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1 **Strain localization and fluid infiltration in the mantle wedge during**
2 **subduction initiation: evidence from the base of the New Caledonia ophiolite**

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23 **Abstract**

24 Despite decades of petrological and geochemical studies, the nature and
25 setting of obducted ophiolites remain controversial: the influence of supra-subduction
26 zone environments on pre-existing oceanic lithosphere is yet to assess, and the
27 processes leading to subduction/obduction initiation are still poorly constrained. Our
28 study documents successive influx of slab-derived fluids and progressive strain
29 localization within the upper mantle in a supra-subduction environment during the
30 first few My of the subduction history. We focus on strongly sheared mafic
31 amphibolites intruding peridotites near the mantle–crust transition of the New
32 Caledonia obducted ophiolite and ~50 to 100 m above the basal thrust contact of the
33 ophiolite. These m- to hm-long and several m-thick shear bands are interpreted as
34 inherited small-scale intrusions of mafic melts, probably dikes or sills, which were
35 derived from a moderately refractory mantle source refertilized by supra-subduction
36 zone fluids. $^{40}\text{Ar}/^{39}\text{Ar}$ age constraints on pargasite at ca. 90 Ma suggest that they
37 could be inherited from the former Pacific west-dipping subduction.

38 Secondary deformation of these mafic intrusions is intimately associated to
39 three major stages of fluid infiltration: (1) the first stage of deformation and
40 metasomatism is marked by syn-kinematic growth of Ca-amphibole (at 700–800 °C
41 and 3–5 kbar) with a distinctive supra-subduction zone signature, and controlled later
42 channelization of aqueous fluids. $^{40}\text{Ar}/^{39}\text{Ar}$ dating on magnesio-hornblende indicates
43 that this deformation episode occurred at ca. 55 Ma, coincident with east-dipping
44 subduction initiation; (2) the main metasomatic stage, characterized by the
45 development of a phlogopite-rich matrix wrapping peridotites and amphibolite
46 boudins, points to the percolation of alkali-rich aqueous fluids at still high

47 temperature (650–750 °C); (3) the last, low temperature (< 600 °C) metasomatic stage
48 results in the formation of deformed veinlets containing talc, chlorite and serpentine.

49

50 **1. Introduction**

51 Obduction emplaces fragments of dense oceanic lithosphere on top of
52 continents over several hundreds of kilometers (e.g. Oman, Newfoundland, Turkey,
53 New Caledonia; Coleman, 1971). Despite numerous petrological and geochemical
54 studies, the exact nature and setting of many obducted ophiolites remain unclear. The
55 Oman ophiolite is a typical example where there are evidences for both a MORB-type
56 signature (Boudier et al., 1988; Ceuleneer et al., 1988; Nicolas et al., 2000; Godard et
57 al., 2000, 2003; Le Mée et al., 2004) and a supra-subduction zone (SSZ) geochemical
58 imprint (Ernewein & al, 1988, Stern and Bloomer, 1992; Shervais, 2001; McLeod et
59 al., 2013).

60 The origin of obducted ophiolites has long been debated (Coleman, 1971,
61 Dewey and Bird, 1971; Nicolas, 1989; Moores et al., 2000), with recent studies
62 focusing on subduction-obduction initiation and early emplacement of ophiolites
63 (Vaughan and Scarrow, 2003; Agard et al., 2007, 2014; Lagabrielle et al., 2013).
64 Most of our knowledge on the initiation of the subduction-obduction system comes
65 from the amphibolite to granulite facies metamorphic soles found at the base of most
66 large-scale non-metamorphosed ophiolites and from magmatic dikes emplaced at
67 different levels of the mantle sequence. These dikes generally record partial
68 refertilization of the mantle wedge by subduction-derived fluids and show little
69 deformation (e.g. Pirard, 2012; Xiong et al., 2014) whereas metamorphic soles are
70 highly deformed portions of upper oceanic crust and sediments, heated and

71 metamorphosed along the interface between the subducting slab and the hotter mantle
72 wedge during early subduction (e.g. Wakabayashi and Dilek, 2003). Metamorphic
73 soles typically predate the final emplacement of the ophiolite onto the continental
74 margin by several millions of years (~20 My for Oman; Agard et al., 2007).

75 We document sheared mafic amphibolite and feldspar-rich veins outcropping
76 over a scale of hundreds of meters, within the peridotites of the New Caledonia
77 ophiolite. They occur near the crust–mantle transition zone (e.g. Prinzhofer et al.,
78 1980; Pirard et al., 2013) and ~50 to 100 m above the basal contact of the ophiolite,
79 south east of Noumea (Fig.1). As these deformed, metamorphosed rocks occur at the
80 base of the New Caledonia ophiolite, it is tempting to relate them to subduction
81 initiation, as for metamorphic soles. These sheared amphibolites could represent
82 metamorphic sole fragments (outcropping in the Thio area; Cluzel et al., 2001, 2012)
83 tectonically imbricated with basal peridotites, or be derived from melt products
84 triggered by subduction-derived fluids (e.g. as the SSZ amphibole-bearing dikes in the
85 Massif du Sud peridotites; Cluzel et al., 2006; Pirard, 2012; Pirard et al., 2013).
86 Alternatively, they could also be inherited from earlier melting processes during
87 ocean basin formation (e.g. near transform fault; Gaggero and Cortesogno, 1997;
88 Constantin, 1999).

89 In this study, we provide a detailed petrological, geochemical and
90 geochronological study to constrain the nature, the tectonic setting and the origin of
91 these deformed amphibole-bearing dikes, and discuss implications for the New
92 Caledonia subduction system in terms of geological setting and refertilization.

93

94 **2. Geological setting**

95 New Caledonia is located at the northern termination of the Norfolk Ridge, a
96 stretched continental fragment rifted off the eastern Australian margin during the
97 Early Cretaceous rollback of the Western Pacific subduction zone and the opening of
98 New Caledonia and Loyalty basins (Weissel and Hayes, 1977; Van de Beuque et al.,
99 1998; Schellart et al. 2006).

100 The present-day structure of New Caledonia (Fig. 1a) results from a Paleogene
101 episode of convergence. This event led to the closure of the South Loyalty Basin after
102 the initiation of a north-east-dipping intra-oceanic subduction (Cluzel et al., 2001;
103 Schellart et al., 2006; Matthews et al., 2015) and was responsible for the obduction of
104 the New Caledonia ophiolite rooted in the Loyalty Basin (Collot et al., 1987).
105 Geochronological and thermochronological data on mafic rocks from the
106 metamorphic sole (Cluzel et al. 2012) and mafic dikes intruding the ultramafic
107 ophiolite (Cluzel et al. 2006; Pirard, 2012) suggest that subduction inception occurred
108 at ca. 56 Ma. At ca. 34 Ma, this intra-oceanic subduction led to the obduction of the
109 South Loyalty Basin lithosphere toward the south-west onto the Norfolk Ridge (e.g.
110 Paris, 1979; Prinzhofer, 1981; Collot et al., 1987).

111 The ophiolite complex of New Caledonia is composed of three allochthonous
112 units of Cretaceous–Eocene age accreted over a calc-alkaline, island-arc derived
113 basement (Fig.1a) (Aitchison et al., 1995; Cluzel et al., 2001). This basement was
114 created from the Permian to the Early Cretaceous with the accretion of three major
115 volcano-sedimentary terranes (i.e. Teremba, Boghen and Koh terranes) along the
116 eastern Gondwana margin. Above this basement, the Tertiary subduction complex is
117 composed of four main units:

118 1 — a Tertiary high-pressure metamorphic belt at the northern end of New-
119 Caledonia (Fig.1a) made of volcano-sedimentary and mafic blueschist and eclogite
120 facies rocks, some of which are interbedded in a serpentinite matrix and minor
121 metasediments (Clarke et al., 1997; Carson et al., 2000; Cluzel et al., 2001; Vitale-
122 Brovarone and Agard, 2013). Mafic eclogites have Late Cretaceous to Eocene
123 protolith ages and share geochemical affinities with the Poya terrane (Cluzel et al.,
124 2001; Spandler et al., 2005)

125 2 — the Poya terrane, a weakly metamorphic mafic rocks unit (e.g. Eissen et al.,
126 1998; Cluzel et al., 1997, 2001), outcrops along most of the west coast of New
127 Caledonia and also forms several isolated exposures along the eastern side (Fig.1a).
128 This terrane consists mainly of massive and pillow basalt, dolerite, and gabbro
129 associated with chert and volcanoclastic rocks (Eissen et al., 1998). The origin of the
130 Poya Terrane is controversial. Geochemical associations of back-arc basin and
131 enriched-MORB signatures constrain its formation to a back-arc environment (Cluzel
132 et al., 1997, 2001). Paleontological studies date its formation between 83 and 55 Ma
133 (Cluzel et al., 2001). This unit was interpreted as representing the crustal cover of the
134 South Loyalty Basin (Spandler et al., 2005; Lagabrielle et al., 2013), with the above
135 ages likely indicating the opening period of the basin.

136 3 — the ultramafic ophiolite (named “Nappe des Péridotites”) found on the
137 south-eastern portion of New Caledonia (Fig.1a) is constituted mainly of a 2–3 km
138 thick depleted mantle sequence (spinel-bearing harzburgite) above a 20 to 100 m
139 thick porphyroclastic serpentine mylonite. In the northern part of New Caledonia,
140 lherzolite is present in the Tiebaghi and the Poum massifs (e.g. Moutte, 1979; Ulrich
141 et al., 2010). This suite of ultramafic rocks is interpreted as representing upper mantle
142 rocks of the South Loyalty Basin (Lagabrielle et al., 2013) and documents a complex

143 history with several supra-subduction zone settings, as attested by magmatic
144 intrusions.

145 In the Massif du Sud, Marchesi et al. (2009) and Pirard et al. (2013) reported
146 gabbro norites and olivine gabbro sills at the top of ophiolitic sequence. These gabbros
147 represent the lower crust of the ophiolite. They are separated from the underlying
148 harzburgite by a dunite complex composed of dunite channels with the presence of
149 dunite cumulates, wherlites, clinopyroxenite and websterite sills and dikes. Amphibole
150 is commonly present in clinopyroxenite bodies but absent from the lower crust
151 gabbro norites (Pirard, 2012; Pirard et al., 2013). These authors interpret this complex
152 as the crust–mantle transition with mafic sills and dikes being the result of
153 percolating, hydrous primitive arc magmas within the peridotite during early stages of
154 the subduction system and nascent arc crust formation.

155 The Massif du Sud is also extensively cross-cut by Early Eocene felsic,
156 anhydrous and hydrous coarse grained dikes (leucodiorite, granitoid, micro-diorite,
157 dolerite and hornblende; Prinzhofer, 1981; Cluzel et al., 2005, 2006; Pirard, 2012;
158 Pirard et al., 2013) discordant with the harzburgite and/or the gabbro sills from the
159 lower crust. These dikes are generally unmetamorphosed and undeformed with the
160 exception of few diorite dikes which are boudinaged, locally micro-folded and where
161 amphibole shows a preferred orientation (Cluzel et al., 2012). Upon geochemical and
162 geochronological features, these dikes have been interpreted as formed in a supra-
163 subduction zone setting. These dikes are essentially considered as synchronous or
164 immediately posterior to magma injection in active faults that formed during the first
165 stages of the 55 Ma subduction (Cluzel et al., 2006, 2012). However, Prinzhofer
166 (1981) obtained 2 different K-Ar ages (i.e. 90 ± 10 Ma and 42 ± 5 Ma) for dolerite
167 dikes intruding the Massif du Sud. The youngest dikes coincide with the Eocene New

168 Caledonia subduction. The Late Cretaceous dikes could be an evidence of the ocean
169 spreading stage of the South Loyalty Basin (Prinzhofer, 1981; Matthews et al., 2015).
170 The Late Cretaceous age of 85 Ma was also measured in zircon cores from the high-
171 pressure Pouebo unit (Spandler et al., 2005). In addition, Cluzel et al. (2005) report
172 granitoid dikes intruding the Massif du Sud at 27 and 24 Ma. They interpret these
173 dikes, coeval with the final obduction stages, as related to the inception of a new
174 subduction along the west coast of New Caledonia in response to the blocking of the
175 previous Eocene subduction. Altogether these dikes appear to record several
176 subduction-related events, suggesting that the ophiolite witnesses several and
177 distinctive stages of migration, accumulation and re-equilibration of different melts
178 (including hydrous melts) in a supra-subduction environment (e.g. Marchesi et al.,
179 2009; Ulrich et al., 2010; Pirard, 2012), rather than being the result of a fractional
180 crystallization from a single magma-type during oceanic spreading (Prinzhofer and
181 Allègre, 1985).

182 4 — an amphibolite to granulite facies metamorphic sole is found in places
183 between the ophiolite and the Poya and Pouebo underlying terranes (Cluzel et al.,
184 2012). This metamorphic sole is interpreted to have been welded to the ophiolite
185 during subduction inception. Using various thermobarometers, Cluzel et al. (2012)
186 estimate peak crystallization conditions around 5 kbar and 800 to 950 °C.
187 Thermochronology on hornblende ($^{40}\text{Ar}/^{39}\text{Ar}$), zircon and sphene (U-Pb) indicate that
188 these mafic rocks crystallized at ca. 56 Ma (Cluzel et al., 2012). This result and
189 inferred pressure–temperature conditions for the New Caledonia metamorphic sole
190 give a geothermal gradient greater than 40 °C/km, consistent with subduction
191 inception occurring at or near the spreading ridge of the South Loyalty Basin.

192

193 **3. Outcrop-scale deformation patterns and mineralogy**

194 The study area is exposed at low tide on the foreshore of Plum Beach (south-
195 east of Noumea), across a ~150x100 m surface (Fig. 1b). This area is bounded to the
196 west by the New Caledonia Basin and by a densely vegetated cliff to the east. It is
197 located near the base of the ophiolite, 50 to 100 m above the north-east dipping
198 contact with the underlying Poya unit, which outcrops 2 km to the north.

199 The study area is a major deformation zone characterized by large-scale mafic
200 shear bands (up to 100 m long in places) and centimetric to decimetric
201 orthopyroxenite veins away from the deformed area. The shear bands and veins cross-
202 cut the peridotite, which is constituted mainly of highly serpentized harzburgite and
203 rare dunite. One to two meters away from the shear bands, the peridotite is massive
204 and has only a tectonite porphyroclastic texture with a high temperature foliation
205 steeply dipping to the NE (parallel to the paleo-Moho according to Prinzhofer et al.,
206 1980). The orientation of this foliation is discordant with respect to the deformation
207 observed in the mafic shear bands.

208 Ten major shear bands have been recognized in the study area. Shear bands #1
209 to #7 are 30 to 50 m long and strike ~N150 (Fig. 1b). They disappear under the
210 eastern cliff but likely extend further to the south-east, where similar shear bands are
211 found in apparent continuity along the coast (#8 to #10). The overall length of these
212 shear bands is thus greater than 150 m.

213 The shear bands are m- to several m-thick and commonly outlined by 10–50 cm
214 thick amphibolite bands, mostly found at the core of the shear bands (and, when
215 found on the margins, only on their south-west side; Fig. 2).

216 Peridotite and amphibolite are strongly boudinaged within the amphibolite-
217 bearing shear zones (Fig. 3a,b,c). Strain increases towards the core of the shear bands,
218 as shown by the marked decrease in boudin size, from the m- to the cm- scale. Both
219 ultramafic and mafic boudins are sigmoidal and indicate consistent sinistral-reverse,
220 top to the north-west shear senses across the whole outcrop (Fig. 2,3a). These boudins
221 are embedded in a phlogopite-rich matrix described below.

222 A network of distributed leucocratic feldspar-rich veins also cross-cuts the
223 peridotite. This network is mainly found on the margins of the high strain bands.
224 Mutual cross-cutting relationships between the dark amphibolites and the feldspar-
225 rich veins suggest that they formed coevally. The vein network is developed only on
226 one side of the amphibolite bands (to the NE); the density, thickness and length of the
227 veins decrease away from the amphibolites (Fig. 3c,d,f,g). The mafic shear band #4bis
228 is a dm-thick dark amphibolite without feldspar-rich veins (Fig. 2).

229 Amphibolite bands (Fig. 3e) chiefly comprise calcic amphibole with 10 to 50
230 vol.% plagioclase, variably altered to reddish-to-yellowish low-grade oxidized
231 aggregates. Pyroxene (always orthopyroxene, as detailed below) amounts to 15–30
232 vol.%.

233 Leucocratic feldspar-rich veins are characterized by a predominance of
234 plagioclase with up to 20 vol.% coarse-grained orthopyroxene crystals (Fig. 4a,b).
235 Most orthopyroxenes are rimmed by calcic-amphibole.

