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New geochemical and age constraints ($^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb) on forearc intrusive rocks from New Caledonia Ophiolite (SW Pacific): diversity of melts generated at hot subduction inception

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1 **New geochemical and age constraints ($^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb) on forearc intrusive rocks from New**
2 **Caledonia Ophiolite (SW Pacific): diversity of melts generated at hot subduction inception.**

3

4 **New Caledonia Eocene forearc magmatism**

5

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25

26 **Abstract**

27

28 New Caledonia Ophiolite is crosscut by coarse to medium grained pyroxenites and hornblende
29 gabbros/diorites dykes intruded between 55.5Ma and 50Ma (U-Pb zircon and $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende),
30 while finer-grained dolerites of tholeiitic affinity are younger (50-47 Ma). Production of hornblende-
31 gabbros/diorites was modelled by moderate degree (20-40%) of partial melting of the HT
32 amphibolites of the metamorphic sole. End-member compositions, hornblendites and anorthosites,
33 resulted from solid-state phase segregation of crystal mushes within tectonically active magmatic
34 conduits. Cascade reactions of slab melts with mantle wedge peridotites successively formed
35 clinoenstatite-boninite magmas, which fed gabbronorite cumulate lenses at the mantle-crust

36 transition, in turn clinoenstatite-boninite melts reacted with peridotites to form websterites. The
37 youngest magmas of tholeiitic affinity, appeared about 6 Ma after subduction inception when the
38 cooler subducting slab plunged more steeply. Incipient slab retreat allowed corner flow, triggering
39 low pressure hydrous melting of the uplifted asthenosphere . The early stages of forearc magmatism
40 were closely associated with transcurrent shear zones, which recorded oblique subduction inception.
41 The lower Eocene tectonic and magmatic features of the New Caledonia ophiolite witness the
42 existence of a north- or northeast-dipping hot (forced) subduction zone in the Southwest Pacific,
43 which appears notably distinct from the slightly younger west-dipping Izu-Bonin-Marianna cold
44 (spontaneous) subduction system.

45

46 **Keywords** : New Caledonia Ophiolite, forearc magmatism, subduction inception, mantle shear zones,
47 ⁴⁰Ar/³⁹Ar, U-Pb zircon

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52 Introduction

53 Veins and dykes are commonly found in mantle sections of Supra Subduction Zone (SSZ)
54 ophiolites; they typically are a few centimeters to several meters thick, either with sharp, diffuse,
55 straight or deformed boundaries, and are diversely oriented with respect to the high-temperature
56 peridotite layering. SSZ veins and dykes may have a purely magmatic and partly or totally
57 metasomatic origin. They record a relatively wide range of temperatures and can be emplaced at
58 various stages of ophiolite cooling. Their compositional diversity allows investigation of the various
59 processes that are active in forearc domains of subduction zones (for a review see Furnes and Dilek,
60 2017). Such investigation is difficult in modern active oceanic forearcs, as these can only be accessed
61 by deep sea drilling. In contrast, SSZ ophiolites that represent obducted forearc lithosphere (e.g.,
62 Stern et al., 2008) are easily accessible targets in spite of erosion and tectonic dismantlement.

63 Pyroxenites are the most common lithologies preserved in the SSZ ophiolite record. Origin of
64 pyroxenite dykes has been attributed to either local harzburgite re-melting (Nicolas, 1989), hydrous
65 melting of depleted mantle wedge peridotites (Xiong et al. 2014), or peridotite-melt interaction
66 (Berly et al., 2006; Rogkala et al., 2017; Le Roux and Liang, 2019) occurring in the early forearc
67 evolution. Owing to their low incompatible elements content, SSZ pyroxenites are generally
68 unsuitable for radio-isotopic dating, and thus rarely constrained temporally. However, some SSZ
69 pyroxenites derived from hydrous melts contain small amounts of magmatic amphibole, which can
70 be dated using the $^{40}\text{Ar}/^{39}\text{Ar}$ method.

71 Hornblende-rich gabbro/diorite dykes are also common in SSZ ophiolites (Yu et al., 2022). In the
72 New Caledonia (NC) ophiolite, their compositions vary from hornblendite with less than 5%
73 plagioclase, to hornblende gabbro/diorite, and to “anorthosite” (less than 5% hornblende). Unlike
74 other SSZ lithologies, they are sufficiently rich in incompatible elements to develop minerals suitable
75 for geochronology ($^{40}\text{Ar}/^{39}\text{Ar}$ and/or U-Pb). Although of minor importance, biotite-
76 granites/pegmatites occur locally. Basalt dykes commonly occur in SSZ ophiolites; they may be
77 generated early in the forearc history during subduction initiation (Reagan et al., 2010, Hickey-Vargas
78 et al., 2018, Whattam and Stern, 2011; Furnes and Dilek, 2017; Shervais et al., 2019), or later during
79 subduction evolution (e.g., Lallemand and Arcay, 2021 and references therein).

80 This article presents new structural, geochemical and geochronological data obtained for a set of
81 dykes from the NC ophiolite. These data provide strong constraints on the timing and evolution of
82 forearc magmatism and its relationship to tectonics in the specific case of the inception of a hot and
83 oblique subduction zone. Our new $^{40}\text{Ar}/^{39}\text{Ar}$ amphibole ages from pyroxenite dykes crosscutting the
84 NNE-SSW trending Bogota shear zone and from a synkinematic pyroxenite intrusion associated with
85 NW-SE trending shear zones near Noumea (Plum) provide the opportunity to obtain information not

86 only on the timing of pyroxenite intrusion, but also on the age of the shearing deformation that
87 affected the host peridotite through the application of the $^{40}\text{Ar}/^{39}\text{Ar}$ method. The widespread
88 occurrence and diversity of magmatic dykes offer new geochronological and geochemical insights
89 into forearc evolution, and more specifically, of a young subduction zone that likely originated at or
90 near a spreading ridge (e.g., Ulrich et al., 2010; Cluzel et al., 2012b).

91

92 **Material and methods**

93 This study is based on surveys of the Peridotite Nappe conducted from the early 1990s to 2020.
94 Some 125 samples were analyzed for major and trace elements, 23 samples for Nd-Sr isotopes, 30
95 were selected for U-Pb dating, and 7 for $^{40}\text{Ar}/^{39}\text{Ar}$ dating.

96

97 ***$^{40}\text{Ar}/^{39}\text{Ar}$ dating at Oregon State University.*** The $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data are organized to comply
98 with FAIR data reporting norms (see for instance Schaen et al., 2020). An Excel workbook is provided
99 with data formatted into individual worksheets with isotope ratio and raw intensity data. Hornblende
100 was separated by standard mineral separation procedures and the samples were irradiated in two
101 packages. NM-312 was at the Oregon State University for 14 hours in the CLICT position, whereas
102 NM-322 was at the USGS reactor in Denver, CO and was for 24 hours. Fish Canyon sanidine was used
103 as the neutron flux monitor and is assigned an age of 28.201 Ma (Kuiper et al., 2008) using a ^{40}K
104 decay constant of $5.463\text{e}^{-10}/\text{a}$ (Min et al., 2000).

