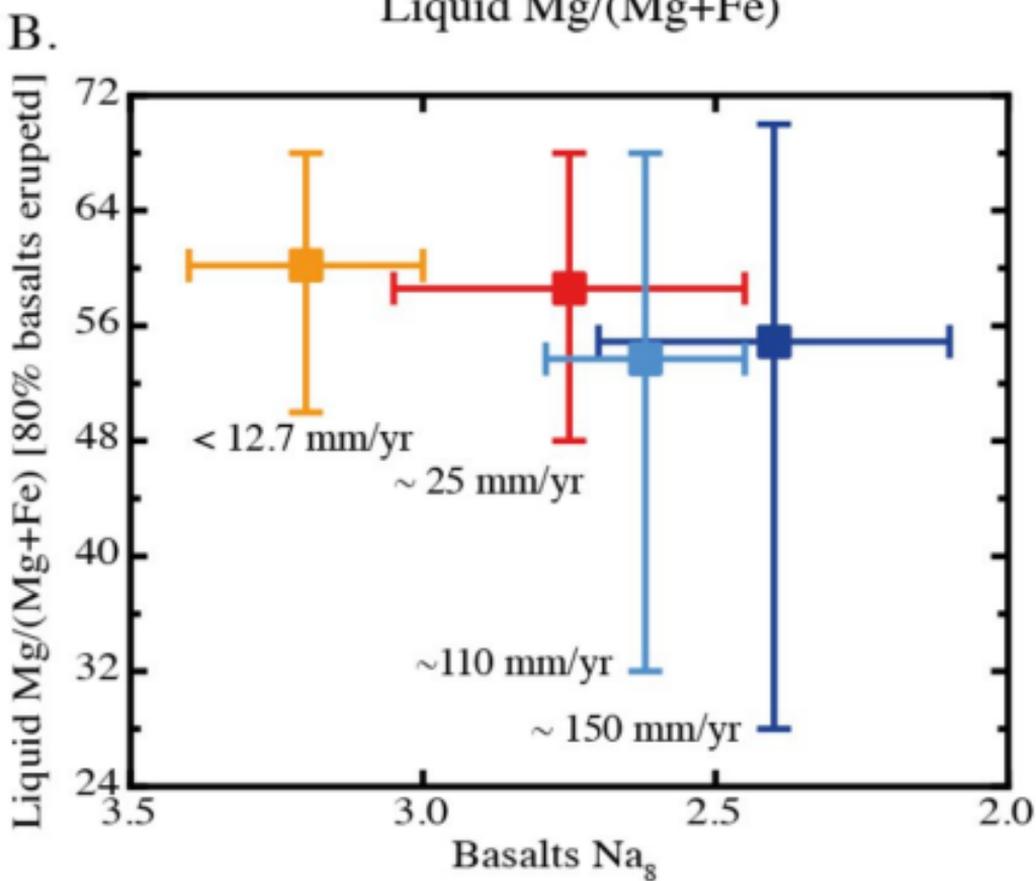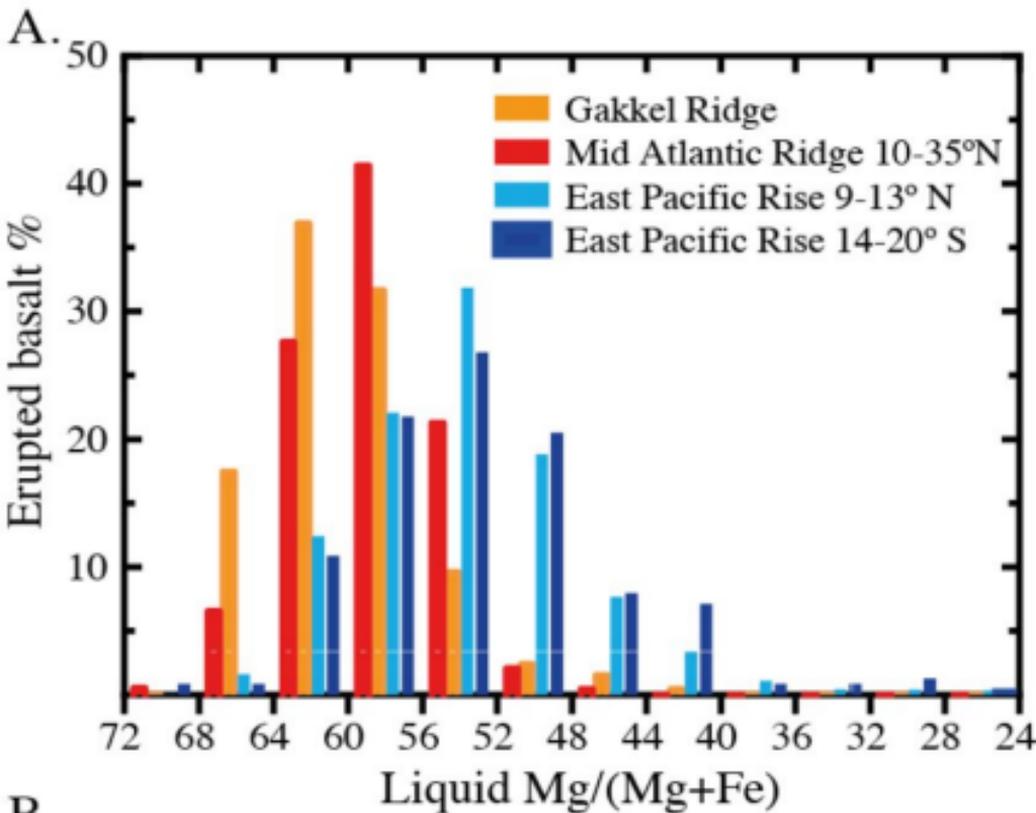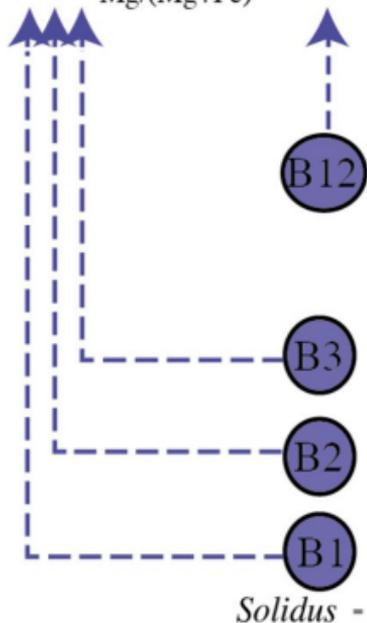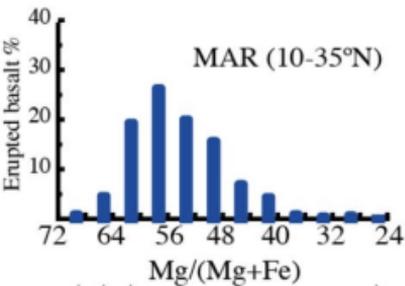
B.