236 A light colored, green to grey matrix comprising phlogopite, chlorite, talc and
237 serpentine is observed within the shear bands and appears to form at the expense of
238 both the mafic and ultramafic rocks (Fig. 3c,d,e,f). Late, milky white cm-thick
239 serpentine veinlets rim peridotite boudins and also cross-cut the boudins and the

240 matrix. Phlogopite is also observed concentrated in m-scale veins next to some of the
241 shear bands (#9).

242

243 **4. Sampling strategy and analytical techniques**

244 *4.1 Sample selection*

245 60 samples (Table 1) were collected in the field across the whole range of
246 textures and mineralogical assemblages of the shear bands (both for mafic -M- and
247 ultramafic -UM- protoliths; Fig. 2). 15 samples are dark amphibolites, either within
248 deformation bands (e.g. M4d; Fig. 2,5a) or isolated in undeformed peridotite (e.g.
249 M8a), whereas 5 samples are deformed feldspar-rich mafic veins (e.g. M8e; Fig. 5b),
250 including an undeformed feldspar-rich veinlet isolated in peridotite (sample M9).
251 Ultramafic samples comprise 17 peridotite boudin samples surrounded by deformed
252 feldspar-rich veins and phlogopite-rich matrix (e.g. UM13, UM10; Fig. 5c,d), 14
253 undeformed peridotite samples (e.g. UM12, UM14) and one orthopyroxenite vein
254 sample.

255

256 *4.2 Electron probe microanalysis (EPMA)*

257 EPMA was carried out at CAMPARIS (UPMC-IPGP, Paris, France) on
258 amphibolites (n=7), feldspar-rich veins (n=5), peridotite boudins (n=5), undeformed
259 peridotites (n=2) and on the biotite-rich matrix (n=7), using classical analytical
260 conditions for spot analyses (1-2 μm spot size; 15kV, 10nA, wavelength-dispersive
261 spectroscopy mode), using Fe_2O_3 (Fe), MnTiO_3 (Mn, Ti), diopside (Mg, Si), CaF_2 (F),

262 orthoclase (Al, K), anorthite (Ca) and albite (Na) as standards for bracketed elements.
263 Representative analyses are given in Table 2.

264

265 *4.3 Thermobarometry*

266 Pressure and temperature conditions of crystallization have been estimated
267 using the hornblende-plagioclase geothermobarometer of Holland and Blundy (1994)
268 modified by Anderson and Smith (1995) in three olivine-bearing amphibolites and in
269 six olivine-free amphibolites. This geothermobarometer is based on the reaction:
270 edenite + albite = richterite + anorthite. The absence of Ti-rich phases in amphibolites
271 invalidates the use of Ti-based geothermobarometers. Representative pressure–
272 temperature estimates are given in Table 2.

273 The Fe^{3+} estimation method in amphibole given in Holland and Blundy (1994)
274 provides the minimum ($\text{Fe}^{3+}/\text{Fe}_{\text{tot}}$) ratio for which sufficient Na is allocated to the A-
275 site. $\text{Fe}^{3+}/\text{Fe}_{\text{tot}}$ here averages around 0.4 in late amphibole (i.e. magnesio-hornblende).
276 With a ($\text{Fe}^{3+}/\text{Fe}_{\text{tot}}$) ratio = 1, estimated after the method of Leake et al. (1997), the
277 estimated temperature is ~ 70 °C higher than that calculated with $\text{Fe}^{3+}/\text{Fe}_{\text{tot}} = 0.4$.

278

279 *4.4 ICPMS and LA-ICPMS:*

280 Whole-rock major element analysis was performed on 7 samples (Table 3) at
281 EOST in Strasbourg, France, by ICP-AES and ICP-MS (see Table 2; analytical
282 procedure given in Omrani et al., 2008) and at UPMC Univ. Paris 6 (following the
283 procedure of Salaün et al. 2010).

284 LA-ICPMS was performed on amphibole in at IRD Bondy using a 7500 cx
285 AGILENT NI Standards coupled with a laser NWR-213 nm. Analyses were carried

286 out using a 10 Hz laser repetition rate and 50 μm beam diameter for amphibole. All
287 analyses were conducted at fixed beam position. For internal standardization ^{43}Ca was
288 used for amphibole. For external standardization, NIST SRM 610 and 612 glasses
289 were used after Pearce et al. (1997). Raw data were processed using GLITTER
290 (GEMOC, Macquarie University, Australia). The following masses were analyzed
291 (Table 4): ^7Li , ^{11}B , ^{47}Ti , ^{85}Rb , ^{88}Sr , ^{89}Y , ^{90}Zr , ^{93}Nb , ^{133}Cs , ^{138}Ba , ^{139}La , ^{140}Ce , ^{141}Pr ,
292 ^{146}Nd , ^{147}Sm , ^{153}Eu , ^{157}Gd , ^{159}Tb , ^{169}Dy , ^{165}Ho , ^{166}Er , ^{172}Yb , ^{175}Lu , ^{178}Hf , ^{181}Ta , ^{208}Pb ,
293 ^{238}U , ^{232}Th .

294

295 *4.5 Radiometric dating*

296 The selected samples were crushed and sieved; single grains of amphibole were
297 handpicked under binocular microscope and cleaned in ultrasonic bath with acetone
298 and distilled water. They were packaged in Al foils and irradiated for 40 hours in the
299 core of the Triga Mark II nuclear reactor of Pavia (Italia) with several aliquots of the
300 Fish Canyon sanidine standard (28.03 ± 0.08 Ma; Jourdan and Renne, 2007) as flux
301 monitor. Argon isotopic interferences on K and Ca were determined by irradiation of
302 KF and CaF_2 , pure salts from which the following correction factors were obtained:
303 $(^{40}\text{Ar}/^{39}\text{Ar})_{\text{K}} = 0.00969 \pm 0.00038$, $(^{38}\text{Ar}/^{39}\text{Ar})_{\text{K}} = 0.01297 \pm 0.00045$, $(^{39}\text{Ar}/^{37}\text{Ar})_{\text{Ca}} =$
304 0.0007474 ± 0.000021 and $(^{36}\text{Ar}/^{37}\text{Ar})_{\text{Ca}} = 0.000288 \pm 0.000016$. Argon analyses
305 were performed at Géosciences Montpellier (France) with two analytical devices that
306 each consist of: (a) an IR-CO₂ laser of 100 kHz used at 5-15% during 60 sec, (b) a
307 lenses system for beam focusing, (c) a steel chamber, kept at 10^{-8} - 10^{-9} bar, with a
308 drilled copper plate, (d) an inlet line for purification of gases including two Zr-Al
309 getters, (e) a multi-collector mass spectrometer (Argus VI from Thermo-Fisher) or an

310 MAP215-50 single collector mass spectrometer depending on the used device. A
311 custom-made software controls the laser intensity, the timing of
312 extraction/purification and the data acquisition. To measure the Ar background within
313 the system, one blank analysis was performed every three sample analyses.
314 ArArCalc© v2.5.2 was used for data reduction and plotting. The one-sigma errors
315 reported on plateau, isochron and total gas ages include the error on the irradiation
316 factor J. Atmospheric ^{40}Ar was estimated using a value of the initial $^{40}\text{Ar}/^{36}\text{Ar}$ of
317 295.5. Data are portrayed as age spectra on figure 11 and summarized in Table 5.

318

319 **5. Petrography**

320 **5.1. Mafic rocks**

321 Dark mafic bands (Fig. 6a,b,c) contain abundant amphibole (40–70 vol.%),
322 plagioclase (10–50 vol.%), orthopyroxene (10–20 vol.%) ± olivine (0–25 vol.%).
323 Olivine is only found in some mafic layers (i.e. shear band #4bis; M8a, Fig. 6c). No
324 clinopyroxene is found. Sulphide inclusions in amphibole, iron oxides and rare apatite
325 (samples M8a and M8b) are also present.

326 Amphibole surrounds all other mineral phases and appears texturally as a late
327 crystallization product with respect to olivine, orthopyroxene (Fig. 6d,e,g, 7a,b,c,d).
328 Sample M8a show two distinct amphibole-bearing layers as an olivine-free
329 amphibolite vein cross-cut a boudinaged olivine-bearing amphibolite band (Fig 6c). In
330 olivine-bearing amphibolites (shear band #4bis), amphibole (referred to as early
331 amphibole) is strongly pleochroic from blue to green under the microscope with a
332 sub-automorph habitus and marked cleavages (Fig. 6d,g). By contrast, amphibole in

333 olivine-free amphibolite (referred to as late amphibole) shows a distinctly paler
334 pleochroism and presents a higher degree of deformation (Fig. 6e).

335 Textural relationships between plagioclase, orthopyroxene and olivine are
336 obscured by the amphibolitization stage. Olivine appears as fine-grained clusters
337 closely associated with, and generally included in, orthopyroxene (Fig. 6g, 7a,b). In
338 sample M4e, olivine was replaced by orthopyroxene before the crystallization of large
339 amphibole grains (Fig. 7c,d). Olivine is also partially destabilized by tremolite.
340 Primary Cr-rich spinel crystals are present in olivine (Fig. 7a). In olivine-bearing
341 samples, plagioclase is rare and partially to totally replaced by amphibole (Fig. 6g).

342 Leucocratic feldspar-rich veins are mainly composed of plagioclase (70–100%)
343 with subordinate orthopyroxene (0–20%) and rare amphibole (0–10%). Undeformed
344 leucocratic veins show euhedral amphibole in a groundmass of large plagioclase
345 crystals (Fig. 6f). Close to the shear bands, veins of plagioclase evidence deformation
346 and grain size reduction. Orthopyroxene is only partially preserved yet retaining
347 textural evidence of former equilibrium with plagioclase. Orthopyroxene is included
348 in amphibole or rimmed by amphibole pressure shadows (Fig. 6e), again suggesting
349 the partial destabilization of coarse grained primary assemblage of orthopyroxene (\pm
350 olivine) and plagioclase to form various generations of amphibole. A network of late
351 serpentine, chlorite and talc is observed in orthopyroxene fractures (Fig. 6d).

352 As for ultramafic rocks, mafic bands are locally replaced by a phlogopite-rich
353 matrix (in which phlogopite is in turn replaced by chlorite). Orthopyroxene and
354 calcic-amphibole appear only as relicts into layers of phlogopite and acicular ortho-
355 amphibole. Plagioclase is destabilized and rimmed by a Ba-rich K-feldspar.

356

357 5.2. Ultramafic rocks (UM) and matrix

358 The mineralogy of undeformed peridotites is dominated by an isotropic network
359 of serpentine (75-90%), with rare relicts of brown orthopyroxene, olivine phenocrysts
360 and black chromiferous spinels. These peridotites belong to the highly depleted
361 harzburgites of the Massif du Sud (i.e. Prinzhofer and Allègre, 1985; Ulrich et al.,
362 2010; Pirard et al., 2013). Few deformation textures are observed, where the extensive
363 late serpentinization may have erased former deformation textures.

364 Peridotites boudinaged in shear bands (Fig. 7b) show a similar mineral
365 assemblage. However, a reaction rim around peridotite boudins contains coarse-
366 grained hornblende pseudomorphing orthopyroxene, and containing chromite
367 inclusions (Fig. 7g). Light colored phlogopite (Fig. 7b,f,h) and acicular ferro-
368 magnesian ortho-amphibole (Fig. 5f) co-crystallized around peridotite boudins as a
369 result of the partial destabilization of ultramafic orthopyroxene and secondary
370 hornblende. Therefore, this assemblage postdates the boudinage of the peridotite and
371 the first hydration recorded by the crystallization of hornblende. Phlogopite increases
372 in modal proportion towards the amphibole-rich mafic bands, along with a grain size
373 decrease but shows no preferred orientation. Chlorite is interlayered with phlogopite
374 and crystallizes as a late stage alteration product responsible for the macroscopically
375 greenish to greyish color of the matrix. Millimetric veinlets of talc–chlorite–
376 serpentine, slightly sheared in places, also cross-cut peridotites boudins (Fig. 7h).
377 These veinlets are themselves cross-cut by a late isotropic serpentine network.

378

379 **6. Mineralogy**

380 *Orthopyroxene*

381 Orthopyroxene from harzburgites displays a typical Mg# [Mg/(Mg+Fe)] of 0.92
382 \pm 0.02 (Fig. 8a). In mafic rocks, the Mg# of orthopyroxene shows lower values
383 ranging between 0.77 and 0.85.

384 Cr₂O₃ and CaO contents (Fig. 7a) show small yet systematic variations between
385 the mafic bodies, with higher values in olivine-bearing rocks (< 0.20 wt% Cr₂O₃;
386 0.35–0.65 wt% CaO, except for three points with 0.82 and 0.97 wt% CaO) than in
387 leucocratic feldspar-rich rocks (< 0.04 wt% Cr₂O₃; 0.15–0.35 wt% CaO). Cr₂O₃ and
388 CaO contents are higher in ultramafic rocks, with large intra-sample variations (0.40-
389 0.80 wt% Cr₂O₃; 0.40-1.80 wt% CaO). Al contents are low (~0.02 p.f.u.) and show no
390 systematic variation.

391 Orthopyroxene in harzburgite falls within the composition domain of
392 orthopyroxene-poor harzburgite described by Prinzhofer et al. (1980) and Pirard et al.
393 (2013; Fig. 8a). Orthopyroxene in olivine-bearing and olivine free amphibolites
394 overlaps the range of orthopyroxene in amphibolite and hornblendite dikes described
395 by Pirard (2012). Orthopyroxene in mafic bands have similar Mg# orthopyroxene
396 from gabbro-norite sills of Pirard et al. (2013), but show lower Ca content.

397

398 *Olivine*

399 The patterns of Mg# of olivine (Fig. 8b) in harzburgite and in olivine-bearing
400 mafic rocks mirror those of orthopyroxene described above (Fig. 8b). Mg# in
401 ultramafic olivine is 0.91–0.92 and the MnO content varies between 0.06 and 0.22
402 wt% (Fig. 8b), in the same range as those of nearby harzburgites (Pirard et al., 2013).
403 Olivine in mafic bands shows similar Mg# (0.78-0.85 wt%) to amphibolite and
404 hornblendite dikes of Pirard (2012). The Mg# is lower and the MnO content (0.20–

405 0.39 wt%) is higher than reported for mafic cumulates by Pirard et al. (2013; Fig. 8b).
406 NiO content in mafic olivine varies between 0.16 and 0.21 wt%, which is consistent
407 with the most evolved olivine found in olivine-bearing cumulates (Pirard et al., 2013).
408 Compositional trends in NiO and MnO vs. Mg# match the fractional crystallization
409 trends evidenced by Pirard et al. (2013).

410

411 *Feldspar*

412 Plagioclase occurs only in mafic rocks, as veins in leucocratic feldspar-rich
413 rocks (with a composition of a norite), or as scarce, isolated crystals in olivine-bearing
414 rocks. In noritic rocks, albite content [$\text{Na\#} = \text{Na}/(\text{Na}+\text{Ca})$] ranges from 0.46 to 0.66
415 (from labradorite to andesine; Fig. 8c), in the domain of amphibolite dikes described
416 by Pirard (2012). In olivine-bearing rocks, plagioclase corresponds to anorthite (Na\# :
417 0.02–0.07; Fig. 8c), in the range of amphibolite dikes of Pirard (2012). Secondary
418 feldspar rimming norite veins are rich in alkalis (~5.5 wt% Na_2O ; ~3.5 wt % K_2O ;
419 ~12.5 wt% BaO) and have intermediate composition between anorthoclase and
420 hyalophane.

421

422 *Ca-amphibole*

423 Amphibole shows Mg# comparable to that of orthopyroxene and olivine in both
424 mafic (0.90–0.74; Fig. 9a) and ultramafic rocks ($\sim 0.91 \pm 0.01$; Fig. 9a), and $\text{Na} < 0.5$
425 p.f.u. and $(\text{Ca}+\text{Na})_{\text{B}} > 1.0$ p.f.u. typical of calcic-amphibole (Leake et al., 1997).

426 In olivine-bearing mafic bodies, early amphibole is pargasitic with $(\text{Na}+\text{K})_{\text{A}} \geq$
427 0.5 p.f.u and Si ranging from 6.12 to 6.30 p.f.u. (Fig. 9b). Late amphibole
428 destabilizing this pargasite hornblende has $(\text{Na}+\text{K})_{\text{A}} \leq 0.5$ p.f.u. and a higher Si

429 content (6.50–7.00 p.f.u.; Fig. 9b), and belongs to the magnesio-tschermakite
430 hornblende (Leake et al., 1997). In olivine-free samples (i.e M3c), late amphibole
431 shows a decrease in $(\text{Na}+\text{K})_{\text{A}}$ and an increase of the Si content (6.37–6.88 p.f.u.; Fig.
432 9b). This evolution is also observed with the content in Ti (0.15–0.05 p.f.u.). This
433 composition is similar to the composition of amphibole from undeformed mafic dikes
434 described by Pirard (2012), and partially fall in the domain of the metamorphic sole
435 amphibolites from Thio considering their Ti and $(\text{Na}+\text{K})_{\text{A}}$ contents, yet with a much
436 lower Mg# (see Cluzel et al., 2012).