105 After irradiation, the samples were step-heated with a diode laser with a heating time of 45
106 seconds per step, followed by 60 seconds of gas clean up with a SAES GP-50 getter operated at 1.6 A.
107 Gas was also exposed to a cold finger operated at $\sim -140^\circ\text{C}$. Argon isotopes were measured using a
108 Thermo-Fisher Scientific Helix MC-plus multi-collector mass spectrometer. Isotopes ^{40}Ar , ^{39}Ar , ^{38}Ar ,
109 and ^{37}Ar were measured on Faraday collectors with ^{40}Ar and ^{37}Ar using a 10^{12} Ohm resistor, ^{39}Ar using
110 a 10^{13} Ohm resistor, and ^{38}Ar using a 10^{14} Ohm resistor. ^{36}Ar was measured on a compact discrete
111 dynode (CDD) ion counter that has a dead time of 20 ns. CaF_2 and K-glass were included in both
112 irradiations to ensure accurate correction for interfering reactions.

113 Calibration gases of air and a gas mixture enriched in radiogenic ^{40}Ar along with ^{39}Ar were
114 analyzed interspersed with the samples to monitor instrument drift and determine detector
115 intercalibration factors. Extraction line blanks were interspersed with sample measurements and
116 are reported in the intensity data worksheet. All data collection was conducted with the in-house
117 Pychron software, and data reduction utilized MassSpec version 7.875. The plateau ages are derived
118 from the weighted mean of the chosen steps with the weighting factor being the inverse variance
119 (Taylor, 1982) and the error is the square root of the sum of $1/\sigma^2$ values. Isochron regressions

120 followed the methods of York (1968). Age errors are reported at 2σ and include the J-factor error and
121 irradiation correction factor uncertainties.

122

123 ***⁴⁰Ar/³⁹Ar dating at Curtin University.*** We selected two fresh samples of hornblendite (VOH 1) and
124 dolerite (SIR 8) for ⁴⁰Ar/³⁹Ar dating. All samples were crushed to centimeter scale using a hydraulic
125 press, after which fresh separates were further crushed to several hundred micrometer in size in a
126 ring mill. The crushates were sieved and washed with deionized water in a sonic bath. Groundmass
127 grains were handpicked from the dolerite sample using a binocular stereomicroscope. For sample
128 VOH1, we isolated amphibole crystals using a Frantz magnetic separator, and then we carefully hand-
129 picked the freshest and inclusion-free crystals under a binocular microscope. The samples were
130 further leached in diluted HF for one minute and then thoroughly rinsed with distilled water in an
131 ultrasonic cleaner.

132 Samples were loaded into large wells of one 1.9 cm diameter and 0.3 cm depth aluminum disc.
133 These wells were bracketed by small wells that included GA1550 biotite used as a neutron fluence
134 monitor for which an age of 99.738 ($\pm 0.1\%$; 1σ) was adopted (Renne et al., 2011). The discs were Cd-
135 shielded (to minimize undesirable nuclear interference reactions) and irradiated for 3 hours in the US
136 Geological Survey nuclear reactor (Denver, USA) in central position in 2011 and in 2014. The mean J-
137 values computed from standard grains within the small pits range from determined as the average
138 and standard deviation of J-values of the small wells for each irradiation disc is provided in table S6.
139 Mass discrimination was monitored using an automated air pipette relative to an air ratio of $298.56 \pm$
140 0.31 (Lee et al., 2006). The correction factors for interfering isotopes were (³⁹Ar/³⁷Ar)Ca = 7.30×10^{-4} (\pm
141 11%), (³⁶Ar/³⁷Ar)Ca = 2.82×10^{-4} ($\pm 1\%$) and (⁴⁰Ar/³⁹Ar)K = 6.76×10^{-4} ($\pm 32\%$). ⁴⁰Ar/³⁹Ar analyses were
142 carried out in the Western Australian Argon Isotope Facility, at Curtin University, Perth on both a
143 MAP 215-50 (sample VOH1) and ARGUS VI (sample SIR8) mass spectrometers.

144 Sample VOH1 (hornblendite) – a multi-grain aliquot of amphibole crystals was wrapped in low-
145 blank niobium foil and step-heated using a 110 W Spectron Laser Systems, with a continuous Nd-YAG
146 (IR; 1064 nm) laser rastered over the sample for 1 minute to ensure a homogenously distributed
147 temperature. The gas was purified in a stainless-steel extraction line using two SAES AP10 and one
148 GP50 getters. Ar isotopes were measured in static mode using a MAP 215-50 mass spectrometer
149 (resolution of ~ 500 ; sensitivity of 4×10^{-14} mol/V) with a Balzers SEV 217 electron multiplier using 9 to
150 10 cycles of peak-hopping. Data acquisition was performed with the Argus program written by M.O.
151 McWilliams and running within a Labview environment. Blanks were monitored every 3 to 4 steps.

152 Sample SIR8 (dolerite) – small groundmass aliquots were step-heated using a continuous 100 W
153 PhotonMachine© CO2 (IR, 10.6 μ m) laser fired on the aliquot material for 60 seconds. All standard

154 crystals were measured on the ARGUS VI and fused in a single step. The gas was purified in an extra
155 low-volume stainless steel extraction line of 240 cm³, set up to run with a single SAES AP10 getter. Ar
156 isotopes were measured in static mode using a low-volume (600 cm³) ARGUS VI mass spectrometer
157 from Thermo Fisher© (Jourdan et al., 2019) set with a permanent resolution of ~200. Measurements
158 were carried out in multi-collection mode using three Faraday cups equipped with three 1012 ohm
159 (masses 40; 38; and 37) and one 1013 ohm (mass 39) resistor amplifiers and a low background
160 compact discrete dynode (CDD) ion counter to measure mass 36. We measured the relative
161 abundance of each mass simultaneously during 10 cycles of peak-hopping and 16 seconds of
162 integration time for each mass. Detectors were calibrated to each other through air shot beam
163 signals. Blanks were analyzed for every three to four incremental heating steps.

164 The raw data (Table S6) were processed using the ArArCALC software (Koppers, 2002), and the
165 ages have been calculated using the decay constants recommended by Renne et al. (2011). All
166 analytical parameters and relative abundance values are provided in Table S6 and have been
167 corrected for blanks, mass discrimination and radioactive decay. Individual errors in Table S6 are
168 given at the 1 σ level. Criteria for the determination of a plateau are as follows: (1) plateaus must
169 include at least 70% of ³⁹Ar; (2) the plateau should be distributed over a minimum of 3 consecutive
170 steps agreeing at 95% confidence level and satisfying a probability of fit (P) of at least 0.05. The use
171 of these criteria means that perturbed spectra are not used for age determination. Plateau ages are
172 given at the 2 σ level and are calculated using the mean of all the plateau steps, each weighted by the
173 inverse variance of their individual analytical error.