Figure 12.

Basalts

MAR (10-35ºN)

Eruped basalt % (y-axis, 0 to 40)

Mg/(Mg+Fe) (x-axis, 72 to 24)

B12

M13 *(24)*
M13=M12-G12-B12-T12
(~5 g)

M4 *(60)*
M4=M3-G3-B3-T3

B3

M3 *(64)*
M3=M2-G2-B2-T2

B2

M2 *(68)*
M2=M1-G1-B1-T1

B1

Solidus

M1 *(72)*   *M1(100g)*

Lower crust

Cumulate   Trapped   Total
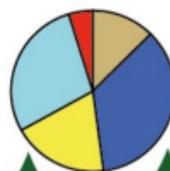
$T_{tot} = \Sigma T_{(1-12)}$

G12   T12

G3   T3

G2   T2

G1   T1

Dunites (tan)
Troctolites (dark blue)
Olivine gabbros (yellow)
Gabbros/norites (light blue)
Fe-Ti oxide gabbros (red)

Figure 13.

A

Plagioclase
Olivine
Clinopyroxene
Ti-Fe oxides

troctolite/Ol-gabbro
gabbro
gabbronorite
Ox-gabbro

Ol   Model
Cpx
Fo mol   16.5°N MAR
Cpx Mg#

Ol and Cpx Mg/(Mg+Fe) (mol%)

B

FC model
Me/Mc ~1:4
Me/Mc ~2:3
Me/Mc ~3:2
Me/Mc ~7:3

Crystalized mass (%)

C

Trapped melt %

D

Liquid density (g/cm³)

Liquid Mg/(Mg+Fe) (mol%)

E Predicted lower crust composition

Dunites
Troctolites
Olivine gabbros
Gabbros/norites
Fe-Ti oxide gabbros

Atlantis Massif

16.5°N

Trapped melt (%)

FC

Mass erupted/mass crystallized (Me/Mc)

F Basalt distribution

MAR 10-35°N

Me/Mc ~1:4
Me/Mc ~2:3
Me/Mc ~3:2
Me/Mc ~7:3

Erupted basalt %

Liquid Mg/(Mg+Fe) (mol%)

Figure 14.

magmatic

granulite-facies
mylonite

Cpx-Opx (L&F, '92)

Am-Pl ultramylonite

Am vein
cross-cutting gabbro

Static replacement of
Cpx porphyroclasts

Am-Pl temperature (°C at 200 MPa)

Troctolites
60-30
60-27a
60-27c
Gabbro
49-18
Ox-gabbro
55-44