437 Early amphibole in olivine-bearing amphibolite has the lowest Ti (0.01 and 0.05
438 p.f.u) and does not fall in the domains of amphibole-bearing dikes (Pirard, 2012) or of
439 the metamorphic sole (Cluzel et al., 2012) (Fig. 9c). Amphibole with the highest Si
440 content is again the poorest in $(\text{Na}+\text{K})_{\text{A}}$ content (Fig. 9b). Tremolite is only present in
441 olivine-bearing rocks and texturally associated to mafic olivine (Fig. 9b).

442 Calcic-amphibole destabilizing ultramafic orthopyroxene are magnesio-
443 hornblende with high Si content (7.19–7.46 p.f.u.), $(\text{Na}+\text{K})_{\text{A}}$ between 0.25 to 0.12
444 p.f.u. (Fig. 9b) and $\text{Ti} < 0.10$ p.f.u. (Fig. 9c).

445 Late acicular amphibole crystallizing within the phlogopite-rich matrix belongs
446 to the ferro-magnesian anthophyllite series ($\text{Si} \sim 8$ p.f.u; $\text{Al}_{\text{tot}} \leq 0.23$ p.f.u.; $(\text{Ca}+\text{Na})_{\text{B}}$
447 ≤ 0.15 p.f.u.; Leake et al., 1997).

448

449 *Phlogopite*

450 Phlogopite presents a Mg# of 0.89, regardless of its structural position in the
451 shear band (Table 1). Only few crystals display the K_2O content of a regular mica

452 owing to retrogression by chlorite. Cr_2O_3 (< 0.02 wt%) and TiO_2 (0.05 wt%) are
453 present in small amounts.

454

455 *Accessory minerals*

456 Early amphibole grains (i.e. pargasite) host various inclusions (e.g. spinel,
457 magnetite, sulfides). Magnetite inclusions derive from the oxidation of iron-sulphide
458 and spinel. Spinel in harzburgite is a chromite with a Cr# [$\text{Cr}\# = (\text{Cr}/(\text{Cr} + \text{Al} + \text{Fe}^{3+}))$] of
459 0.72 ± 0.05 and a Mg# of 0.35 ± 0.05 (Fig. 9d). Spinel is consistently more aluminous
460 in mafic rocks, with a Cr# from 0.18 to 0.38 for a Mg# varying between 0.25 and 0.32
461 (Fig. 9d). TiO_2 in mafic and ultramafic spinel is below the detection limit. Cu- and
462 Ni-rich Fe-sulphides (i.e. pentlandite-pyrrhotite) are found as inclusions in early
463 amphibole in olivine-bearing mafic rocks.

464

465 **7. Geochemistry**

466 **7a. Whole rock geochemistry**

467 Rare earth elements (REE) normalized to N-MORB of both olivine-bearing and
468 olivine-free amphibolites display values lower than 1 [$(\text{REE})_N \sim 0.4$], with the
469 exception of the sample M4f [$(\text{REE})_N \sim 1$] (Fig. 10a). They display a slight
470 enrichment in light REE (LREE) relative to heavy REE (HREE) [$(\text{La}/\text{Lu})_N = 1.89 \pm$
471 0.35]. High field strength elements (HFSE; Ti, Hf, Zr, Nb, Ta) are slightly depleted
472 relative to REE [$(\text{Nb}/\text{Lu})_N = 0.41 \pm 0.05$] (Fig. 10a). Large Ion Lithophile Elements
473 (LILE; Pb, U, Th, Ba, Rb, Cs) show large enrichments relative to REE, increasing
474 with their degree of mobility and incompatibility [$(\text{Cs}/\text{Lu})_N = 62 \pm 39$] (Fig. 10a).

475 The compositions range of metamorphic sole (Cluzel et al., 2012) display a
476 similar pattern to amphibolites though slightly shifted towards higher values by a
477 factor of ~2 (Fig. 10a). Gabbro-norite sills (Pirard et al., 2013) are highly depleted in
478 all incompatible elements. However, they display a pattern roughly similar to that of
479 amphibolites, with a mean relative depletion factor of ~10 but with larger Pb and Sr
480 relative anomalies (Fig. 10a). Arc tholeiitic dikes (Cluzel et al., 2006) have
481 incompatible elements compositions overlapping with those amphibolites (Fig. 10a).
482 It strongly suggests that tholeiitic dikes and amphibolites have a same origin.

483 In phlogopite-rich rocks, concentrations in HREE are similar to those of olivine-
484 free and olivine-bearing amphibolite (Fig. 10a). However, phlogopite-rich rocks have
485 a larger enrichment in LREE relative to HREE [(La/Lu)_N = 3.55 ± 1.77]. These rocks
486 are depleted in HFSE relative to REE but less depleted than olivine-bearing and
487 olivine-free amphibolite (with the exception of the sample M4f). Sr shows a large
488 depletion relative to REE and LILE display a large enrichment especially in Cs.

489

490 **7b. In-situ trace elements**

491 Incompatible trace elements compositions of early and late amphibole,
492 respectively found in olivine-bearing and olivine-free amphibolites display similar N-
493 MORB normalized patterns (Fig. 10b) with however slightly higher content in late
494 amphibole (i.e. magnesio-hornblende) compared to early amphibole (i.e. pargasite).
495 REE patterns from heavy REE (HREE) to light REE (LREE) are flat and close to N-
496 MORB values, with (La/Lu)_N = 0.87 ± 0.39. HFSE are slightly depleted relative to
497 REE [(Nb/Lu)_N = 0.21 ± 0.07]. LILE show positive anomalies [(Cs/Lu)_N = 12.1 ±
498 5.9]. Overall, similar behaviors (i.e. negative anomalies in HFSE and positive

499 anomalies in LILE) are observed in whole-rock data on amphibolites and in-situ data
500 on amphibole, suggesting that the trace element budget in the rocks is mainly
501 controlled by amphibole. However, higher HREE content in late amphibole compared
502 to that in early amphibole is difficult to explain with a fluid mediated
503 recrystallization. This higher content more likely mirrors the more differentiated
504 nature of the norite dikes compared to the troctolite dikes (prior to metamorphic
505 hydration).

506 Amphibole in mafic amphibole-bearing dikes analyzed by Pirard (2012)
507 displays a pattern similar to that of the amphibole reported here (grey overlay; Fig.
508 10b): it shows a similar enrichment in Pb and depletions in Ti–Sr–Nb–Ta with
509 comparable absolute values, including in the most mobile and incompatible trace
510 elements, advocating for similar origins. Ba-content is similar in both analyses series
511 and Rb-content is lower in the amphibole of Pirard (2012) than all our early and late
512 amphiboles (Fig. 10b). Cs has not been measured.

513

514 **8. $^{40}\text{Ar}/^{39}\text{Ar}$ constraints on mafic rocks formation**

515 $^{40}\text{Ar}/^{39}\text{Ar}$ laser probe step-heating dating was carried out on 20 isolated single
516 amphibole crystals from 11 different rocks, using the procedure described above.
517 These samples are mainly olivine-free and olivine-bearing amphibolites from the
518 Plum beach. They also include amphibolites from the Thio metamorphic sole (sample
519 C10) and undeformed gabbroic dikes from the Massif du Sud (samples SNC03 and
520 SNC04, col des Deux Tétions). Age spectra and inverse isochron plots are reported on
521 figure 11 for some of these samples. The closure temperature for the $^{40}\text{Ar}/^{39}\text{Ar}$
522 method in hornblende is around 550 ± 50 °C (Harrison, 1981). For most samples,

523 errors are relatively large due to the fact that only one grain was analyzed each time
524 and that the K content (and therefore the amount of radiogenic Ar) is low in these
525 amphiboles.

526 Amphibole from the Thio metamorphic sole provides an age of 56.9 ± 2.6 Ma
527 (Fig. 11a,12) consistent with the age reported by Cluzel et al. (2012) from a nearby
528 exposure. Similar ages are obtained on the two undeformed gabbros from the Massif
529 du Sud for which intercept ages are in the range $51\text{--}52 \pm 4$ Ma (Fig. 11a,12; see the
530 Online Fig. S1 for isotope correlation plot), with initial $^{40}\text{Ar}/^{36}\text{Ar}$ values close to the
531 present-day atmospheric ratio (see Table 5). Large amphibole from a partially
532 weathered pegmatite dike yields more scattered results with some evidence of excess
533 argon as a result of fluid infiltration and only a maximum age of 60 ± 5 Ma can be
534 attributed to this sample.

535 At Plum beach, 8 hornblende crystals from 4 olivine-free amphibolites (M3d,
536 M4d, M8a-Z2, M8f) and 5 hornblende crystals from two weakly deformed olivine-
537 bearing amphibolite (M15 and M8a-Z1) yield ages ranging between 54 and 93 Ma,
538 with the exception of one crystal (M15) that displays a strongly discordant age pattern
539 and an integrated age of 167 ± 21 Ma likely related to trapped excess argon. For M8f,
540 4 amphiboles were analyzed and representative age spectra are shown on figure 11b,
541 with plateau dates ranging from 54.9 ± 2.0 Ma to 59.2 ± 2.7 Ma and a similar age in
542 the isotope correlation plot with an initial atmospheric ratio (reported in the Online
543 Fig. S1). The consistency of the results on 4 grains from M8f and the lack of evidence
544 of excess argon for this sample suggest that amphibolitisation and related deformation
545 occurred at ca. 55 Ma, which is supported by the age of 54.1 ± 3.4 Ma obtained on
546 amphibolite M3b from another shear zone. According to microprobe analyses and
547 textural observations, it is likely that this age of ca. 55 Ma corresponds to the

548 crystallization of secondary magnesio-hornblende that partially replaced pargasite and
549 orthopyroxene. This age range coincides with the formation of the metamorphic sole
550 at Thio (sample C10 and Cluzel et al., 2012), the emplacement of some gabbros
551 (samples SNC03 and SNC04) and pre-obduction felsic dikes (53 Ma; Cluzel et al.,
552 2006).

553 8 amphiboles from 5 olivine-bearing samples yield older ages ranging from 73
554 to 93 Ma (Table 5) with relatively discordant age spectra (Fig. 11c) and large
555 experimental errors due to the low amount of K. However, the isotope correlation
556 diagram and the flat K/Ca ratios do not provide evidence of excess argon (reported in
557 the Online Fig. S1), which could indicate that ages around 90 Ma are meaningful.
558 This suggests that early amphibole is preserved in the dated rocks, which is consistent
559 with the chemical data that point to the coexistence of primary pargasite predating
560 synkinematic magnesio-hornblende developed around it.

561 **9. Interpretations and discussion**

562 **9.1. Nature of the protolith and subduction signature**

563 The sharp contact between mafic rocks and surrounding peridotites (preserved
564 from metasomatic alteration in places) suggests that these mafic rocks were emplaced,
565 prior to amphibolitization, as 100 m-scale intrusions, most likely as dikes or melt
566 lenses. The lack of pervasive deformation in the peridotite a few meters away from
567 the shear bands, and the presence of olivine and orthopyroxene in mafic rocks, rule
568 out the possibility that the mafic bands represent sheared metamorphic sole
569 amphibolites imbricated within mantle wedge peridotites. The protolith of the mafic
570 bands is therefore interpreted as emplaced in the form of small-scale melt intrusions at
571 temperatures where peridotites were comparatively brittle.

572 Partially preserved orthopyroxene, anorthitic plagioclase and high Mg# olivine
573 indicate that the protolith of the amphibolite bands was a norite or a troctolitic norite.
574 Late hornblende (i.e. magnesio-hornblende) and Ca–Na plagioclase, which abound in
575 amphibolite bands, may have recrystallized from orthopyroxene, olivine and
576 anorthitic plagioclase (such as found in the “noritic” pods) during amphibolitisation
577 and/or later fluid infiltration. Clinopyroxene was either absent or is now completely
578 replaced (undeformed hornblendite dikes from Rivière des Pirogues show late
579 amphibole replacing large clinopyroxene crystals; Pirard, 2012).

580 Olivine-bearing domains cross-cut by olivine-free domains (Fig. 6c) point to
581 evolving compositions of the magma intruding the peridotite. The positive correlation
582 between Cr# of spinel and Mg# of olivine in olivine-bearing amphibolites suggests
583 that the mafic protolith formed through melting of harzburgite with a supra-
584 subduction signature followed by fractional crystallization (Marchesi et al., 2009).
585 The presence of orthopyroxene, plagioclase and early amphibole (i.e. pargasite) is
586 consistent with peridotite melting triggered by aqueous slab-derived fluids (Pirard,
587 2012).

588 Amphibolites display flat N-MORB normalized patterns with lower REE
589 abundances than N-MORB ($REE_{\text{Amphibolite}} / REE_{\text{N-MORB}} \sim 0.5$; Fig. 10). A simple
590 melting model shows that these amphibolites are too enriched, however, to derive
591 directly from melting of the host depleted harzburgite and require a less depleted
592 ultramafic source (either refertilized by fluids or silica-rich melts). Trace elements
593 patterns of amphibolites in fact closely fit those of unmetamorphosed microdioritic
594 dikes found in the Massif du Sud, interpreted as subduction-related magma intrusions
595 by Cluzel et al. (2006).

596 The slab signature is best recorded by HFSE and LILE. HFSE in the bulk rock
597 display negative anomalies relative to REE (Fig. 10). These are characteristic of
598 subduction zone magmas, as these elements are retained in rutile and zircon in the
599 subducting slab (e.g. Coltorti et al., 2007; Zack and John, 2007). The enrichment in
600 LILE (and to a lesser extent in LREE) is also consistent with a subduction signature,
601 since LILE are mobilized during the dehydration of the upper part of the slab (e.g.
602 Pearce et al. 1984). Such enrichments are observed in metamorphic soles for Rb, Ba
603 and also Pb (Ishikawa et al., 2005). These enrichments cannot result from late sub-
604 surface alteration, since LILE and HFSE are mainly concentrated in high temperature
605 amphiboles which impose their geochemical signature in the whole rock (Fig. 10b;
606 Table 4).

607

608 **9.2 Petrological evolution, deformation and fluid ingress**

609 Field relationships and mineralogy suggest that the dark amphibolite bands and
610 the more leucocratic noritic veins are cogenetic, with percolation apparently favored
611 towards the north. This direction is compatible with upward melt migration expected
612 at plate boundary (e.g. Matthews et al., 2015) within the mantle wedge (Fig. 2, 13a)
613 and close to the nascent crust (e.g. Pirard et al., 2013). Progressive aqueous fluid
614 infiltration and amphibolitization of the preexisting mafic bands would have
615 preferentially localized deformation in the former dikes during the first stages of the
616 Eocene subduction. Fig. 4c and Fig. 5c exemplify mafic veins with peridotite boudins.
617 Deformation localization is here closely tied with fluid infiltration, consistently with
618 the weaker rheology of amphibole and plagioclase compared to pyroxene (Homburg
619 et al., 2010) and olivine.

620 Three main hydration stages can be seen in these rocks and related to
621 deformation events, as schematized in Fig. 13:

622 — (a) *amphibolitization stage*:

623 Replacement of orthopyroxene by late amphibole is the second event of hydrous
624 fluid intrusion and the major deformation stage (Fig. 13b). Whereas magnesio-
625 hornblende in undeformed leucocratic veins has an euhedral shape and are associated
626 to a groundmass of large plagioclase crystals (Fig. 6f), magnesio-hornblende within
627 the shear bands makes up the foliation, together with plagioclase (Fig. 6e), and
628 presents syn-kinematic texture (e.g. pressure shadow around orthopyroxene; Fig. 6e).
629 Euhedral late amphibole could be of magmatic origin, but syn-kinematically
630 deformed late amphibole has likely crystallized from an aqueous fluid. Nonetheless,
631 evolving late amphibole (and possibly plagioclase) compositions (Fig. 8c, 9b,c)
632 indicate that mafic bands underwent progressive recrystallization during fluid
633 infiltration and deformation (localized within the shear bands).

634 Although mafic bands are strongly boudinaged and folded, amphibole crystals
635 themselves are only mildly deformed. This is consistent with plagioclase being
636 weaker than amphibole (Getsinger and Hirth, 2014) and with some static
637 recrystallization after the peak of deformation. Peridotites in contact with mafic dikes
638 are also boudinaged during this stage and record hydration through the crystallization
639 of amphibole with a Mg# around 0.92 (Fig. 9a).

640 Magnesio-hornblende (i.e. late amphibole) compositions yield pressure-
641 temperature conditions 3–5 kbar and 750 ± 50 °C (Table 2). These estimations are
642 similar to that calculated for leucodiorite dikes in the uppermost harzburgite (i.e. $3.5 \pm$

643 0.6 kbar and 800 ± 45 °C; Pirard, 2012). As for metamorphic sole, the pressure
644 estimates for the dikes cannot be explained by the present ophiolite thickness.