174

175 ***U-Pb zircon dating at the University of Tasmania.*** Twenty representative samples were selected
176 from a set of some 120 dyke samples collected throughout the Peridotite Nappe and dated by LA-
177 ICPMS zircon geochronology at the University of Tasmania.

178 Approximately 100 g of rock was repeatedly sieved and crushed in a Cr-steel ring mill to a grain
179 size <400 μ m. Non-magnetic heavy minerals were then separated using a gold pan and a Fe-B-Nd
180 hand magnet. The zircons were hand-picked from the heavy mineral concentrate under the
181 microscope in cross-polarised transmitted light. The selected crystals were placed on double sided
182 sticky tape and epoxy glue was then poured into a 2.5 cm diameter mould on top of the zircons. The
183 mount was dried for 12 hours and polished using clean sandpaper and a clean polishing lap. The
184 samples were then washed in distilled water in an ultrasonic bath.

185 The analyses were performed on Agilent 7500cs quadrupole ICPMS instruments between 2004
186 and 2015. Data collected prior to 2010 used the New Wave UP193 nm solid state laser equipped with
187 a custom-made laser ablation cell (see Sack et al. 2011 for details). Data collected after 2010 used a
188 193 nm Coherent Ar-F gas laser and the Resonetics M50 ablation cell. The downhole fractionation,

189 instrument drift and mass bias correction factors for Pb/U ratios on zircons were calculated using 2
190 analyses on the primary (91500 standard of Wiendenbeck et al. 1995) and 1 analysis on each of the
191 secondary standard zircons (Temora standard of Black et al. 2003 & JG1 of Jackson et al. 2004)
192 analyzed at the beginning of the session and every 12 unknown zircons (roughly every 1/2 hour)
193 using the same spot size and conditions as used on the samples. Additional secondary standards (The
194 Mud Tank Zircon of Black & Gulson, 1978) were also analysed. The correction factor for the
195 $^{207}\text{Pb}/^{206}\text{Pb}$ ratio was calculated using 3 large spot of NIST610 analysed at the beginning and end of
196 the day and corrected using the values recommended by Baker et al. (2004).

197 Each analysis on the zircons began with a 30 second blank gas measurement followed by a further
198 30 seconds of analysis time when the laser was switched on. Zircons were sampled on 32 micron
199 spots using the laser at 5 Hz and a density of approximately 1.5 J/cm^2 . A flow of He carrier gas carried
200 particles ablated by the laser out of the chamber to be mixed with Ar gas and carried to the plasma
201 torch. Elements measured include ^{49}Ti , ^{96}Zr , ^{146}Nd , ^{178}Hf , ^{202}Hg , ^{204}Pb , ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th and ^{238}U
202 with each element being measured sequentially every 0.16 s with longer counting time on the Pb
203 isotopes compared to the other elements. The data reduction used was based on the method
204 outlined in detail in Meffre et al. (2008) similar to that outlined in Black et al. (2004) and Paton et al
205 (2010). Element abundances in zircons were calculated using the method outlined by Kosler (2001)
206 using Zr as the internal standard element, assuming stoichiometric proportions and using the 91500
207 to standard correct for mass bias.

208

209 ***U-Pb dating at the University of Queensland.*** Analyses were performed using a Thermo Fisher iCAP
210 RQ quadrupole inductively coupled plasma mass spectrometer (ICP-MS) equipped with an ASI
211 RESOLUTION SE 193 nm laser at Centre for Geoanalytical Mass Spectrometry (The University of
212 Queensland). Zircon data were collected with a $30 \mu\text{m}$ spot size in 30 cycles of 1s at a repetition rate
213 of 7Hz and a fluence of 3J/cm^3 . U-Pb age determinations for zircons were undertaken using the
214 91500 zircon standard (Wiedenbeck et al., 1995) for background and instrumental bias (Jackson et
215 al., 2004). TEMORA2 (Black et al., 2003) or GJ1 zircons (Jackson et al., 2004) were used as secondary
216 standards to monitor precision and accuracy. Data produced indicate precision and accuracy better
217 than 0.5% for U-Pb age measurements for both analytical sessions. The raw data were processed
218 using the Lolite software (Paton et al., 2011) and later normalized using Zr as the internal standard.

219

220 ***Whole rock and mineral data.*** Whole-rock major (ICP-OES) and trace-elements (ICP-MS) analyses of
221 hornblende-rich dikes and dolerites were carried out at the CNRS-SARM-CRPG of Nancy (France);
222 analytical procedures, errors and detection limits may be found at <https://sarm.cnrs.fr/index.html> .

223 Major element mineral analyses of pyroxenites were carried out at Department of Chemistry, Life
224 Sciences and Environmental Sustainability in Parma (Italy) using a JEOL-6400 electron microprobe
225 equipped with a LINK-ISIS energy dispersive micro analytical system. The electron beam was
226 produced at an accelerating voltage of 15 kV and probe current of 0.25 nA. Both natural minerals and
227 synthetic compounds were used as standards.

228

229 **Geologic setting**

230 The main island of New Caledonia (named Grande Terre) is the emerged northern part of a
231 continent-size area of thinned continental crust termed Zealandia (Luyendyk, 1995; Mortimer et al.,
232 2017), which was rifted from the south Gondwana margin (Antarctica and Australia) during the Late
233 Cretaceous (Hayes and Ringis, 1973). In the northern part of Zealandia, marginal rifting isolated the
234 Norfolk Ridge, a narrow slice of thinned continental crust bound to the west by the New Caledonia
235 Basin and to the east by the proto-Loyalty Basin. New Caledonia geology records a collage of
236 Permian-Mesozoic island-arc related terranes (Meffre, 1995; Aitchison et al., 1995a; Maurizot et al.,
237 2020a) accreted during the late Early Cretaceous and unconformably overlain by a Late Cretaceous-
238 Eocene sedimentary cover (Maurizot et al., 2020b). In turn, pre-Oligocene rocks were tectonically
239 overlain by the obducted oceanic lithosphere of the Loyalty Basin (for details see Cluzel et al., 2012a
240 and Maurizot et al., 2020c).

241 During the Eocene, the proto-Loyalty Basin was the site of north or northeast dipping intra-
242 oceanic subduction, which resulted in off-scraping of the oceanic lower plate and forearc accretion
243 (Cluzel et al., 2001; Cluzel et al., 2017). Subduction erosion resulted in the formation of serpentinite
244 melange (Cluzel, 2020), while high-pressure low-temperature (HP-LT) metamorphism developed with
245 a northeastward increasing grade (Clarke et al., 1997; Spandler et al., 2005; Vitale Brovarone and
246 Agard, 2013; Maurizot et al., 2020c and references therein). Meanwhile, in-sequence southwest-
247 directed thrusting affected the sedimentary cover of the Norfolk Ridge and generated coeval
248 turbidite basins. Oblique subduction eventually led to arc-continent collision and forearc obduction
249 when the northern tip of the continental Norfolk Ridge blocked the subduction (Aitchison et al.,
250 1995a) at the Eocene-Oligocene boundary (Cluzel et al., 1998).

251 As a result of Eocene obduction and subsequent erosion, about 5,500 km² of allochthonous upper
252 mantle rocks (Avias, 1967) of the NC Ophiolite, also known as Peridotite Nappe (Paris, 1981), are
253 exposed in the Grande Terre, Belep Island and Isle of Pines (Fig. 1). The Peridotite Nappe is rooted in
254 the Loyalty Basin (Collot et al., 1987) and consists of highly depleted harzburgites (e.g., Nicolas and
255 Prinzhofer, 1983; Marchesi et al., 2009; Pirard et al., 2013; Secchiari et al., 2020), dunites and minor
256 lherzolites (e.g., Ulrich et al., 2010; Secchiari et al., 2016). The ophiolite sheet is up to 2,500 m thick

257 and is underlain by a serpentinite sole, 20 to 200 m thick. Locally, lenses of high-temperature
258 amphibolites are pinched between the serpentinites and the underlying Poya Terrane (Cluzel et al.,
259 2012b). Near the top of the ultramafic rocks a ~300-500 m thick dunitic transition zone (DTZ) consists
260 of replacive/reactive dunites, which resulted from interaction of olivine-saturated melts with the
261 underlying harzburgites (Marchesi et al., 2009; Pirard et al., 2013). Some wehrlites on top of the DTZ
262 may have a replacive origin as well (Nicolas and Prinzhofer, 1983; Marchesi et al., 2009; Pirard et al.,
263 2013). The DTZ is in turn overlain by mafic-ultramafic cumulate lenses with a distribution, from base
264 to top, of dunites-pyroxenites, wehrlites and websterites-gabbronorites (e.g., Pirard et al., 2013;
265 Secchiari et al., 2018) (Suppl. Fig. S1). The remarkable absence of an upper section of oceanic crust
266 (volcanic rocks and sediments) suggests either deep erosion or removal by intra-crustal tectonic
267 detachment (Cluzel et al., 2001). Notably, the absolute age of the lower crustal intrusive sequence is
268 unknown at present, mainly due to the lack of suitable lithologies for radio-isotopic dating. An upper
269 time constraint of ca. 60 Ma is however provided by Nd model ages of the layered gabbronorites
270 (Secchiari et al., 2018), consistent with their formation during the Early Eocene (Cluzel et al. 2006).

271 North- or northeast-dipping subduction, which eventually led to obduction, originated within the
272 proto-Loyalty Basin to the east of New Caledonia, which evolved from marginal basin (Late
273 Cretaceous to Paleocene) to forearc setting (Eocene) (Aitchison et al., 1995a, 1995b; Cluzel et al.,
274 2001). Intra-oceanic subduction started shortly before 56 Ma, probably close to a spreading ridge as
275 recorded by U-Pb zircon (55.9 ± 0.8 Ma) and $^{40}\text{Ar}/^{39}\text{Ar}$ amphibole cooling age (55.8 ± 1.7 Ma) of HT
276 amphibolites from the metamorphic sole (Cluzel et al., 2012b; Soret et al., 2016). Obduction occurred
277 after 34 Ma, as constrained by the youngest pre-obduction marine sediments (Cluzel et al., 1998),
278 and before 24.5 Ma, as indicated by post-obduction magmatism (Paquette and Cluzel, 2007).
279 Paleomagnetic data from laterite ferricrete developed upon peridotites suggest that the ophiolite
280 was exposed and subject to weathering for some time before 25 Ma (Sevin et al., 2012).

281 A variety of dykes displaying supra-subduction geochemical signatures (e.g., Nb-Ta and Ti negative
282 anomalies) crosscut the ophiolite at all levels. These dykes were intruded during the lower Eocene
283 (55.5 - 50 Ma; U-Pb zircon) (Cluzel et al., 2006) shortly after subduction inception. They are also found
284 dismembered in the porphyroclastic serpentinite sole of the Peridotite Nappe and tectonically mixed
285 with elements of the mafic Poya Terrane, in blueschist and eclogite-facies serpentinite melanges of
286 the Eocene HP-LT belt (Cluzel et al., 2001; Spandler et al., 2005; Cluzel, 2020).

287 Several high-temperature shear zones crosscut NC ophiolite: the Bogota Transform Fault oriented
288 N20°E (Prinzhofer and Nicolas, 1980; Titus et al., 2011), the Bélep Shear Zone oriented N150°E
289 (Nicolas, 1989; Sécher, 1981; Titus et al., 2011), and the Humboldt Corridor trending N165°E (Ferré et
290 al., 2004; Vogt and Podvin, 1983) (Fig. 1). All these high-temperature shear zones are characterized
291 by steeply dipping foliation and horizontal stretching lineation. No time constraints related to their

292 activity are currently available so that their correlation with either the Late Cretaceous marginal
293 basin stage or the Eocene fore-arc stage remains speculative.

294

295

296 **Tectonics, geochemistry and geochronology of lower Eocene dykes**

297 ***General features***

298 The majority of dykes that cross-cut the Peridotite Nappe are typically coarse to medium-grained,
299 suggesting they were intruded into relatively hot host rocks. They are found throughout the
300 Peridotite Nappe, regardless of the structural level and nature of the ultramafic host rock
301 (harzburgite, dunite or lherzolite), but they appear more frequent near its base. The dykes are well
302 preserved in the main body of the ultramafic allochthon but are severely disrupted in the
303 serpentinite sole. Blocks of similar supra-subduction magmatic rocks with 55-50 Ma U-Pb zircon ages
304 are found in serpentinite melanges of the HP-LT belt of northern New Caledonia as a result of
305 subduction erosion (Cluzel, 2020).

306 Widespread coarse-grained felsic dykes suggest relatively slow cooling. In addition, pegmatitic
307 and sub-solvus textures and local formation of retrograde mineral associations suggest the
308 involvement of hydrous fluids during the late- and post-magmatic evolution. Some dykes underwent
309 internal ductile deformation (mineral boudinage, mineral preferred orientation, flow folding or
310 microfolding, mylonitisation), which is not observed in the peridotite host. A compositional layering
311 parallel to dyke walls is locally developed, thus suggesting high-temperature shearing and magma
312 injection in active faults. Two-step intrusion is recorded by some dykes, which display a foliated,
313 earlier, outer part while the rest of the dyke displays granular texture. Shearing therefore only
314 occurred during the early stage of dyke emplacement (Cluzel, 2020) (see outcrop pictures in online
315 appendix, suppl. Figs. S2 and S4).

316 Due to the large spacing, crosscutting relationships are rarely observed; however, dyke textures
317 allow for establishing a rough chronology. Some dykes display ductile deformation (e.g. Poum Massif,
318 Bogota Peninsula; Titus et al., 2011), thus indicating intrusion within a still hot peridotite. Although
319 coarse-grained, most felsic dykes were emplaced in colder host rocks and developed anthophyllite-
320 chlorite reaction rims or no rims at all. Dolerite/basalt dykes tend to be more common in the upper
321 level of the pile, they show fine-grained texture and locally chilled margins indicating intrusion in
322 cooler host rocks.