645 Age constraints on amphibole reveal an age cluster at 55 ± 2 Ma (Fig. 12),
646 which is well defined for late amphibole. Although the closure temperature for the
647 $^{40}\text{Ar}/^{39}\text{Ar}$ method in hornblende is ~ 200 °C lower than our temperature estimates for
648 amphibole formation, we interpret these ages as crystallization ages in a context of
649 fast cooling. Considering available constraints on the latest oceanic magmatic events,
650 pressure–temperature estimates and the age of the New Caledonia metamorphic sole
651 (Cluzel et al., 2012 and this study), high temperature metamorphism and cooling
652 below the closure temperature must have occurred in a very short lapse of time after
653 the Eocene subduction initiation.

654 — *(b) formation of the alkali-rich matrix:*

655 The alkali-rich matrix is composed of an association of Ba-rich feldspar,
656 phlogopite and ortho-amphibole crystallized at the expense of pyroxene, olivine and
657 Ca-amphibole, from both the mafic and ultramafic protoliths, with less deformation
658 than during the main amphibolitization stage (Fig. 13c). This matrix points to an
659 episode of fluid ingress postdating amphibolitization, with a markedly distinct
660 fluid composition (i.e. richer in K, Na and Ba). Ortho-amphibole (anthophyllite)
661 indicates temperatures of $\sim 700 \pm 50$ °C (Chernosky and Autio, 1979, Nozaka, 2011).

662 — *(c) Late reaction with serpentine, talc, chlorite:*

663 Finally, veinlets composed of hydrous phases (serpentine, talc and chlorite)
664 form mm to cm-scale rims around mafic and peridotite boudins or cross-cut the
665 matrix (Fig. 13d). With temperatures lower than 500–600 °C, according to presence

666 of the talc–chlorite assemblage, this stage reflects the latest and coldest episode of
667 reequilibration of these shear bands.

668

669 **9.3. Geodynamics implications**

670 Both petrology and geochemistry (through the similarity to undeformed
671 amphibole-bearing dikes reported in the Massif du Sud; Cluzel et al., 2006; Pirard,
672 2012) suggest that mafic shear bands correspond to former hm- to km-scale mafic
673 intrusions emplaced in the highly refractory host harzburgite, deriving from the
674 melting of a moderately depleted mantle source fluxed by subduction-derived fluids
675 (see § 9.1).

676 Trace element geochemistry (i.e. HFSE depletion and LILE enrichment) and
677 age constraints at ca. 55 Ma (Fig. 12) show that secondary amphibolitization
678 coincides with the east-dipping subduction initiation preluding to obduction (Spandler
679 et al., 1995; Cluzel et al., 2001, 2012). Fluid infiltration and at least part of the trace
680 element enrichment could result from dehydration of the upper part of the slab (as for
681 metamorphic sole dehydration; Ishikawa et al., 2005). In this interpretation, the shear
682 bands reported here are a direct record of early subduction processes in New
683 Caledonia, deformation and mantle metasomatism in an active subduction setting.

684 By contrast, the magmatic origin of the mafic protoliths remains uncertain. The
685 high degree of amphibolitization affecting these mafic dikes hampers modeling the
686 degree of partial melting of the peridotites and to constrain the exact origin of the
687 mafic protoliths.

688 Age constraints on amphiboles at both ca. 90 Ma (early pargasite) and ca. 55
689 Ma (magnesio-hornblende) can be accounted for in two different ways:

690 (1) amphibole-bearing troctolite to norite mafic intrusions would have formed at
691 ca. 55 Ma as a result of the east-dipping subduction preluding to obduction, as for the
692 relatively abundant dikes interpreted as supra-subduction and dated at ca. 55 Ma by
693 Cluzel et al. (2006) or dikes found in the nearby Massif du Sud exposures (also
694 interpreted as Eocene, based on structural data; Pirard, 2012), both interpreted as
695 formed in a supra-subduction setting at the expense of a metasomatized, HFSE-
696 depleted mantle. In this view, ages at ca. 90 Ma would be attributed to excess argon.

697 (2) mafic intrusions would represent inherited mafic melts from the Cretaceous
698 South Loyalty basin accretion (dating back to ≥ 90 Ma; e.g. Cluzel et al., 2012 and
699 this study), as several early amphiboles yield an age of ca. 90 Ma (Fig. 12). This
700 hypothesis is supported by the similarity with ages found in amphibole-bearing dikes
701 in the Massif du Sud by Prinzhofer (1981). Hydrous mafic melt intrusions would have
702 been emplaced possibly as a result of the west-dipping Pacific subduction, and later
703 been fluxed by fluids leading to late amphibole crystallization and ages at ca. 55 Ma
704 (during the second, east-dipping subduction initiation). This secondary high
705 temperature event has only partially reset the ca. 90 Ma age in pargasite.

706 The second interpretation is favored by the lack of evidence for excess argon,
707 early amphibole (pargasite) compositions that are distinct from those of Pirard (2012),
708 and the peculiar deformation of these dikes (those of Pirard and Cluzel are not or only
709 mildly deformed).

710 Deformation and fluid infiltration later localized on these preexisting mafic
711 intrusions. Shear bands developed as a result of a threefold stage of fluid infiltration
712 from slab-derived fluids. These mafic bands (and the later metasomatic matrix) are
713 good witnesses of the subduction initiation processes, both in terms of fluid influx and
714 progressive deformation near the base of the obducted New Caledonia ophiolite, from

715 temperatures of 800–700 °C down to less than 600 °C. Localization of deformation
716 along inherited structures such as mafic dike swarms also supports the view that
717 subduction initiation is generally triggered near a preexisting weakness zone (i.e.
718 Agard et al., 2014).

719

720 **10. Conclusions**

721 This study focuses on strongly sheared mafic amphibolites within the uppermost
722 mantle of the New Caledonia obducted ophiolite, near the transition zone and ~50–
723 100 m above the basal thrust contact. Using a combination of petrological,
724 geochemical, and geochronological data, we show that these amphibolites record
725 melting and polyphase metasomatism within the mantle wedge during the early
726 history of the subduction.

727 Shear bands indicate that deformation and high temperature fluid infiltration
728 (from ~ 800 to ~500 °C) localized along preexisting hm-scale mafic intrusions
729 emplaced within the upper mantle of the New Caledonia ophiolite and derived from
730 peridotites refertilized by slab-derived fluids in a supra-subduction zone environment.
731 The timing of these melt intrusions remains uncertain. Age constraints on early
732 amphibole at ca. 90 Ma and the lack of evidence for argon excess suggest that they
733 could be inherited from the former west-dipping Pacific subduction.

734 Secondary deformation and amphibolitization of these mafic dikes were the
735 result of a threefold infiltration of slab-derived fluids. The age cluster of amphibole at
736 ca. 55 Ma indicates that amphibolitization of the mafic protoliths (as well as later
737 alkali-metasomatism) relate to the onset of the Eocene east-dipping subduction
738 initiation preluding to the obduction of the New Caledonia ophiolite.

739 These outcrops bring important information on the processes and timing of fluid
740 influx, progressive deformation and peridotites refertilization during subduction
741 initiation, showing how m- to hm-scale heterogeneities (inherited or early-emplaced
742 magmatic dikes) localize deformation and metasomatism in the oceanic lithosphere.
743 In turn this suggests that similar small-scale heterogeneities may have controlled the
744 development of larger thrusts, ultimately leading to the tectonic slicing (and later
745 emplacement) of the New Caledonia ophiolite.

746

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757

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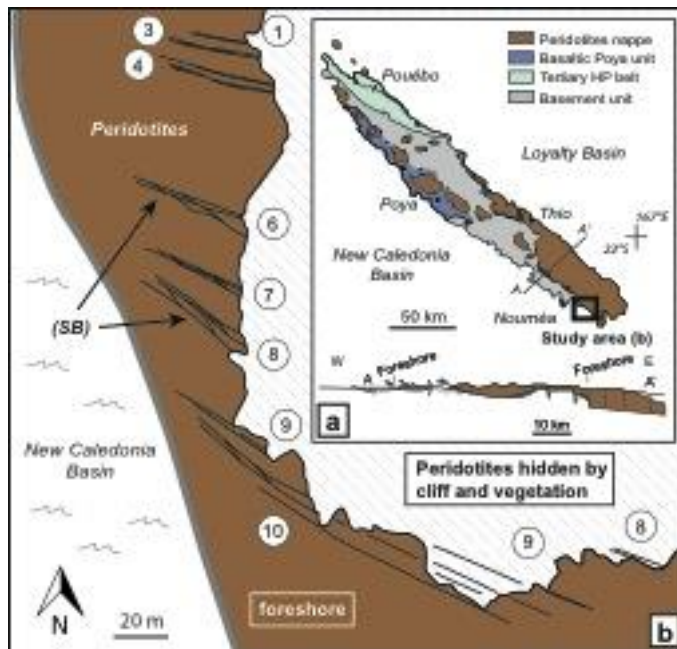
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- 948

949 Captions

950 Figure 1

951



952

953 a) Simplified geological map and cross-section of New Caledonia Island (after Cluzel

954 et al., 2001); b) Structural map of the study area near to Plum Beach (S.E. of

955 Nouméa) showing the distribution of m-scale shear mafic bands within the peridotite

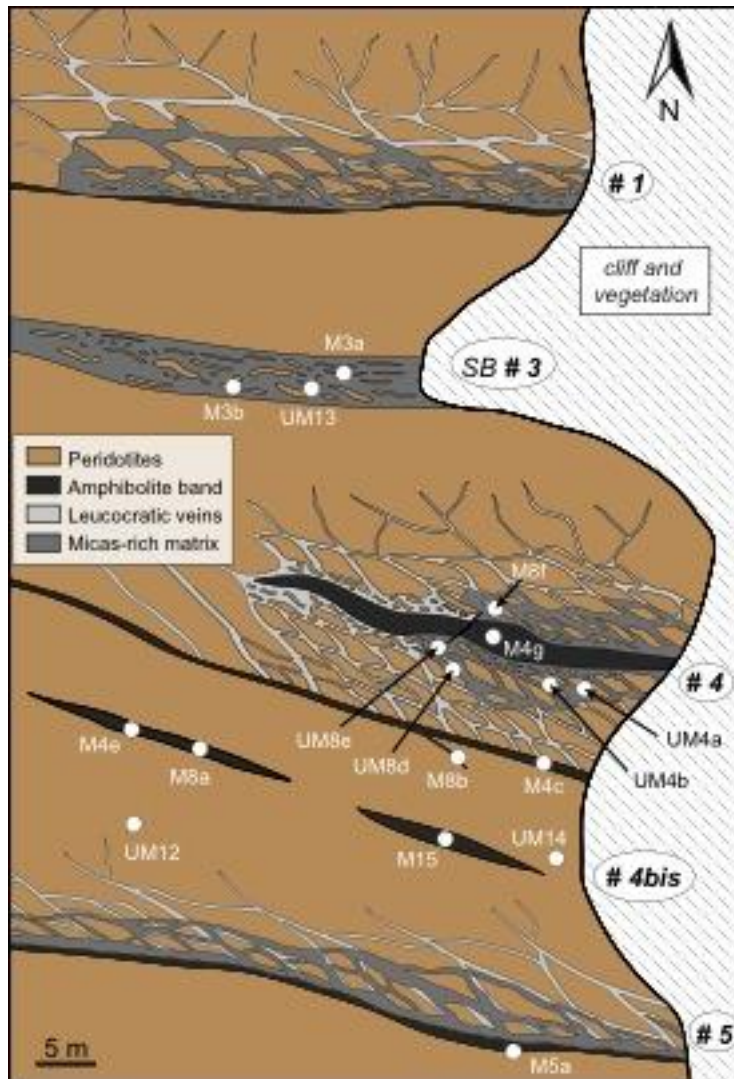
956 of the ophiolite. Shear band (SB) #2 is not represented at this scale.

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958

959 **Figure 2**

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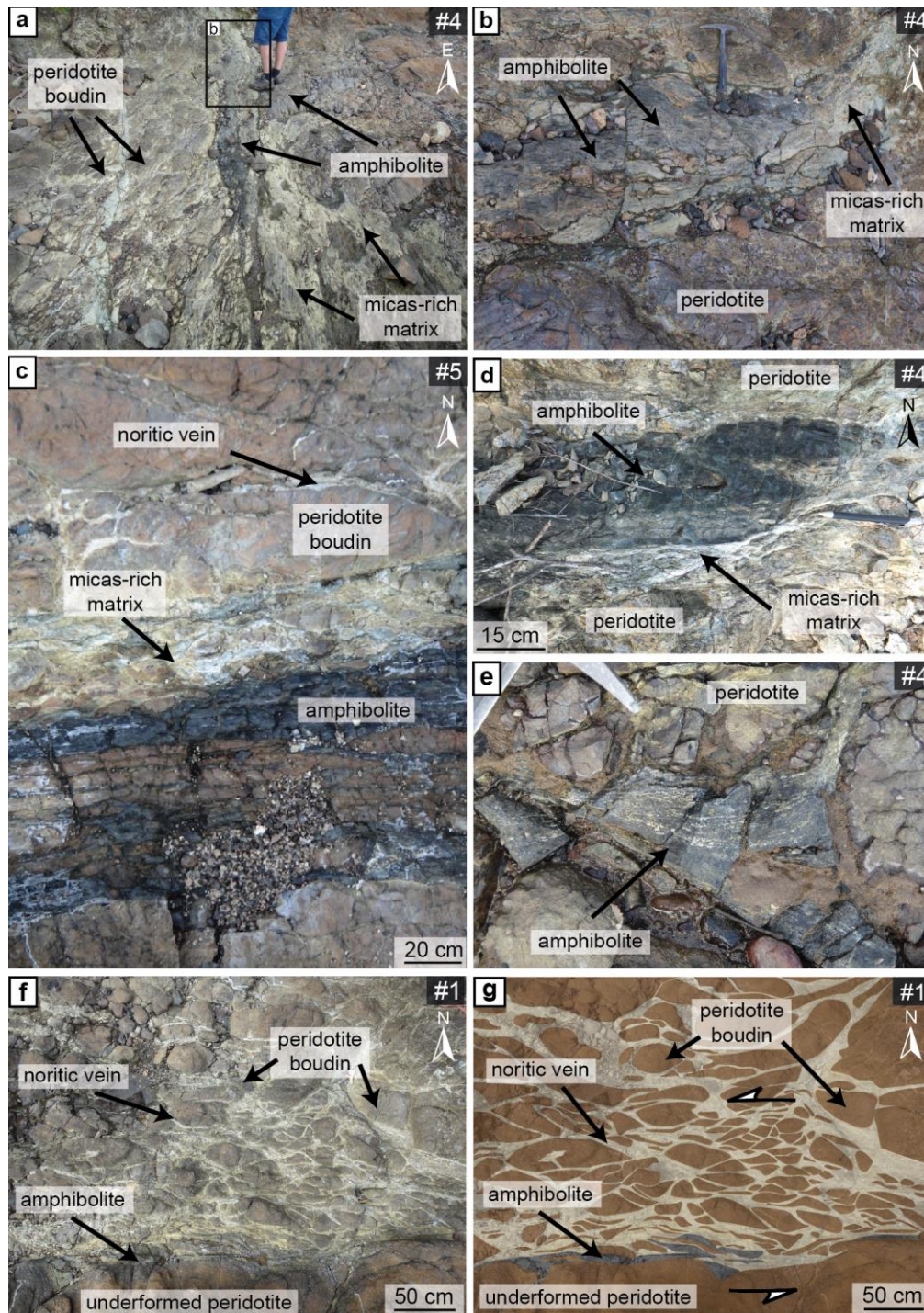
963 Structural map of the northern shear bands. Fig. 2 schematizing the relations between

964 shear bands within the undeformed peridotite of the ophiolite. White dots indicate

965 sample location (Table 1). Shear band (SB) #2 is not represented at this scale.