323 Dyke orientations do not fit a simple tension crack model since shallow dipping sills are
324 widespread and two or more sets of upright dykes commonly coexist. At island scale, the dykes
325 roughly cluster at *ca.* N130°E, N20°E, and N70°E; i.e., parallel or perpendicular to the main structural
326 lineaments of the Peridotite Nappe (Cluzel et al., 2021). Magma was injected through joints that
327 likely developed during the cooling of peridotite. For example, dolerite dykes of the Bogota
328 Peninsula, in spite of some dispersion, cluster at *ca.* N20°E (Titus et al., 2011), i.e., roughly parallel to
329 the Bogota Shear Zone (see below).

330 Sheared dykes are predominantly found near the base of Peridotite Nappe in the south of the
331 island (Plum, Ouen Island, Casy Islet; Fig. 1). They also occur, although less frequently, in northern
332 massifs. Dextral motion is recorded by WNW-ESE trending sheared dykes while rare NNW-SSE
333 trending dykes display sinistral motion, attesting NW-SE directed shortening and NE-SW stretching
334 directions (in the present orientation of the island; Cluzel et al., 2021).

335 Based on field and major elements classification, dyke compositions vary from ultramafic
336 (pyroxenite) to mafic (gabbro, diorite, hornblendite, dolerite, boninite), and felsic (leucogabbro,
337 leucodiorite, granite) (Fig. 2). Post-magmatic alteration is evidenced in some rocks by the variability
338 of mobile elements and occurrence of secondary mineral phases (tremolite, chlorite, etc). Whole-
339 rock major and trace element analyses of all dyke types (new and published data) are summarized in
340 the online appendix (Suppl. Table S1).

341 ***Plum pyroxenite intrusion***

342 The Plum composite intrusion in the Massif du Sud is located E-SE of Noumea near the base of
343 Peridotite Nappe (Fig.1). It includes a sub-circular pyroxenite body about 2 km wide and some
344 smaller elliptical bodies. Besides Upper Oligocene post-obduction granodiorites, it represents the
345 only intrusive body of such size in the Peridotite Nappe. It is intruded in strongly serpentinized
346 harzburgites and truncated by the basal thrust, thus providing evidence of subduction erosion
347 (Cluzel, 2020). The intrusion consists of roughly concentric envelopes with dunite, wehrlite, and
348 websterite inwards. The pyroxenite is intruded by hornblende-gabbro dykes oriented either parallel
349 to the curved shape of the intrusion, or perpendicularly crosscutting it (Fig. 3a). Some dykes are
350 sheared and mylonitized (Suppl. Figs. S2c, S2d), especially those oriented NW-SE (Cluzel et al., 2021).
351 The overall Z-shaped structure, with directions changing eastward from E-W to N-S and E-W, and
352 partitioned internal foliation (Suppl. Fig. S2b) suggests synkinematic intrusion in a dextral
353 transcurrent setting (Figs. 3b and 3c).

354 The main petrographic and geochemical features of the Plum pyroxenite were investigated in
355 detail by Secchiari et al. (2022b). In the following paragraph, the most essential characteristics of
356 these rocks will be summarized. The pyroxenite body consists of medium-grained websterites with
357 magmatic textures ranging from allotriomorphic to orthocumulitic. The websterites are mainly
358 composed of orthopyroxene (~ 30–75 vol %) and Al₂O₃ and TiO₂-poor clinopyroxene (~ 20–50 vol %),
359 coupled to variable amount of amphibole (~ 2-20 vol.%) and minor plagioclase (up to 9 vol.%) with
360 highly calcic composition (An₈₈–An₉₆). Gabbronorite dykes crosscut the pyroxenite body. In contrast
361 to the pyroxenite core, the dykes are characterized by variable grain size, often showing the
362 concomitant presence of mm-sized crystals (up to 1-2mm) surrounded by a fine-grained (~ 0.1–0.5
363 mm) crystal matrix. Mineral modes of the dykes are characterized by lower pyroxene contents (Opx ~
364 3–40 vol %; Cpx ~ 5–25 vol %) together with higher amphibole and plagioclase abundances (~ 20–30
365 vol % and ~ 10–55 vol %, respectively) compared to the main pyroxenite body. Pale green amphibole
366 mainly occurs as interstitial phase texturally associated with clinopyroxene. Abundant amphibole
367 blebs, with rounded or irregular margins, are commonly found within clinopyroxene, often in optical
368 continuity with the interstitial amphibole. The amphibole may also form rims on clinopyroxene.
369 According to Secchiari et al. (2022b), the amphibole crystallization may be ascribed to replacement
370 reactions involving the early-crystallized clinopyroxene in a crystal mush environment. Despite the
371 slightly different mineral modes, in situ major element investigation revealed similar compositional
372 features for the pyroxenite core and the crosscutting dykes. Major element pyroxene thermometry
373 (Brey and Kohler, 1990; Taylor, 1998) applied to Plum intrusive rocks provided a temperature interval
374 of 860-955 °C. This temperature range suggests some subsolidus cooling after magmatic
375 crystallization, as evidenced by mutual exsolutions in ortho- and clinopyroxene. The amphibole
376 geothermometer of Putirka (2016) also yields low temperatures (800-910 °C), which are consistent
377 with the late appearance of amphibole in the crystallisation sequence. The pyroxenite intrusion and
378 the associated dykes register similar equilibrium temperatures, thus implying a common cooling
379 history.

380 The Plum pyroxenites are depleted in incompatible elements and display negative Nb, Zr, Hf, Ta
381 and Ti anomalies on MORB-normalized trace elements diagrams (Fig. 4b). The associated dykes are
382 slightly more depleted and display Eu positive anomalies that suggest plagioclase segregation.
383 Geochemical modelling based on whole-rock and clinopyroxene trace element compositions (Figs. 4a
384 and 4b) indicates that putative melts in equilibrium with the Plum intrusives have a geochemical
385 signature strongly recalling those of the clinoenstatite-bearing (CE) boninites of Népoui (west coast
386 of NC; Secchiari et al., 2022b).

387 $^{40}\text{Ar}/^{39}\text{Ar}$ ages have been obtained for amphiboles from a websterite (Sample PL-AM7) and a
388 gabbro dyke (PL-AM10) using the step heating technique. According to the classification of Leake et
389 al. (1997) the amphiboles are magnesio-hornblende (Appendix Table S2) characterized by very low
390 K_2O contents (0.10-0.20 wt. %) and high Ca/K ratios (> 50, up to 187). Sample PL-AM7 has an initial
391 step near 58 Ma, older than the remaining steps that yield a flat segment with a well-defined plateau
392 age of 52.3 ± 0.6 Ma (Fig. 5a). PL-AM10 has an overall age pattern that is downward stepping with
393 the majority of the spectrum defined by ages between about 70 to 60 Ma (Fig. 5b). Isochron analysis
394 of the plateau steps for PL-AM7 yields an age equal to the plateau age of 52.5 ± 1.9 Ma along with an
395 initial $^{40}\text{Ar}/^{36}\text{Ar}$ equal to atmospheric argon (Fig. 5c). The isochron array for PL-AM10 yields an age of
396 53.9 ± 0.6 Ma and reveals an excess argon trapped initial $^{40}\text{Ar}/^{36}\text{Ar}$ of 337.1 ± 1.3 (Fig. 5d) (Appendix
397 Table S3). Based on cross-cutting relationships, the gabbro dyke (PL-AM10) is younger than the
398 websterite (PL-AM7) and the isochron dates are consistent with this as they are analytically
399 indistinguishable at the 2σ level. The $^{40}\text{Ar}/^{39}\text{Ar}$ ages are within the range of U-Pb zircon ages of
400 hornblende-bearing gabbro and diorite dykes (55.6-50.0 Ma) (Cluzel et al., 2006; this study).

401 ***Pyroxenite dykelets from the Bogota Peninsula.***

402 The Bogota Peninsula ultramafic rocks represent the NW extension of the Massif du Sud along the
403 northeastern coast of New Caledonia (Fig. 1). The orientation of the peridotite layering varies
404 throughout the Bogota Peninsula from E-W to N-S, dipping 35° to 65° ; a stretching lineation marked
405 by chromite streaks and elongated/boudinaged orthopyroxene grains with a pitch of about 45° in any
406 direction (Fig. 6) is also observed. The moderately dipping layering connects to, or is crosscut by high-
407 temperature dextral shear zones defined by steep mylonitic foliation bearing a shallow dipping
408 lineation (Suppl. Fig. S4c). The main shear zone was interpreted as an ancient oceanic transform and
409 referred to as Bogota Transform Fault (BTF) (Prinzhofer and Nicolas, 1980; Titus et al., 2011). BTF is
410 about 2 km wide around Ouassé Village, while narrower high-strain zones are found elsewhere (Fig.
411 6); a secondary shear zone was also described 20 km to the NW near Kouaoua Village (Titus et al.,
412 2011).

413 A number of mafic to ultramafic dykes are distributed throughout the peninsula (Fig. 6). Although
414 they are widely spaced, often preventing crosscutting relationships from being observed, the relative
415 timing from earliest to latest was suggested to be pyroxenite, plagioclase-bearing pyroxenite,
416 hornblende gabbro and dolerite/basalt (Prinzhofer and Nicolas, 1980). Such relative chronology is
417 consistent with decreasing grain size from pyroxenites to basalts, thus suggesting emplacement at
418 decreasing temperature/pressure conditions of the host peridotite.

419 Depending on their orientation and location within the high-strain domain, pyroxenite dykelets
420 are generally undeformed, post-dating ductile shear deformation of the peridotite. However, some

421 pyroxenite dykes display ductile shearing or boudinage when parallel to foliation in high strain
422 domains, or ptygmatic folding when perpendicular to it (Titus et al., 2011). Sheared or folded
423 pyroxenite dykes are therefore more or less contemporaneous with transcurrent motion. The
424 orientations of hornblende-gabbro dykes cluster around N5°E, N40°E and N130°E (Fig. 6), broadly
425 consistent with the orientation of dykes in the rest of the ophiolite (Cluzel et al., 2021), bearing no
426 direct relationship to BTF. Finally, undeformed basalt dykes display straight boundaries and cluster
427 around N30°E (Titus et al., 2011). They were seemingly channelized in already cold foliated rock.

428 In the Ouassé bay (Fig. 6), the pyroxenites occur as 5-15 cm thick dykelets that typically intersect
429 the host peridotite foliation at variable angles (Suppl. Fig. S2d). Rare concordant layers of
430 orthopyroxenites occur as well and some are boudinaged (Suppl. Fig. S2e). Pyroxenites mainly consist
431 of amphibole-bearing (Amp ~5-40 vol. %) websterites and minor orthopyroxenites containing ~5 vol.
432 % of olivine or clinopyroxene. Pyroxenite textures are highly variable, ranging from cumulitic to
433 porphyroclastic or granoblastic-polygonal (Ferrari, 2021). The enclosing harzburgites have textures
434 ranging from porphyroclastic to mylonitic. Their foliation is defined by aligned and/or stretched
435 orthopyroxene, with aspect ratio up to 12:1, and olivine porphyroclasts.

436 Two hornblendes from websterite samples were dated by the $^{40}\text{Ar}/^{39}\text{Ar}$ method (samples OU1 and
437 OU7B) and consist of orthopyroxene (~25-40 vol%), clinopyroxene (~25 vol%) and light brown
438 amphibole (30-40 vol %). The age spectra are concordant and simple to interpret with sample OU1
439 yielding a cooling age of 56.01 ± 0.19 Ma, and OU7B gave an analytically identical age of 56.05 ± 0.17
440 Ma (Fig. 7; Appendix Table S3), without any evidence for partial degassing or excess argon.

441 The sample OU1 displays a disequigranular texture with medium- to coarse-grained pyroxenes set
442 into a fine-grained granoblastic polygonal matrix composed of pyroxenes + amphibole and accessory
443 amounts of Ca-rich plagioclase and Cr-spinel. Conversely, sample OU7B has a coarse-grained
444 cumulitic texture with interstitial/poikilitic amphibole and accessory apatite and Fe-Ni sulfides.
445 Pyroxene geothermometry of Brey and Köhler (1990) and Taylor (1998) applied to pyroxenites
446 provided temperatures between 920-990°C. Consistent values (970 °C) were obtained using the
447 amphibole-plagioclase thermometer of Holland and Blundy (1994). Ca-in-orthopyroxene method
448 applied to the the porphyroclastic (920-970°C) and neoblastic (925-945°C) assemblage of the
449 adjacent harzburgites yielded comparable temperature estimates (Ferrari, 2021).

450 Recently, Xu et al. (2021) proposed that the websterites from the Ouassé bay resulted from
451 migration of boninitic melts into the host harzburgite. However, trace element composition of the
452 calculated melts in equilibrium with clinopyroxene of the amphibole websterites of this work are
453 distinctly enriched in incompatible elements with respect to boninitic melts and are similar to high-
454 Mg adakites (Ferrari, 2021). By contrast, the orthopyroxenite-forming liquids have a boninitic affinity
455 (Fig. 4c and 4d).

456 The dated amphiboles are analogous in chemistry to those dated from Plum pyroxenite intrusion
457 (Appendix, Table S2), i.e. they are magnesio-hornblendes with low TiO₂ (0.68-1.21 wt%) and K₂O
458 contents of 0.15 and 0.28 wt%. They have nearly identical ⁴⁰Ar/³⁹Ar cooling ages of 56.05 ± 0.17 Ma
459 and 56.01 ± 0.19 Ma respectively (Fig. 7) without any evidence for partial degassing or excess argon.