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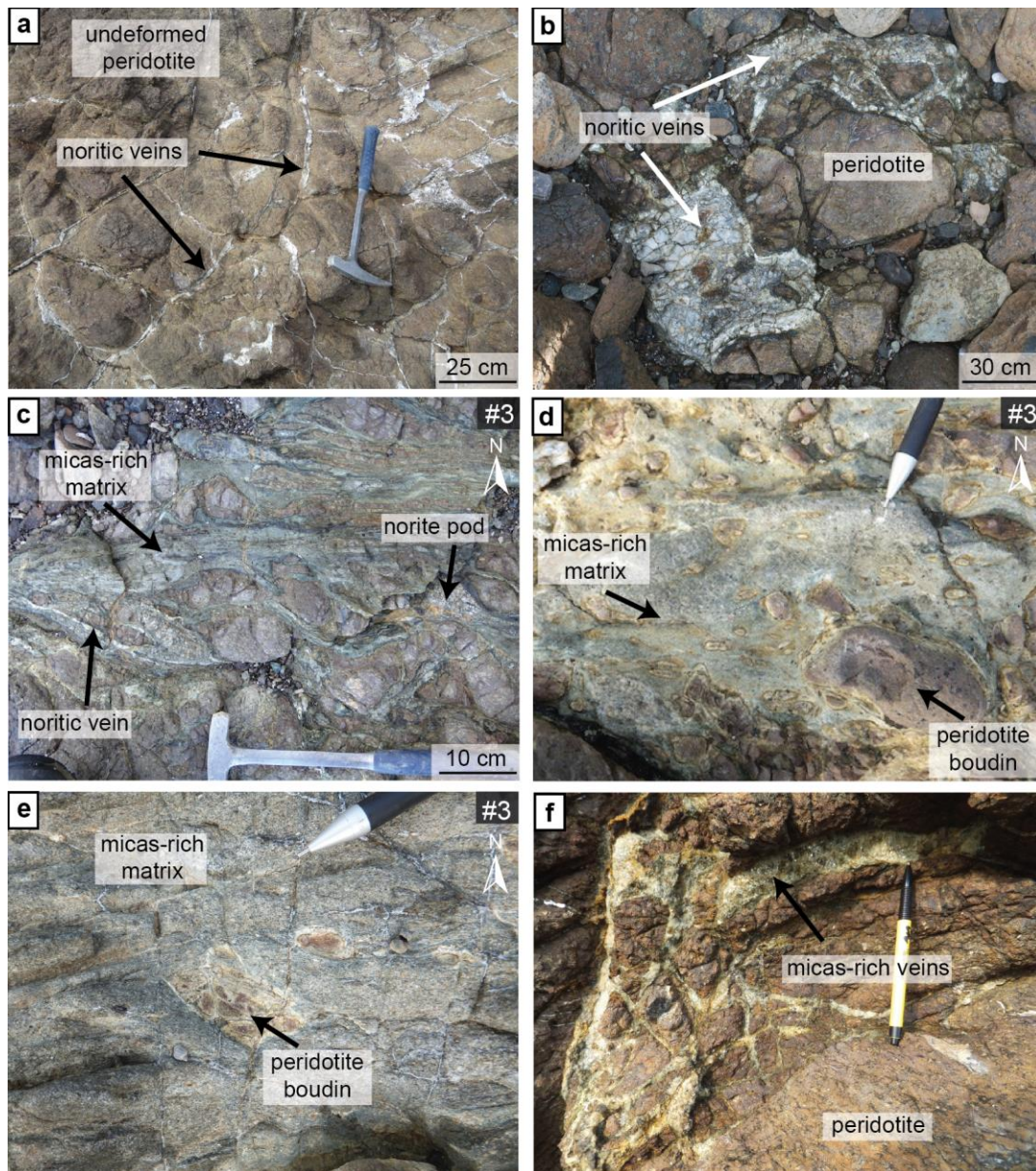
970 Structures of the mafic shear bands. Note that peridotite is highly boudinaged near
 971 dark dm-scale amphibolite bands. a) Amphibolite band found at the core of shear
 972 band #4. b) A small-scale amphibolite band cut by the major amphibolite #4. c) Shear
 973 band #5 is asymmetrical. The vein network is developed only on one side of the

974 amphibolite band (to the NE). d) and e) Boudinaged and foliated amphibolite. f) and

975 g) Photograph and corresponding schematized sketch of asymmetrical shear band #1.

976 Vein density, thickness and length decrease away from the amphibolites.

977

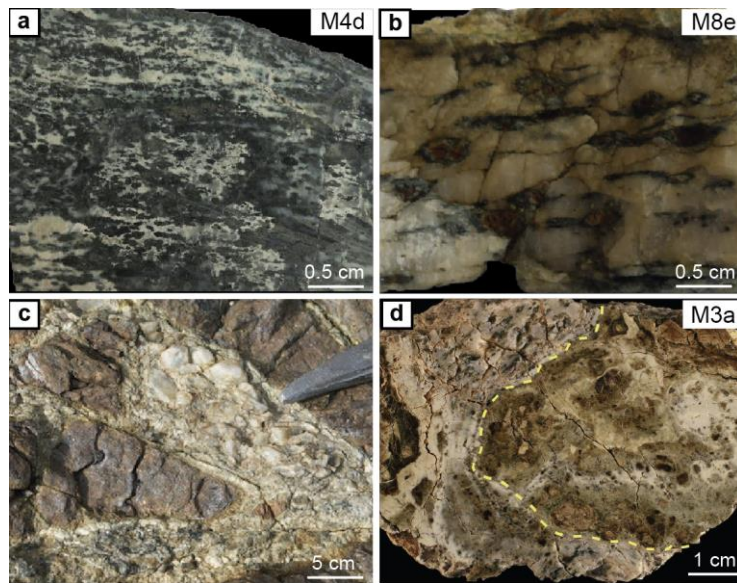


979

980 Photographs of the peridotite and the micas-rich matrix within and away from shear
 981 bands. a) and b) Weakly deformed feldspar-rich veins within undeformed peridotites
 982 away from shear bands. c), d), e) and f) Late light colored matrix with biotite,
 983 chlorite, talc and serpentine minerals within the shear bands. The matrix appears to
 984 form at the expense of both the mafic and ultramafic rocks.

985

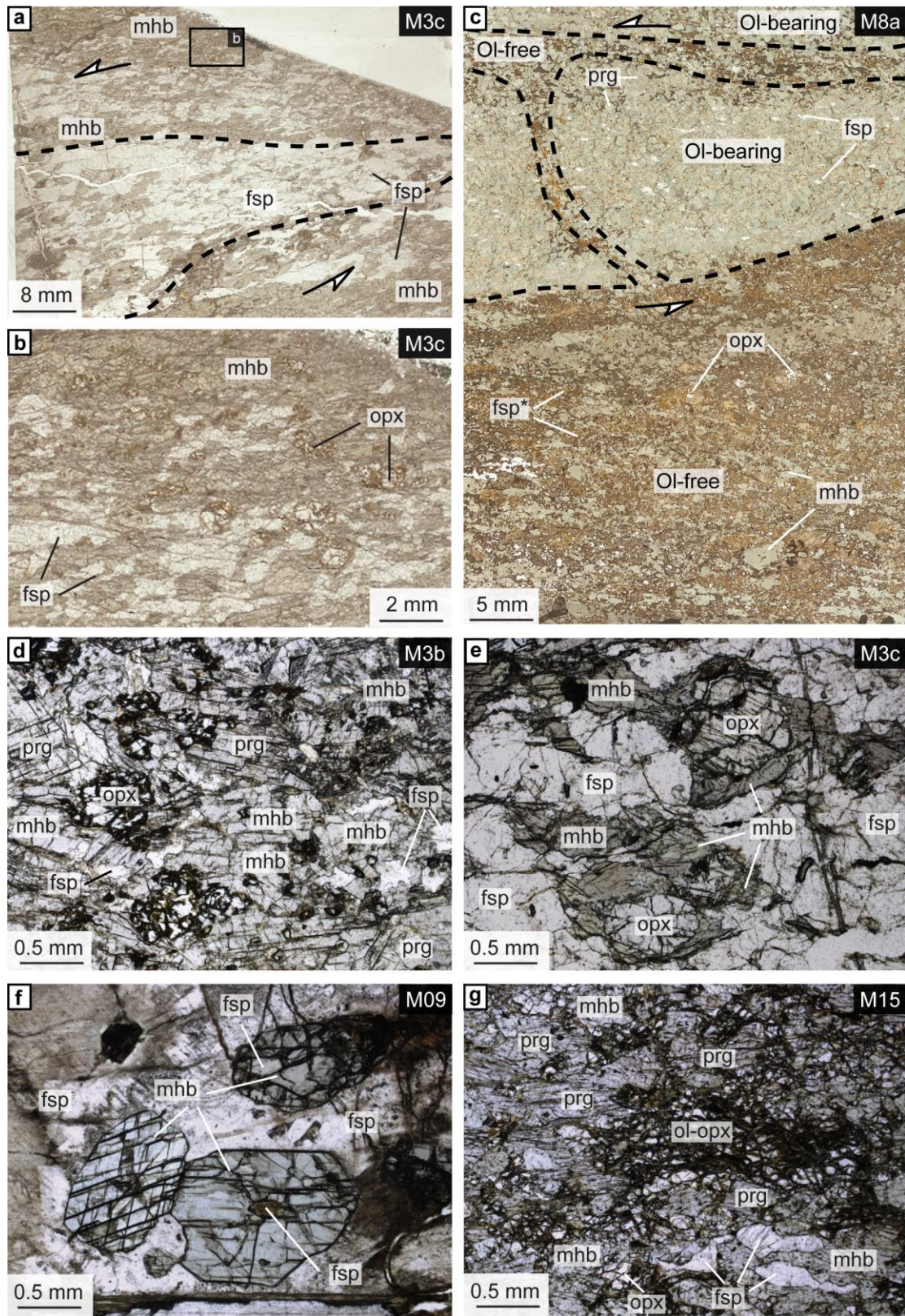
986 **Figure 5**



987

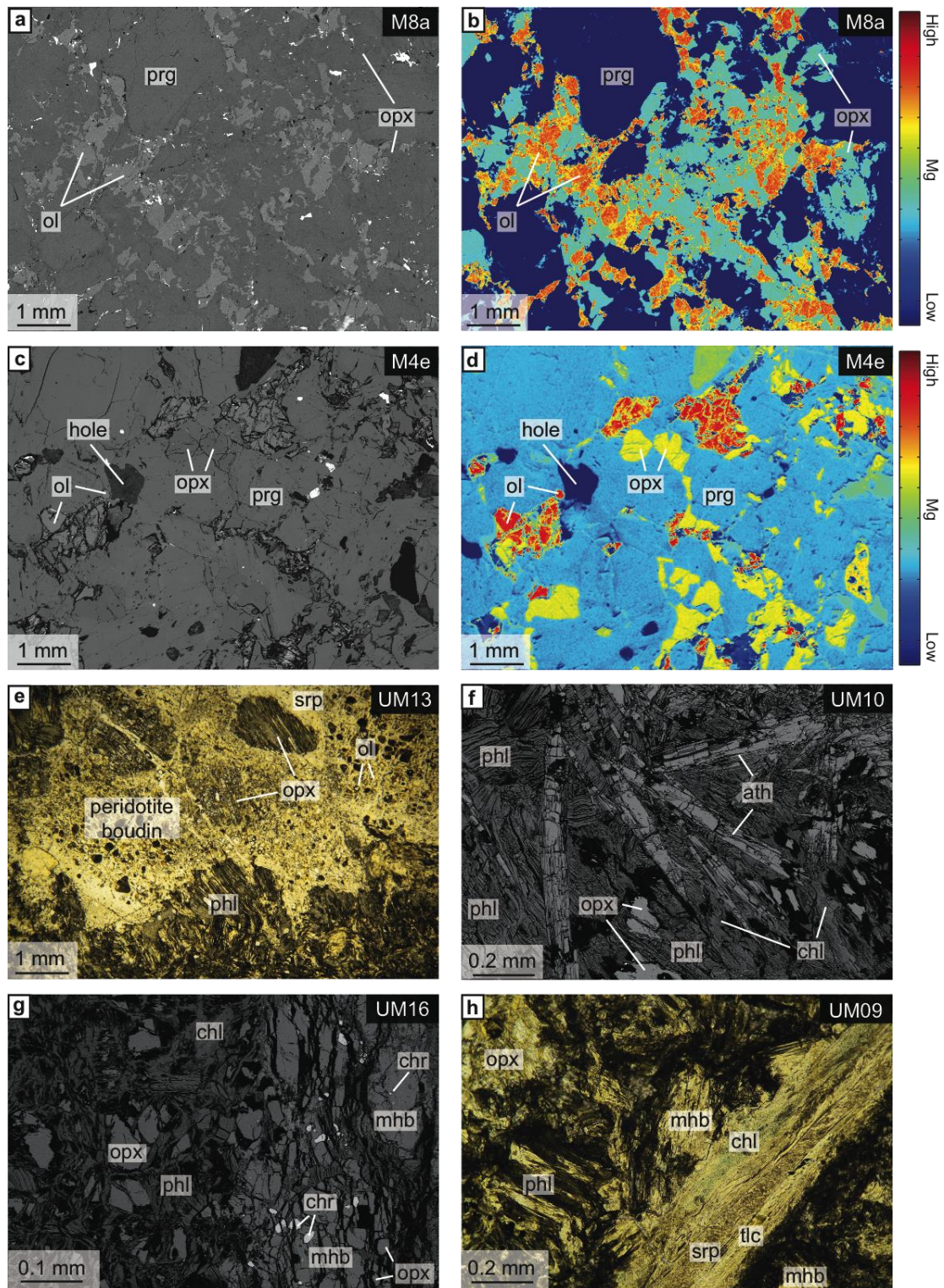
988 Representative samples of the different units observed in shear bands. a) Foliated
989 amphibolite. White layers are mainly composed of plagioclase, darker layers contains
990 amphibole and orthopyroxene. b) Feldspar-rich veins with brown coarse-grained
991 orthopyroxene rimmed by green amphiboles. c) Feldspar-rich vein wrapping around a
992 peridotite boudin. Note the serpentine-bearing white rim around the boudin. d)
993 Peridotite boudin wrapped by feldspar-rich vein and cross-cut by serpentine- talc-
994 chlorite-rich white veinlets.

995



999 Representative mineral assemblages in amphibolite dikes. a) Olivine-free amphibolite
1000 with a boudinaged feldspar-rich layer within a hornblende-rich matrix. b) Zoom into
1001 the hornblende-rich matrix showing the destabilization of orthopyroxene. c) Pale
1002 olivine-bearing amphibolite band boudinaged and cross-cut by an olivine-free
1003 amphibolite layer. d) e) Assemblage of orthopyroxenes and plagioclases destabilized
1004 by the growth of amphiboles. f) Euhedral blue-pale amphibole growing in feldspar-
1005 rich vein away from deformation. g) Olivine, orthopyroxene and plagioclase
1006 boudinaged and destabilized by the syn-cinematic growth of amphibole.
1007 Abbreviations after Whitney and Evans (2010).

1008



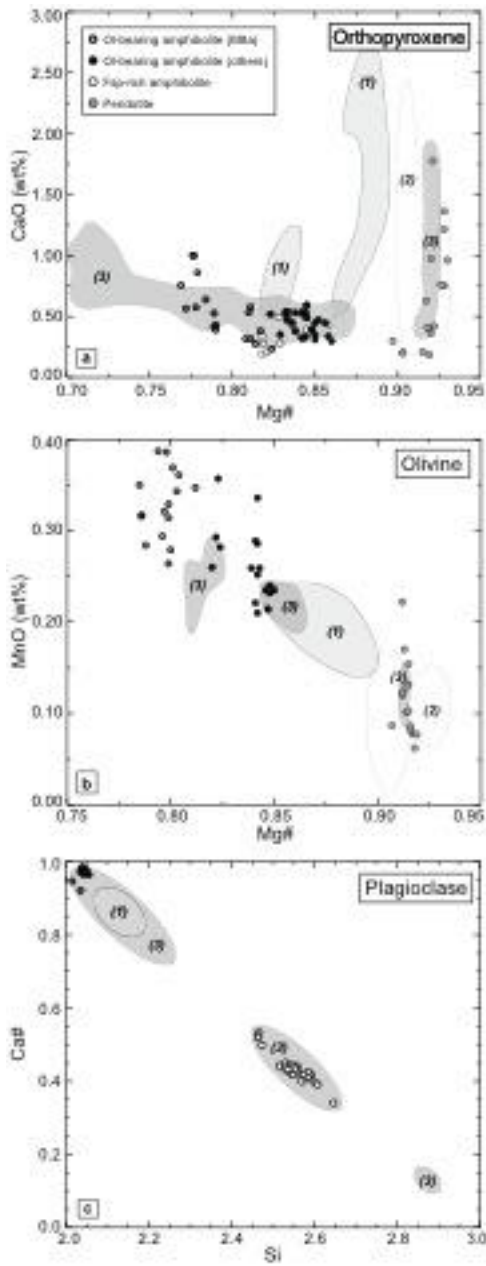
1010

1011 Representative mineral assemblages in olivine-bearing amphibolite and in micas-rich
 1012 matrix. a), b), c) and d) Olivine and plagioclase destabilized by orthopyroxene and
 1013 then amphibole. a) and c) are SEM back-scattered electron images, b) and d) are

1014 corresponding SEM abundancy maps for Mg, where relative concentration is scaled
1015 on the right-end side color bar. e) and f) Microphotograph and SEM back-scattered
1016 electron image of peridotites boudins destabilized by the phlogopite-rich matrix
1017 associated to the anthophyllitic amphiboles. g) Coarse-grained hornblende
1018 pseudomorphing orthopyroxene and partially destabilized by phlogopite. h)
1019 Phlogopite-rich matrix cross-cut by veins of talc, chlorite and serpentine.
1020 Abbreviations after Whitney and Evans (2010).