460 The two pyroxenite occurrences (Plum and Ouassé) partly differ by chemical composition,
461 structure, size and, to some extent, their timing. The Ouassé samples are measurably older than the
462 Plum samples at the 2σ confidence level. The slightly older one (Ouassé) is associated with the HT
463 Bogota shear zone. Here, the emplacement of different types of pyroxenite-forming liquids
464 (boninites and high-Mg adakites) was contemporaneous or closely followed ductile deformation. The
465 age of 56 Ma places the genesis of Ouassé pyroxenites and ductile shearing at the onset of
466 subduction. The apparently younger Plum pyroxenites occurred in much larger bodies and were
467 dominated by more depleted parental magmas, yet they were also associated with large-scale
468 transcurrent tectonics.

469

470 ***Nepoui clinoenstatite-boninite and boninite-series dykes***

471 Boninite outcrops from Plaine des Gaiacs, referred to as Nepoui clinoenstatite (CE)-boninite (Fig.
472 1), have been extensively studied since their discovery (Sameshima et al., 1983). Reassessment of
473 these outcrops has shown that Nepoui CE-boninite consists of blocks enclosed within the
474 serpentinite sole of Peridotite Nappe together with other dismembered magmatic rocks (Cluzel et al.,
475 2016). The mineralogical and geochemical features of the Nepoui boninite were already described in
476 detail (Sameshima et al., 1983; Cameron et al., 1983; Cameron, 1989; Ohnenstetter and Brown,
477 1996; Solovova et al., 2012) and are outlined here. The Nepoui boninite displays hyaloporphyritic
478 texture with some 50–60% dark glassy groundmass including poly-synthetically twinned
479 clinoenstatite, orthopyroxene (bronzite) and olivine microphenocrysts, and accessory clinopyroxene,
480 hornblende and chromite. Clinoenstatite laths, a diagnostic component of boninites, 1–4 mm long,
481 contain minute chromite grains and appear as decussate aggregates.

482 The Nepoui boninites have andesitic major elements composition (SiO₂ = 55.5 wt.%, Al₂O₃ = 10.0
483 wt.% on av. for 10 samples) with low CaO (3.8 wt.% on av.) and FeO contents (7.6 wt.%), low
484 CaO/Al₂O₃ ratio (0.4 on av.), very low K₂O (0.6 wt.%) and TiO₂ (0.2 wt%). In contrast, they display high
485 MgO (11–15 wt.%) and high contents in compatible elements contents (Cr = 486–1371 ppm, Ni =
486 190–1640 ppm) (Cluzel et al., 2016).

487 In contrast with the classical U-shaped REE pattern of boninites of the type locality (Bonin islands;
488 Crawford et al., 1998; Shervais et al., 2022), Nepoui CE-boninites display slightly concave “spoon-

489 shaped" REE patterns with significant LREE enrichment ($(La/Sm)_N=2.2$; Fig. 4a), which are similar to
490 that of the less REE-depleted boninites from Cape Vogel (Walker and Cameron, 1983; König et al.,
491 2010; Kamenetsky et al., 2002; Cluzel et al, 2016). According to the classification scheme of Pearce
492 and Reagan (2019), they diagnostically plot in the domain of high-silica boninites (Suppl. Fig. S4a). On
493 the expanded incompatible trace elements diagram normalized to N-MORB, they exhibit a well-
494 defined negative anomaly in Nb, Ta and Ti and a positive anomaly in Zr and Hf very similar to the
495 patterns of Cape Vogel boninites. In contrast, Nepoui boninites exhibit slightly higher incompatible
496 elements, more pronounced Ti negative anomalies (Cluzel et al., 2016; Fig. 4b) and lower contents in
497 compatible elements compared to Cape Vogel boninites.

498 Based on multi-element geochemical modeling Cluzel et al. (2016) proposed that Nepoui CE-
499 boninites derived from a moderately depleted mantle rock modified by low degree partial melts of a
500 mafic source similar to the BABB-like amphibolites of the metamorphic sole. Remarkably, the
501 calculated melting residue strongly differs from the ultra-depleted harzburgites of Peridotite Nappe,
502 ruling out that these harzburgites represent the mantle residue left after boninites extraction.

503 $^{40}Ar/^{39}Ar$ dating of glassy fragments yielded Early Eocene cooling ages (47.4 ± 0.9 Ma and $50.4 \pm$
504 1.3 Ma) (Cluzel et al., 2016). Considering the glassy character of the rock, the older age is likely to
505 approach the emplacement age.

506 Numerous dykes within the Peridotite Nappe exhibit geochemical features akin to those observed
507 in the Nepoui CE-boninites, i.e. REE contents, high-MgO, negative Nb, Ta and Ti anomalies, positive Zr
508 and Hf anomalies (Figs. 4e and 4f), and were referred to as boninite-series by Cluzel et al. (2016).
509 However, none of them are *sensu stricto* boninites and they dominantly plot in the domain of high-
510 Mg andesites on the classification diagram of Pearce and Reagan (2019) (Suppl. Fig. S5). Although no
511 modeling was performed for these rocks, a process similar to that of CE-boninites may be postulated
512 for their origin; i.e. partial melting from a mantle source modified by slab melts (e.g. Wood and
513 Turner, 2009). U-Pb zircon dating of these rocks provided a narrow age range between 54.5 ± 0.7 Ma
514 and 53.2 ± 0.6 Ma (Cluzel et al. 2016), within the age range of the other dykes of the Peridotite Nappe.

515 ***Hornblende-bearing dykes and hornblendites***

516 Hornblende-gabbro and diorite dykes are common in ophiolites from various tectonic settings. In
517 New Caledonia they occur at all levels of the ophiolite but decrease in frequency from the base to the
518 top of the Peridotite Nappe. Their compositions vary from hornblendite (with less than 5%
519 plagioclase) to leucocratic anorthosite (less than 5% hornblende) through hornblende gabbro/diorite.
520 Many hornblende-bearing dykes display structures that record high- to medium-temperature
521 deformation compatible with transcurrent kinematics (Cluzel et al., 2021). Textures such as

522 amphibole prisms with preferred orientation, compositional layering, isoclinal folding, sigmoidal
523 stretched amphibole grains and mylonitic fabrics suggest a syn- to late-magmatic deformation of the
524 dykes (Suppl. Figs. S3c and S3d). The host peridotites do not display any sign of similar ductile
525 deformation that could be coeval to these textures suggesting that the hydrous magmas were
526 injected in active faults in a transcurrent tectonic setting (Cluzel et al., 2021). Hornblendite samples
527 display slight LREE depletion (Fig. 8a) with hook-shaped patterns typical of amphibole-dominated
528 modal compositions (Henderson, 1984). In the REE and trace elements MORB-normalized diagram
529 these samples display Nb, Ta and Ti negative anomalies diagnostic of subduction-related magmas
530 (Fig. 8b). Zircon U-Pb (50 ± 3 Ma; Suppl. Fig. S5a) and hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ dating (54.9 ± 2.8 Ma; Suppl.
531 Fig. S5b) of representative hornblendite samples (RPYR 6 and VOH 1 respectively, Appendix Table S4
532 and S5) consistently indicate lower Eocene ages, which are within uncertainty of each other.