1021

1022

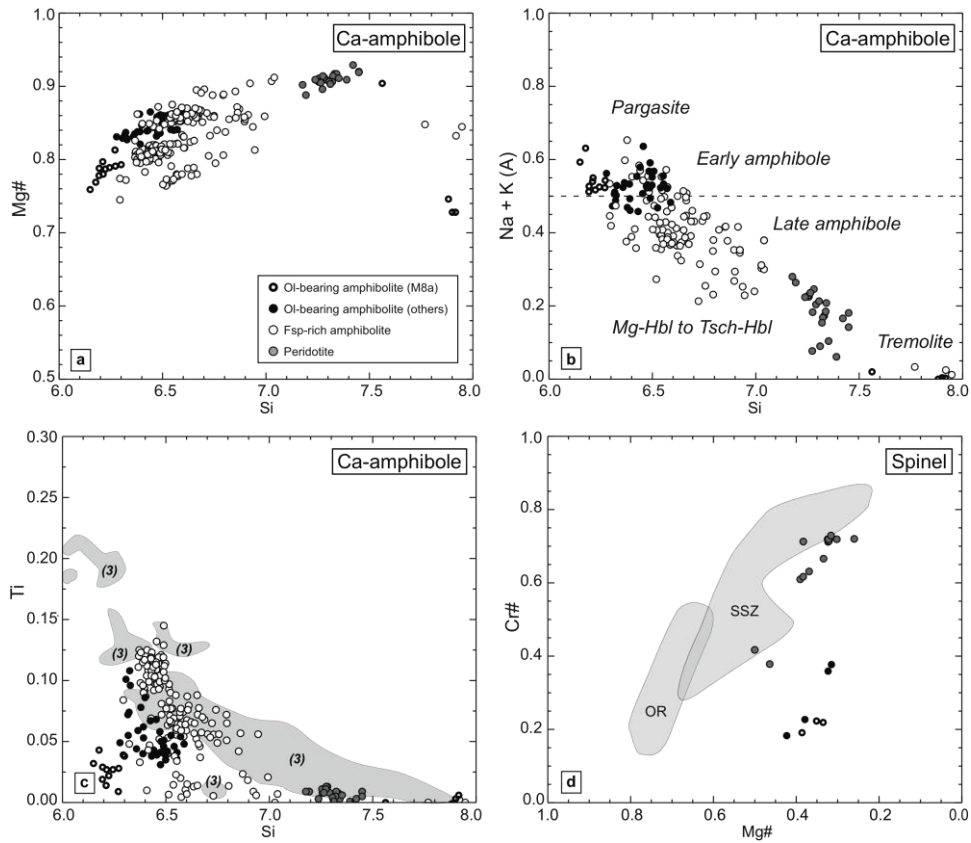


1024

1025 EPMA composition of selected minerals. a) Mg# [Mg/(Mg+Fe)] vs. CaO (wt%) in
 1026 mafic and ultramafic orthopyroxene. b) Mg# vs. MnO (wt%) in mafic and ultramafic
 1027 olivine. c) Si (p.f.u) vs. Ca# [Ca/(Ca+Na)] in plagioclase. (1) Gabbronorite sills
 1028 (crustal cumulates; Pirard et al., 2013) ; (2) Peridotites from Massif du Sud (Marchesi
 1029 et al., 2009) (3) Horblendite and amphibolite dikes (Pirard, 2012).

1030

1031 **Figure 9**



1032

1033

1034 EPMA composition of amphibole and spinel in selected samples. a) Si (p.f.u) vs. Mg#

1035 in Ca-amphibole. b) Si (p.f.u) vs. (Na+K)_A in Ca-amphibole after the classification of

1036 Leake et al. (1997). The dash line separates the pargasite domain (early amphibole)

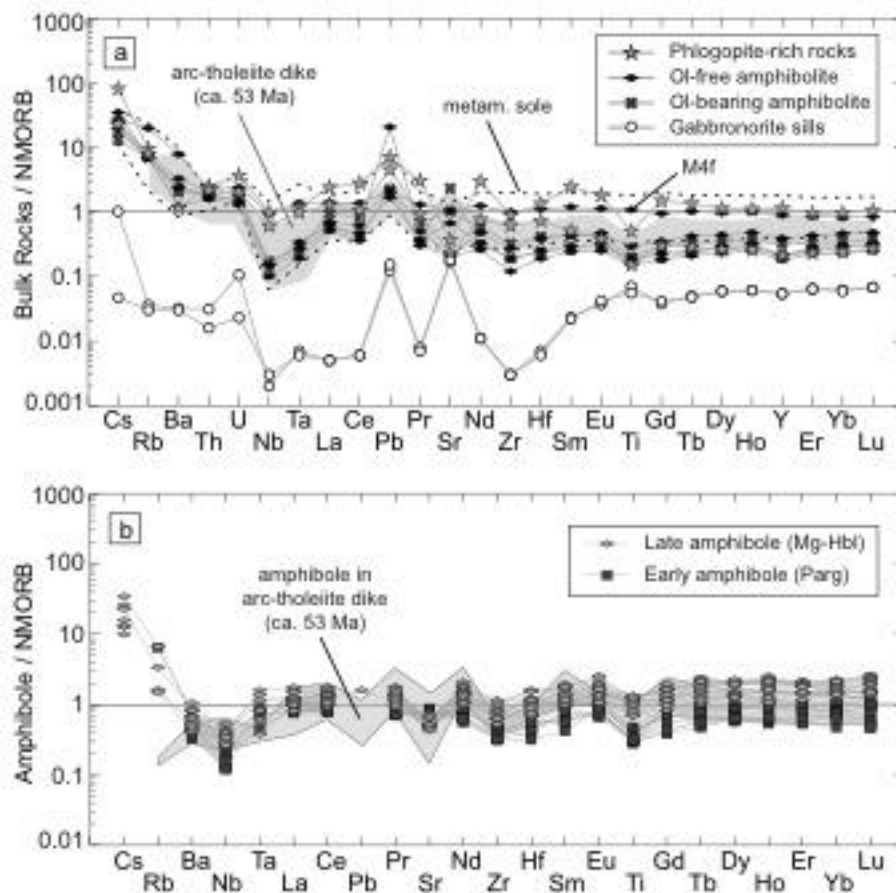
1037 from the Mg-hornblende to tsch-hornblende domains (late amphibole). c) Si (p.f.u) vs

1038 Ti (p.f.u) in Ca-amphibole. d) Mg# vs. Cr# in spinel. Or: Ocean ridge domain. SSZ:

1039 suprasubduction domain. (3) Hornblende and amphibolite dikes (Pirard, 2012)

1040

1041 **Figure 10**



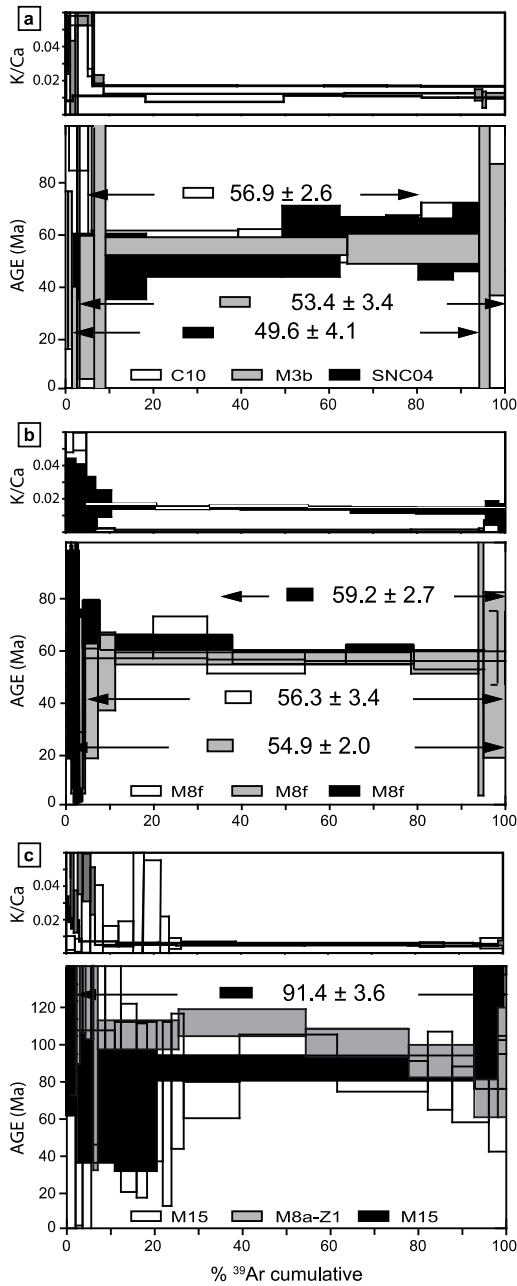
1042

1043 (a) Trace elements concentrations of amphibolite. The N-MORB normalization values
1044 are from Sun and Mcdonough (1989). Data for the metamorphic sole are from
1045 Cluzel et al. (2012) and for pre-obduction dikes (53 Ma) from Cluzel et al. (2006).

1046 Gabbronorite sills from the crustal cumulates are from Pirard et al. (2013). (b) Trace
1047 elements spider-diagrams of hornblende in olivine-free and olivine-bearing
1048 amphibolite M8a. Normalization values for the NMORB are from Sun and
1049 Mcdonough (1989). Grey area represents the trace elements values for amphibole in
1050 hornblendite and diorite dikes described by Pirard (2012).

1051

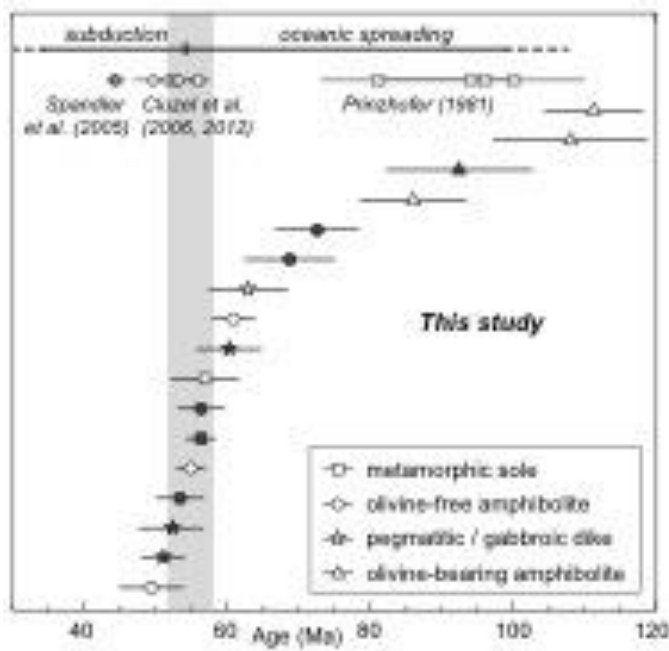
1052



1054

1055 Representative $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra of amphibole where K/Ca and calculated age are
 1056 plotted as function of the cumulative fraction of ^{39}Ar released. Samples: C10:
 1057 metamorphic sole sampled of Thio; SNC04: undeformed gabbroic dikes cross-cutting
 1058 the Massif du Sud ophiolite; M8a-Z1 and M15: olivine-bearing; M8f and M3b:
 1059 olivine-free amphibolite. Corresponding reverse isochron plots ($^{36}\text{Ar}/^{40}\text{Ar}$ vs.
 1060 $^{39}\text{Ar}/^{40}\text{Ar}$) with intercept age are reported in the Online Fig. S1.

1061 **Figure 12**



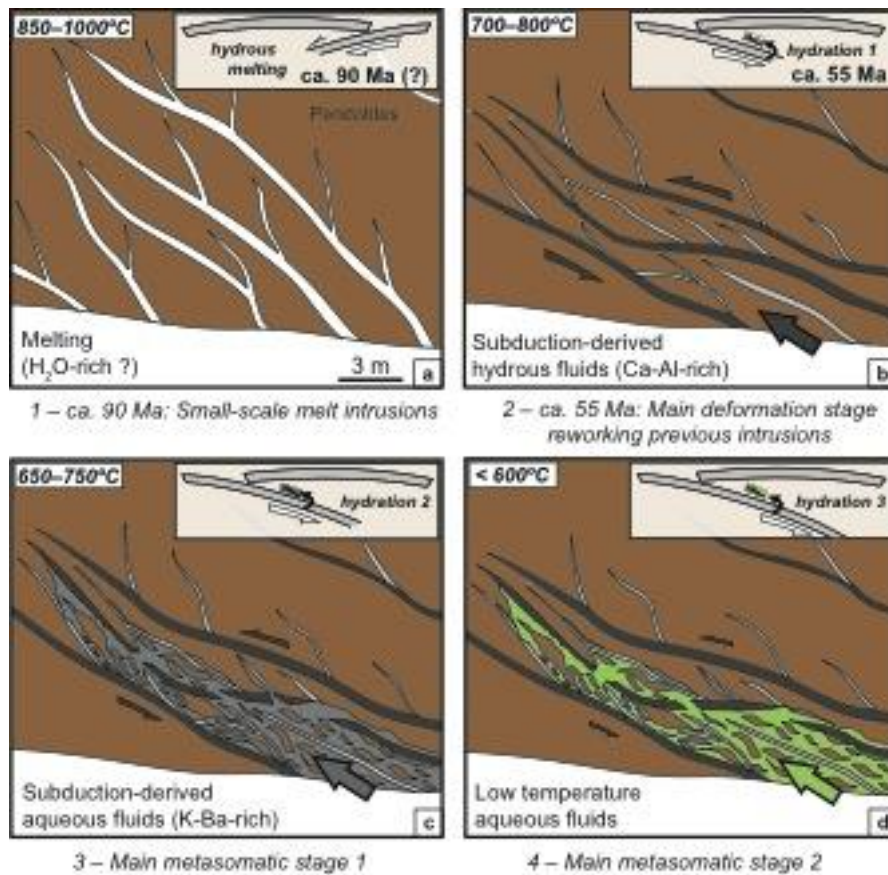
1062

1063

1064 Plots of $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age on amphibole from the different mafic units of the New
1065 Caledonia ophiolite in comparison of the timing of intrusion of pre-obduction dikes
1066 from Cluzel et al. (2006) and the metamorphic sole is from Cluzel et al. (2012). High-
1067 pressure rocks are from Spandler et al. (2005) and Cretaceous gabbroic dikes cross-
1068 cutting the peridotite are from Prinzhofer (1981). White-colored data are for $^{40}\text{Ar}/^{39}\text{Ar}$
1069 plateau ages. Black-colored data are for $^{40}\text{Ar}/^{39}\text{Ar}$ preferred ages (plateau age or
1070 intercept age). See also Table 5.

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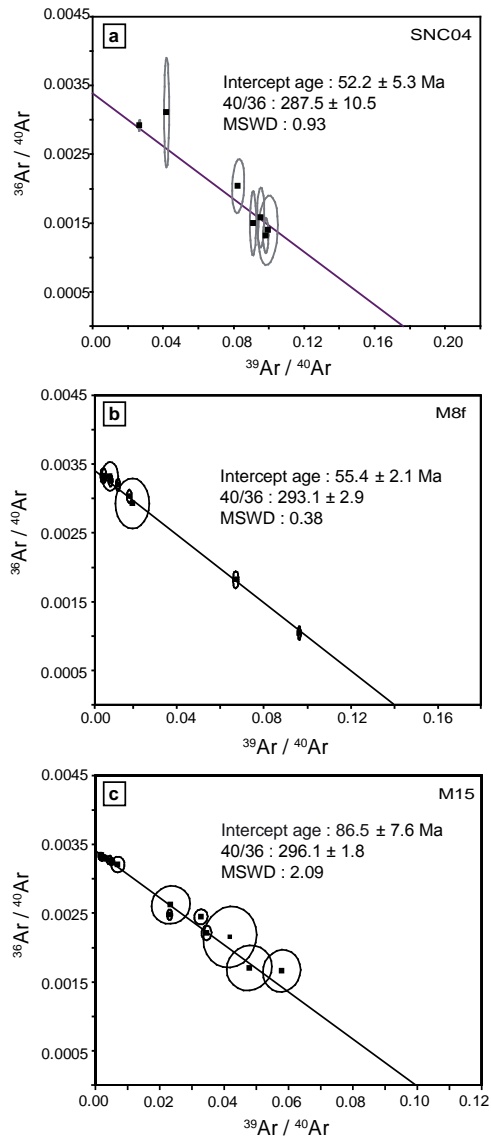
1075 Schematized crystallization and deformation events for the emplacement of the mafic
 1076 dikes at the study area scale. The inset shows the corresponding simplified
 1077 geodynamic evolution with successive stages of infiltration of subduction-related
 1078 fluids. a) Small-scale melt intrusions, enriched in Ca and Al compared to the
 1079 surrounding harzburgites. Fluids at the origin of these mafic veins (with
 1080 orthopyroxene, olivine, plagioclase ± early amphibole) were probably gabbroic melts
 1081 derived from hydrated melting of the peridotite, caused by slab dehydration. b)
 1082 Amphibolitization stage is the main deformation stage reworking previous mafic
 1083 intrusions mechanically weaker than the peridotite host. Sheared amphibolites
 1084 indicate that these mafic intrusions were deformed shortly after emplacement, as
 1085 suggested by amphibole ages at 55 ± 2 Ma. c) The following emplacement of the
 1086 phlogopite-rich matrix records the main metasomatic stage. Anthophyllite indicates

1087 temperature around 700 °C. The intensity of deformation decreased significantly
1088 compared to b). d) Talc–chlorite–serpentine assemblage representing the last stage of
1089 metasomatism affecting the shear bands and likely starting around temperatures of
1090 600 °C.

1091

1092

1094 Figure S1



1095

1096 Corresponding reverse isochron plots ($^{36}\text{Ar}/^{40}\text{Ar}$ vs. $^{39}\text{Ar}/^{40}\text{Ar}$) are shown for some of
 1097 the measurements, with intercept age, $^{36}\text{Ar}/^{40}\text{Ar}$ ratio and final reduced chi square
 1098 (MSWD). Samples: SNC04: undeformed gabbroic dikes cross-cutting the Massif du
 1099 Sud ophiolite; M15: olivine-bearing amphibolite; M8f and M3b: olivine-free
 1100 amphibolite.

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1102 **Tables**

Table 1: Mineral occurrence for selected samples.

| Unit | Sample | ol | opx | fsp | Ba-fsp | hbl | ath | phl | chl/tlc | srp |
|------------------------|--------|----|-----|-----|--------|-----|-----|-----|---------|-----|
| Undeformed peridotite | M12 | x | x | | | | | | | 1 |
| | M14 | x | x | | | | | | | 1 |
| Peridotite boudin | M9 | x | x | v | | | o | o | 1 | 1 |
| | M10 | x | x | v | o | o | o | o | 1 | 1 |
| | M16 | x | x | v | | o | o | o | 1 | 1 |
| Ol-bearing amphibolite | M8a | x | x | x | | o | | | 1 | |
| | M8b | x | x | x | | o | | | 1 | |
| | M4e | x | x | x | | o | | | | |
| | M15 | x | x | x | | o | | | | |
| Ol-free amphibolite | M3c | | x | x | | o | | | | |
| | M3b | | x | x | | o | | | | |
| | M4c | | x | x | | o | | | | |
| | Mx | | x | x | | o | | | | |
| | M4d | | x | x | | o | | 1 | 1 | |
| | M4g | | x | x | | o | | 1 | 1 | |
| | 8e | | x | x | | o | | | | |

x: primary phases, o: secondary phases; l: late phases; v: vein wrapping peridotite boudin.
 Mineral abbreviations after Whitney and Evans (2010). Ba-fsp: celsian.

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Table 2: Selected representative EPMA analyses for olivine, orthopyroxene, plagioclase and amphibole. Fe³⁺ for amphibole calculated after Leake et al. (1997)

| Olivine | | | | | Orthopyroxene | | | | | | | | |
|--------------------------------|--------------|--------------|--------------|--------------|--------------------------------|--------------|--------------|--------------|--------------|--------------|--------------|--------------|--------------|
| Min. | | | | | Min. | | | | | | | | |
| Samp. | M4e | M8a1 | M8b | UM10 | Samp. | M3c | M4e | M8a1 | M8a2 | M8b | M8e | UM10 | UM16 |
| SiO ₂ | 40.05 | 39.56 | 40.24 | 41.62 | SiO ₂ | 55.85 | 55.02 | 54.51 | 54.69 | 56.04 | 55.86 | 57.44 | 58.08 |
| TiO ₂ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | TiO ₂ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | 0.12 | <i>b.d.l</i> | <i>b.d.l</i> |
| Al ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Al ₂ O ₃ | 1.34 | 2.12 | 1.88 | 1.23 | 1.60 | 1.55 | 1.50 | 0.70 |
| Cr ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Cr ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | 0.08 | <i>b.d.l</i> | 0.07 | <i>b.d.l</i> | 0.72 | 0.09 |
| FeO | 14.25 | 18.31 | 16.12 | 8.14 | FeO | 11.12 | 10.00 | 13.74 | 14.78 | 10.85 | 10.38 | 5.24 | 6.69 |
| MnO | 0.23 | 0.28 | 0.36 | 0.08 | MnO | 0.23 | 0.25 | 0.49 | 0.42 | 0.28 | 0.25 | 0.08 | 0.11 |
| MgO | 45.46 | 41.73 | 43.11 | 50.38 | MgO | 31.24 | 32.28 | 28.76 | 28.13 | 30.26 | 31.78 | 34.94 | 34.89 |
| CaO | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | CaO | 0.28 | 0.47 | 0.53 | 0.57 | 0.54 | 0.39 | 0.42 | 0.20 |
| Na ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Na ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| K ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | K ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| BaO | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> |
| Sum | 100.0 | 99.9 | 99.8 | 100.2 | Total | 100.1 | 100.1 | 100.0 | 99.8 | 99.7 | 100.3 | 100.3 | 100.8 |
| Formula unit | | | | | | | | | | | | | |
| Si | 1.00 | 1.01 | 1.02 | 1.01 | Si | 1.96 | 1.92 | 1.94 | 1.96 | 1.98 | 1.95 | 1.97 | 1.99 |
| Ti | 0.00 | 0.00 | 0.00 | 0.00 | Ti | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Al | 0.00 | 0.00 | 0.00 | 0.00 | Al | 0.06 | 0.09 | 0.08 | 0.05 | 0.07 | 0.06 | 0.06 | 0.03 |
| Cr | 0.00 | 0.00 | 0.00 | 0.00 | Cr | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.02 | 0.00 |
| Fe 3+ | 0.00 | 0.00 | 0.00 | 0.00 | Fe 3+ | 0.02 | 0.08 | 0.03 | 0.02 | 0.00 | 0.02 | 0.00 | 0.00 |
| Fe 2+ | 0.30 | 0.39 | 0.34 | 0.00 | Fe 2+ | 0.31 | 0.21 | 0.38 | 0.42 | 0.32 | 0.28 | 0.15 | 0.19 |
| Mn | 0.00 | 0.01 | 0.01 | 0.17 | Mn | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.00 | 0.00 |
| Mg | 1.70 | 1.59 | 1.63 | 0.00 | Mg | 1.64 | 1.68 | 1.53 | 1.50 | 1.60 | 1.66 | 1.78 | 1.78 |
| Ca | 0.00 | 0.00 | 0.00 | 1.82 | Ca | 0.01 | 0.02 | 0.02 | 0.02 | 0.02 | 0.01 | 0.02 | 0.01 |
| Na | 0.00 | 0.00 | 0.00 | 0.00 | Na | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| K | 0.00 | 0.00 | 0.00 | 0.00 | K | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Ba | - | - | - | - | Ba | - | - | - | - | - | - | - | - |
| Mg# | 0.85 | 0.80 | 0.83 | 0.92 | Mg# | 0.83 | 0.85 | 0.79 | 0.77 | 0.83 | 0.85 | 0.92 | 0.90 |

b.d.l.: below detection limit; *n.a.*: non analyzed

Table 2: Selected representative EPMA analyses for olivine, orthopyroxene, plagioclase and amphibole. Fe³⁺ for amphibole calculated after Leake et al. (1997)

| Olivine | | | | | Orthopyroxene | | | | | | | | | Feldspar | | | | | |
|--------------------------------|--------------|--------------|--------------|--------------|--------------------------------|--------------|--------------|--------------|--------------|--------------|--------------|--------------|--------------|--------------------------------|--------------|--------------|--------------|--------------|--------------|
| Min. | | | | | Min. | | | | | | | | | Min. | | | | | |
| Samp. | M4e | M8a1 | M8b | UM10 | Samp. | M3c | M4e | M8a1 | M8a2 | M8b | M8e | UM10 | UM16 | Samp. | M8a1 | M8a1 | M8e | UM9 | UM10 |
| SiO ₂ | 40.05 | 39.56 | 40.24 | 41.62 | SiO ₂ | 55.85 | 55.02 | 54.51 | 54.69 | 56.04 | 55.86 | 57.44 | 58.08 | SiO ₂ | 43.24 | 43.55 | 56.68 | 56.62 | 51.19 |
| TiO ₂ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | TiO ₂ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | 0.12 | <i>b.d.l</i> | <i>b.d.l</i> | TiO ₂ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| Al ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Al ₂ O ₃ | 1.34 | 2.12 | 1.88 | 1.23 | 1.60 | 1.55 | 1.50 | 0.70 | Al ₂ O ₃ | 37.18 | 36.77 | 27.62 | 26.70 | 22.52 |
| Cr ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Cr ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | 0.08 | <i>b.d.l</i> | 0.07 | <i>b.d.l</i> | 0.72 | 0.09 | Cr ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| FeO | 14.25 | 18.31 | 16.12 | 8.14 | FeO | 11.12 | 10.00 | 13.74 | 14.78 | 10.85 | 10.38 | 5.24 | 6.69 | FeO | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| MnO | 0.23 | 0.28 | 0.36 | 0.08 | MnO | 0.23 | 0.25 | 0.49 | 0.42 | 0.28 | 0.25 | 0.08 | 0.11 | MnO | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| MgO | 45.46 | 41.73 | 43.11 | 50.38 | MgO | 31.24 | 32.28 | 28.76 | 28.13 | 30.26 | 31.78 | 34.94 | 34.89 | MgO | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| CaO | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | CaO | 0.28 | 0.47 | 0.53 | 0.57 | 0.54 | 0.39 | 0.42 | 0.20 | CaO | 20.33 | 19.40 | 9.13 | 9.48 | 0.13 |
| Na ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Na ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | Na ₂ O | 0.21 | 0.54 | 6.47 | 6.61 | 0.46 |
| K ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | K ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | K ₂ O | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | 7.92 |
| BaO | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | BaO | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | <i>n.a</i> | 15.78 |
| Sum | 100.0 | 99.9 | 99.8 | 100.2 | Total | 100.1 | 100.1 | 100.0 | 99.8 | 99.7 | 100.3 | 100.3 | 100.8 | Sum | 101.0 | 100.3 | 99.9 | 99.4 | 98.0 |
| Formula unit | | | | | | | | | | | | | | | | | | | |
| Si | 1.00 | 1.01 | 1.02 | 1.01 | Si | 1.96 | 1.92 | 1.94 | 1.96 | 1.98 | 1.95 | 1.97 | 1.99 | Si | 1.98 | 2.00 | 2.46 | 2.54 | 2.65 |
| Ti | 0.00 | 0.00 | 0.00 | 0.00 | Ti | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | Ti | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Al | 0.00 | 0.00 | 0.00 | 0.00 | Al | 0.06 | 0.09 | 0.08 | 0.05 | 0.07 | 0.06 | 0.06 | 0.03 | Al | 2.00 | 1.99 | 1.52 | 1.42 | 1.37 |
| Cr | 0.00 | 0.00 | 0.00 | 0.00 | Cr | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.02 | 0.00 | Cr | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Fe 3+ | 0.00 | 0.00 | 0.00 | 0.00 | Fe 3+ | 0.02 | 0.08 | 0.03 | 0.02 | 0.00 | 0.02 | 0.00 | 0.00 | Fe 3+ | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Fe 2+ | 0.30 | 0.39 | 0.34 | 0.00 | Fe 2+ | 0.31 | 0.21 | 0.38 | 0.42 | 0.32 | 0.28 | 0.15 | 0.19 | Fe 2+ | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Mn | 0.00 | 0.01 | 0.01 | 0.17 | Mn | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.00 | 0.00 | Mn | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Mg | 1.70 | 1.59 | 1.63 | 0.00 | Mg | 1.64 | 1.68 | 1.53 | 1.50 | 1.60 | 1.66 | 1.78 | 1.78 | Mg | 0.00 | 0.00 | 0.00 | 0.00 | 0.03 |
| Ca | 0.00 | 0.00 | 0.00 | 1.82 | Ca | 0.01 | 0.02 | 0.02 | 0.02 | 0.02 | 0.01 | 0.02 | 0.01 | Ca | 1.00 | 0.95 | 0.53 | 0.46 | 0.01 |
| Na | 0.00 | 0.00 | 0.00 | 0.00 | Na | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | Na | 0.02 | 0.05 | 0.48 | 0.58 | 0.05 |
| K | 0.00 | 0.00 | 0.00 | 0.00 | K | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | K | 0.00 | 0.00 | 0.00 | 0.00 | 0.36 |
| Ba | - | - | - | - | Ba | - | - | - | - | - | - | - | - | Ba | - | - | - | - | 0.52 |
| Mg# | 0.85 | 0.80 | 0.83 | 0.92 | Mg# | 0.83 | 0.85 | 0.79 | 0.77 | 0.83 | 0.85 | 0.92 | 0.90 | Ca# | 0.98 | 0.95 | 0.52 | 0.44 | 0.13 |

Table 2 (continued)

| Samp. | Amphibole | | | | | | | | | | | | | |
|--|--------------|--------------|--------------|--------------|-------------|------------|--------------|--------------|--------------|-------------|--------------|--------------|--------------|--------------|
| | M3c mhb | M3c mhb | M8a1 prg | M8a1 tr | M8a2 mhb | M8b prg | M8e mhb | M8e mhb | M09 mhb | UM10 mhb | UM10 tr | UM10 ath | UM16 mhb | UM16 tr |
| SiO ₂ | 47.08 | 45.41 | 43.33 | 55.40 | 46.30 | 46.11 | 46.04 | 49.96 | 46.99 | 48.13 | 53.96 | 57.08 | 51.52 | 54.23 |
| TiO ₂ | 0.19 | 1.01 | 0.40 | <i>b.d.l</i> | 0.79 | 0.39 | 0.06 | 0.06 | 0.68 | 0.44 | 0.07 | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> |
| Al ₂ O ₃ | 11.10 | 11.47 | 15.16 | 3.14 | 11.82 | 12.24 | 11.54 | 8.42 | 11.66 | 10.58 | 5.73 | 1.57 | 8.44 | 4.76 |
| Cr ₂ O ₃ | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | 0.10 | 0.44 | <i>b.d.l</i> | <i>b.d.l</i> | <i>b.d.l</i> | 0.85 | <i>b.d.l</i> | 0.05 | <i>b.d.l</i> | <i>b.d.l</i> |
| FeO | 7.03 | 7.38 | 8.20 | 4.16 | 8.85 | 6.11 | 7.20 | 6.32 | 5.86 | 4.09 | 3.51 | 9.14 | 3.83 | 3.05 |
| MnO | 0.18 | 0.05 | 0.10 | 0.12 | 0.09 | 0.08 | 0.15 | 0.17 | 0.09 | 0.06 | 0.04 | 0.26 | 0.05 | 0.12 |
| MgO | 18.26 | 17.64 | 15.27 | 21.91 | 16.08 | 17.47 | 17.55 | 19.28 | 18.88 | 19.62 | 21.81 | 28.58 | 20.97 | 22.41 |
| CaO | 11.51 | 11.27 | 12.03 | 12.73 | 11.71 | 12.35 | 11.38 | 11.29 | 11.52 | 11.72 | 12.06 | 0.73 | 12.02 | 11.90 |
| Na ₂ O | 2.20 | 2.50 | 2.80 | 0.53 | 2.13 | 2.38 | 2.47 | 1.82 | 2.20 | 2.48 | 1.26 | 0.12 | 1.67 | 1.09 |
| K ₂ O | 0.12 | 0.15 | 0.10 | <i>b.d.l</i> | 0.06 | 0.07 | 0.13 | 0.08 | 0.13 | 0.14 | 0.05 | <i>b.d.l</i> | 0.08 | <i>b.d.l</i> |
| Sum | 97.67 | 96.89 | 97.38 | 97.99 | 97.94 | 97.63 | 96.51 | 97.41 | 98.04 | 98.11 | 98.50 | 97.51 | 98.58 | 97.56 |
| Si | 6.57 | 6.41 | 6.18 | 7.56 | 6.52 | 6.49 | 6.51 | 6.94 | 6.52 | 6.66 | 7.33 | 7.77 | 7.03 | 7.42 |
| Ti | 0.02 | 0.11 | 0.04 | 0.00 | 0.08 | 0.42 | 0.01 | 0.01 | 0.07 | 0.05 | 0.01 | 0.00 | 0.00 | 0.00 |
| Al | 1.83 | 1.90 | 2.55 | 0.51 | 1.96 | 2.03 | 1.92 | 1.38 | 1.90 | 1.73 | 0.92 | 0.25 | 0.97 | 0.77 |
| Cr | 0.00 | 0.00 | 0.00 | 0.00 | 0.01 | 0.05 | 0.00 | 0.00 | 0.00 | 0.09 | 0.00 | 0.01 | 0.00 | 0.00 |
| Fe 3+ | 0.82 | 0.87 | 0.55 | 0.47 | 0.70 | 0.48 | 0.85 | 0.73 | 0.68 | 0.47 | 0.40 | 0.17 | 0.35 | 0.44 |
| Fe 2+ | 0.00 | 0.00 | 0.43 | 0.00 | 0.34 | 0.24 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.87 | 0.00 | 0.00 |
| Mn | 0.02 | 0.01 | 0.01 | 0.01 | 0.00 | 0.01 | 0.02 | 0.02 | 0.01 | 0.01 | 0.01 | 0.03 | 0.01 | 0.01 |
| Mg | 3.70 | 3.72 | 3.24 | 0.46 | 3.37 | 3.66 | 3.70 | 3.99 | 3.90 | 4.05 | 4.42 | 5.80 | 4.57 | 4.26 |
| Ca | 1.72 | 1.71 | 1.84 | 1.86 | 1.77 | 1.86 | 1.73 | 1.68 | 1.71 | 1.74 | 1.76 | 0.11 | 1.75 | 1.76 |
| Na | 0.60 | 0.69 | 0.78 | 0.14 | 0.58 | 0.65 | 0.68 | 0.49 | 0.59 | 0.66 | 0.33 | 0.03 | 0.29 | 0.44 |
| K | 0.02 | 0.03 | 0.02 | 0.00 | 0.01 | 0.01 | 0.02 | 0.02 | 0.02 | 0.02 | 0.01 | 0.00 | 0.01 | 0.00 |
| Mg# | 0.82 | 0.81 | 0.77 | 0.49 | 0.76 | 0.84 | 0.81 | 0.84 | 0.86 | 0.90 | 0.92 | 0.85 | 0.93 | 0.91 |
| Geothermobarometry after Holland and Blundy (1994) modified by Anderson and Smith (1995) | | | | | | | | | | | | | | |
| An# | 0.52 | 0.52 | 0.95 | - | 0.52 | 0.52 | 0.52 | 0.52 | 0.44 | 0.44 | - | - | 0.44 | - |
| T (°C) | 718 | 776 | 1013 | - | 756 | 963 | 739 | 702 | 680 | 677 | - | - | 615 | - |
| P (kbar) | 5.16 | 4.33 | error | - | 4.98 | error | 5.23 | 3.34 | 5.97 | 3.61 | - | - | 3.78 | - |

Table 3: Whole rock geochemistry for representative amphibolite dikes.

| Sample | M15_Z1 amph | M4f amph | M15_Z2 ol-amph | M8a ol-amph | M4e ol-amph | M4a amph(1) | M4g amph(1) |
|--------------------------------|----------------|-------------|-------------------|----------------|----------------|----------------|----------------|
| SiO ₂ (wt %) | 43.83 | 49.45 | 43.61 | 44.66 | 43.09 | 46.96 | 45.81 |
| TiO ₂ | 0.38 | 1.38 | 0.00 | 0.26 | 0.24 | 0.19 | 0.64 |
| Al ₂ O ₃ | 7.41 | 14.09 | 9.35 | 10.41 | 8.94 | 7.29 | 10.13 |
| Fe ₂ O ₃ | 9.59 | 9.78 | 11.10 | 8.12 | 9.89 | 5.88 | 5.59 |
| MnO | 0.18 | 0.15 | 0.17 | 0.14 | 0.15 | 0.07 | 0.08 |
| MgO | 21.54 | 8.12 | 20.99 | 20.30 | 25.40 | 25.86 | 21.90 |
| CaO | 7.65 | 9.71 | 8.25 | 9.21 | 7.03 | 2.60 | 8.56 |
| Na ₂ O | 1.42 | 3.96 | 1.70 | 1.80 | 1.61 | 1.53 | 1.61 |
| K ₂ O | <i>b.d.l.</i> | 0.74 | <i>b.d.l.</i> | 0.02 | <i>b.d.l.</i> | 0.10 | 0.07 |
| P ₂ O ₅ | 0.03 | 0.13 | 0.03 | 0.06 | 0.03 | 0.02 | 0.03 |
| Perte 1000 | 6.41 | 1.52 | 3.16 | 4.72 | 3.54 | 9.47 | 5.21 |
| Sum | 98.43 | 99.03 | 98.60 | 99.69 | 99.91 | 99.78 | 99.00 |
| Cr (ppm) | 960.00 | 336.00 | 1249.00 | 1652.00 | 2212.00 | 1140.00 | 802.00 |
| Ni | 348.00 | 94.00 | 523.00 | 544.00 | 819.00 | 922.00 | 648.00 |
| Cs | 0.14 | 0.19 | 0.17 | 0.12 | 0.09 | 0.16 | 0.56 |
| Rb | 3.90 | 11.10 | 4.00 | 4.00 | 3.90 | 4.40 | 5.30 |
| Ba | 20.70 | 49.40 | 14.90 | 12.80 | 15.10 | <i>n.d.</i> | 7.10 |
| Th | 0.25 | 0.27 | 0.22 | 0.22 | 0.21 | 0.26 | 0.30 |
| U | 0.09 | 0.11 | 0.07 | 0.08 | 0.07 | 0.09 | 0.17 |
| Nb | 0.37 | 2.24 | 0.34 | 0.40 | 0.31 | 1.42 | 2.28 |
| Ta | 0.04 | 0.18 | 0.05 | 0.04 | 0.03 | 0.13 | 0.16 |
| La | 1.76 | 3.49 | 1.53 | 2.37 | 1.39 | 3.19 | 5.96 |
| Ce | 4.56 | 10.26 | 3.48 | 6.16 | 3.29 | 8.03 | 20.82 |
| Pb | 6.19 | 0.62 | 0.50 | 0.66 | 0.64 | 1.36 | 2.12 |
| Pr | 0.66 | 1.72 | 0.49 | 0.90 | 0.46 | 1.22 | 4.01 |
| Sr | 19.40 | 95.60 | 60.60 | 91.80 | 208.50 | 20.90 | 33.60 |
| Nd | 3.45 | 9.07 | 2.42 | 4.20 | 2.25 | 5.44 | 21.72 |
| Zr | 20.00 | 74.00 | 14.00 | 24.00 | 14.00 | 45.00 | 65.00 |
| Hf | 0.80 | 2.30 | 0.51 | 0.84 | 0.49 | 1.50 | 2.72 |
| Sm | 1.14 | 3.13 | 0.76 | 1.17 | 0.72 | 1.33 | 6.45 |
| Eu | 0.48 | 1.15 | 0.33 | 0.43 | 0.30 | 0.39 | 1.84 |
| Gd | 1.27 | 3.45 | 0.83 | 1.11 | 0.75 | 1.10 | 5.44 |
| Tb | 0.28 | 0.68 | 0.17 | 0.22 | 0.16 | 0.19 | 0.90 |
| Dy | 2.00 | 4.39 | 1.30 | 1.53 | 1.21 | 1.18 | 5.11 |
| Ho | 0.48 | 1.00 | 0.32 | 0.37 | 0.29 | 0.26 | 1.10 |
| Y | 11.00 | 25.00 | 6.00 | 9.00 | 6.00 | 6.00 | 32.00 |
| Er | 1.29 | 2.51 | 0.83 | 0.97 | 0.78 | 0.68 | 2.80 |
| Yb | 1.39 | 2.48 | 0.90 | 1.07 | 0.87 | 0.74 | 2.94 |
| Lu | 0.22 | 0.38 | 0.15 | 0.17 | 0.14 | 0.12 | 0.47 |

(1) mafic amphibolite altered by 1- the phlogopite-rich matrix and 2- late chlorite-rich matrix
b.d.l.: below detection limit

Table 4:: LA-ICP-MS geochemistry for representative Ca-amphibole in amphibolite dikes.

| Sample | M8a1 | M8a1 | M8a1 | M8a2 | M8a2 | M8a2 |
|--------------------------------|---------------|---------------|---------------|---------------|---------------|---------------|
| Mineral | prg | prg | prg | mhb | mhb | mhb |
| SiO ₂ (wt%) | 43.12 | 45.46 | 43.98 | 48.14 | 43.82 | 46.32 |
| TiO ₂ | 0.30 | 0.09 | 0.21 | 0.53 | 0.77 | 0.85 |
| Al ₂ O ₃ | 13.99 | 12.79 | 14.11 | 10.48 | 13.13 | 11.82 |
| Cr ₂ O ₃ | 0.07 | 0.15 | 0.24 | 0.17 | 0.10 | 0.07 |
| FeO | 8.97 | 6.53 | 7.85 | 7.70 | 9.29 | 8.76 |
| MnO | 0.16 | 0.07 | 0.10 | 0.13 | 0.08 | 0.06 |
| MgO | 15.84 | 17.43 | 16.28 | 16.80 | 15.24 | 16.10 |
| CaO | 11.83 | 12.37 | 12.12 | 11.64 | 11.89 | 12.01 |
| Na ₂ O | 2.79 | 2.39 | 2.44 | 1.82 | 2.48 | 2.14 |
| K ₂ O | 0.07 | 0.04 | 0.05 | 0.04 | 0.07 | 0.07 |
| Sum | 97.13 | 97.31 | 97.37 | 97.47 | 96.87 | 98.20 |
| Cs (ppm) | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | 0.24 | 0.07 |
| Rb | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | 3.72 | <i>b.d.l.</i> |
| Ba | 2.18 | 2.05 | 2.77 | 2.68 | 5.21 | 3.79 |
| Nb | 0.36 | 0.31 | 0.40 | 0.68 | 0.65 | 0.87 |
| Ta | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | 0.11 | 0.15 |
| La | 2.23 | 1.90 | 2.26 | 2.57 | 2.67 | 2.81 |
| Ce | 6.31 | 5.76 | 6.63 | 7.95 | 7.09 | 9.64 |
| Pb | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> | <i>b.d.l.</i> |
| Pr | 0.95 | 0.92 | 1.23 | 1.60 | 1.21 | 1.67 |
| Sr | 73.11 | 68.75 | 67.36 | 41.47 | 46.78 | 48.35 |
| Nd | 4.05 | 4.59 | 5.67 | 7.81 | 8.66 | 10.30 |
| Zr | 24.49 | 23.12 | 30.51 | 48.31 | 38.45 | 56.43 |
| Hf | 1.13 | 0.85 | 1.04 | 1.95 | 1.62 | 2.08 |
| Sm | 1.50 | <i>b.d.l.</i> | 1.64 | 3.23 | 2.96 | 2.74 |
| Eu | 0.87 | 0.76 | 0.81 | 1.08 | 1.09 | 1.46 |
| Gd | <i>b.d.l.</i> | 1.41 | 3.34 | 3.10 | 3.62 | 6.02 |
| Tb | 0.33 | 0.32 | 0.62 | 0.82 | 0.69 | 1.20 |
| Dy | 2.76 | 2.86 | 3.86 | 5.97 | 6.50 | 8.42 |
| Ho | 0.67 | 0.68 | 0.93 | 1.31 | 1.18 | 2.03 |
| Er | 2.36 | 1.60 | 2.80 | 4.20 | 3.43 | 5.65 |
| Yb | 2.26 | 1.44 | 2.53 | 4.72 | 4.13 | 5.58 |
| Lu | 0.22 | 0.36 | 0.49 | 0.62 | 0.67 | 0.76 |

See appendix for spider diagrams.

Abbreviations after Whitney and Evans (2010)

Table 5: $^{40}\text{Ar}/^{39}\text{Ar}$ age performed on amphiboles from mafic dikes cross-cutting the peridotite of the New Caledonia ophiolite.

| Sample | Lithology | Location | Total fusion age | Plateau age | % ^{39}Ar | Intercept age | ($^{40}\text{Ar}/^{36}\text{Ar}$) _i | MSWD | Preferred age | 38Ar/37Ar | 38Ar/39Ar |
|------------|------------------------|---------------|------------------|----------------|--------------------|---------------|--|-----------|---------------|-----------|-----------|
| C10 (1) | MS amphibolite | Thio | 59.12 ± 2.24 | 56.21 ± 2.14 | 95.6 | <i>nd</i> | <i>nd</i> | <i>nd</i> | 56.21 ± 2.14 | <i>nd</i> | <i>nd</i> |
| C10 (2) | MS amphibolite | Thio | 68.23 ± 2.97 | 56.85 ± 4.79 | 96.0 | <i>nd</i> | <i>nd</i> | <i>nd</i> | | <i>nd</i> | <i>nd</i> |
| SNC01 (1) | Pegmatite dike | Massif du Sud | 53.67 ± 3.14 | 60.20 ± 4.62 | 92.5 | <i>nd</i> | <i>nd</i> | <i>nd</i> | 60.20 ± 4.62 | 0.011 | 0.41 |
| SNC01 (2) | Pegmatite dike | Massif du Sud | 87.66 ± 5.94 | 62.87 ± 5.54 | 61.8 | <i>nd</i> | <i>nd</i> | <i>nd</i> | | 0.01 | 0.36 |
| SNC03 (1) | Gabbro | Massif du Sud | 51.13 ± 4.49 | 51.25 ± 3.12 | 81.5 | 51.32 ± 4.18 | 295.7 ± 21.1 | 0.65 | 51.25 ± 3.12 | 0.013 | 0.53 |
| SNC04 (1) | Gabbro | Massif du Sud | 47.59 ± 4.36 | 49.56 ± 3.98 | 97.8 | 52.22 ± 5.34 | 287.5 ± 10.5 | 0.93 | 52.22 ± 5.34 | 0.007 | 0.27 |
| M3b (1) | ol-free amphibolite | Plum beach | 53.64 ± 4.27 | 53.35 ± 3.39 | 78.0 | 54.10 ± 3.40 | 293.6 ± 5.0 | 1.25 | 54.10 ± 3.40 | 0.005 | 0.18 |
| M4d (1) | ol-free amphibolite | Plum beach | 238.54 ± 7.50 | / | / | / | / | / | | 0.02 | 1.72 |
| M4d (2) | ol-free amphibolite | Plum beach | 73.10 ± 5.49 | 68.67 ± 6.37 | 97.4 | 67.59 ± 5.06 | 303.7 ± 5.0 | 1.02 | 67.59 ± 5.06 | 0.02 | 1.73 |
| M8f (1) | ol-free amphibolite | Plum beach | 58.34 ± 3.61 | 56.29 ± 3.28 | 95.8 | 52.86 ± 5.34 | 318.6 ± 20.6 | 1.79 | 56.23 ± 4.42 | <i>nd</i> | <i>nd</i> |
| M8f (2) | ol-free amphibolite | Plum beach | 62.08 ± 2.61 | 60.82 ± 3.06 | 92.2 | 60.25 ± 2.84 | 299.6 ± 4.2 | 1.77 | | 0.004 | 0.11 |
| M8f (3) | ol-free amphibolite | Plum beach | 93.78 ± 3.08 | 49.34 ± 4.47 | 49.31 | 56.46 ± 7.40 | 282.5 ± 12.1 | 0.23 | | 0.004 | 0.07 |
| M8f (4) | ol-free amphibolite | Plum beach | 53.43 ± 2.85 | 54.86 ± 2.04 | 99.5 | 55.36 ± 2.11 | 293.1 ± 2.9 | 0.38 | | 0.003 | 0.08 |
| M8a-Z1 (1) | ol-bearing amphibolite | Plum beach | 107.69 ± 7.09 | 107.94 ± 10.80 | 67.7 | 92.81 ± 13.43 | 349.1 ± 19.0 | 1.56 | | | 1.71 |
| | | | | | | | | | 92.50 ± | | |
| M8a-Z1 (2) | ol-bearing amphibolite | Plum beach | 125.24 ± 4.51 | 111.22 ± 6.93 | 70.8 | 92.50 ± 11.65 | 351.7 ± 3.7 | 1.28 | 11.65 | 0.017 | 1.68 |
| M8a-Z2 (1) | ol-free amphibolite | Plum beach | 76.61 ± 7.31 | 72.54 ± 5.83 | 86.1 | 73.85 ± 6.74 | 297.0 ± 9.6 | 1.55 | 73.85 ± 6.74 | 0.024 | 1.87 |
| M15 (2) | ol-bearing amphibolite | Plum beach | 81.80 ± 4.20 | / | / | / | / | / | | 0.019 | 1.59 |
| M15 (3) | ol-bearing amphibolite | Plum beach | 89.97 ± 8.42 | 85.94 ± 7.50 | 87.9 | 86.54 ± 7.58 | 296.1 ± 1.8 | 2.09 | | 0.017 | 1.28 |
| M15 (4) | ol-bearing amphibolite | Plum beach | 167.09 ± 21.48 | / | / | / | / | / | / | 0.013 | 1.74 |

Preferred age is determined after plateau age, intercept age or a mean of different analysis for a same sample when the plateau age is not significant.

MS: metamorphic sole; MSWD: mean square weighted deviation.