533 The chemistry of hornblende-gabbro/diorite dykes shows higher incompatible elements
534 abundances with respect to hornblendites, with positive and negative Eu anomalies, systematic
535 negative Nb, Ta and Ti anomalies, and Zr-Hf positive anomalies (Fig. 8c and 8d). Variable high-field
536 strength elements (HFSE) depletion in many hornblendites and hornblende gabbro/diorite samples
537 (Fig. 8b and 8d) are consistent with the occurrence of residual garnet in their sources. U-Pb dating of
538 zircons from 12 samples provides crystallization ages from 55.7 ± 1.3 Ma to 52.3 ± 0.9 Ma with an
539 average of ca. 54 Ma (Suppl. Fig. S6) (Appendix Table S4). These new data are consistent within error
540 with the previously published average of 53 Ma (Cluzel et al., 2006); however, in detail, such older
541 average value accounts for the new sampling of the middle and northern parts of Peridotite Nappe
542 where dykes are slightly older than in Massif du Sud (see discussion below).

543 Anorthosite dykes are light-colored rocks sharply contrasting in the field with the host peridotite.
544 They form cm- to m-thick coarse to medium-grained dykes that represent the leucocratic
545 (amphibole-poor) end-member of the hydrous rock suite. They typically show variable REE contents
546 with slightly U-shaped patterns, some of which are much depleted, with a prominent Eu positive
547 anomaly (Fig. 8e) that suggest plagioclase accumulation. However, the positive Eu anomaly does not
548 increase together with decreasing bulk REE content, suggesting that plagioclase enrichment did not
549 occur by fractional crystallization solely. In contrast, plagioclase, and in counterpart, hornblende
550 enrichment may result from solid-state mineral segregation within crystal mushes. Examples of
551 mineral segregation within dykes are particularly well exposed in Ouen Island where a compositional
552 layering (Suppl. Fig. S3e) is isoclinally folded in some internally deformed dykes (Suppl. Fig. S3f). The
553 occurrence of compositional layering and hornblendite and anorthosite enclaves within the same
554 gabbro dyke (Suppl. Figs. S3g and S3h) are evidence for their cogenetic character. Therefore, it is
555 suggested that solid-state segregation of plagioclase in anorthosites represents the counterpart of

556 amphibole enrichment in hornblendite. Due to the low Zr content (2-100 ppm) zircons are rare in
557 these rocks and only samples with $Zr > 60$ ppm could be dated. Zircons from sample NFC 4 yielded a U-
558 Pb age of 52.6 ± 0.4 Ma (Appendix Suppl. Fig. S7, Table S4); i.e. within the age range as the bulk of
559 lower Eocene dykes.

560

561 ***Granites***

562 Biotite and quartz-bearing felsic dykes are not uncommon in the Peridotite Nappe and similarly to
563 other dyke types display various textures, from granular to pegmatitic, due to relatively slow cooling.
564 They typically contain 71-78.7 wt% SiO_2 , 12.3-15.3 wt% Al_2O_3 , 3.7-6.7 wt% Na_2O+K_2O , low K_2O (0.1-2
565 wt%) and low FeO and MgO contents (Appendix Table S1) plotting in the granite field of the TAS
566 diagram (Le Bas et al., 1986; Le Maitre et al., 2002; Fig. 2). The sample OUEN 3 differs by its extreme
567 depletion in some incompatible elements (e.g. $Na_2O+K_2O = 0.81$) and HREE. Samples KO6 and KO11
568 display higher K_2O contents (4.4-4.7 wt%). On the REE diagram normalized to the chondrite C1
569 (Evensen et al., 1978) two kinds of patterns are distinguished: type 1) spoon-shaped patterns with or
570 without Eu anomalies, and type 2) slightly U-shaped patterns with or without Eu anomalies (Fig. 8g).
571 The sample OUEN 3 (type 1) displays an extreme depletion in HREE and a prominent Eu positive
572 anomaly. On the REE and trace elements diagram normalized to N-MORB (Fig. 8h), the type 1
573 granites display variable Nb-Ta negative anomalies and Zr-Hf positive anomalies and the type 2 does
574 not display Nb-Ta anomalies. Type 1 granites display trace-elements patterns typical of adakite series
575 and their similarities with hornblende-bearing dykes likely denote a similar origin through a lesser
576 melting degree and/or fractionation. In contrast, the U-shaped REE patterns of type 2 granites
577 strongly recall that of boninite series rocks, although they are not boninites sensu stricto (Appendix
578 Fig. S5). Similarly to the other dykes except dolerites, their U-Pb zircon ages define a narrow range
579 between 54.2 Ma and 53.1 Ma (Appendix Fig. S7 and Table S4).

580

581 ***Dolerite dykes***

582 Dolerite dykes 1-10 m thick, are relatively rare in the Peridotite Nappe and a crosscutting
583 relationship with other dykes was never observed; however, microcrystalline texture and occurrence
584 of chilled margins suggest intrusion in an already cold peridotite and late emplacement compared to
585 other dyke types. Dolerite dykes are oriented $N20^\circ E$ on average (Titus et al., 2011; Cluzel et al.,
586 2021), suggesting brittle NW-SE-directed extension at the time of dolerite intrusion.

587 The dolerite samples examined in this study were taken from various locations, including the
588 Massif du Sud (Montagne des Sources, Tontouta, and Dumbea river valleys) and the west coast
589 massifs (Boulinda, Tiebaghi). As such, they provide a representative overview of the entire Peridotite
590 Nappe. The analyzed samples are basalts or basaltic andesites (Fig. 9a) that display LREE depletion
591 similar to MORB in chondrite-normalized diagrams (Fig. 9b). They also have variable but prominent
592 negative anomalies in Nb and Ta and weakly negative Ti anomalies normalized to N-MORB (Fig. 9d).
593 These features and Hf-Th-Ta relationships (Wood, 1980; Vermeesch, 2006; Fig. 9c) allow a
594 classification as island arc tholeiites (IAT). Only one sample that displays a negative slope on C1-
595 normalized REE diagram plots in the Island arc basalt (IAB) domain of the Hf-Th-Ta diagram (Fig. 9c).

596 A K-Ar age of 50 Ma was previously reported for a dolerite crosscutting a cumulate lens at
597 Montagne des Sources (Massif du Sud) (Paris, 1981), but no analytical data were provided in this
598 work. One dolerite dyke from Boulinda Massif (Si Reis nickel mine) was selected for $^{40}\text{Ar}/^{39}\text{Ar}$ dating
599 (SIR 8, this study). This sample is one of the most primitive in the dolerite sample set (Fig. 9b) and
600 indicated a cooling age of 50.4 ± 0.2 Ma (Fig. 10) (Appendix Table S6).

601 A dredge on the top of continental slope a few km to the west of Koumac (DR4, IPOD program)
602 collected vitric and vesicular volcanic rock fragments embedded in a late Pliocene micritic matrix, the
603 location of which was judged enigmatic by the authors (Mortimer et al., 2018). This rock, classified as
604 andesite based on Nb/Y ratio (Fig. 9a), has 'immobile' trace-elements features (Nb-Ta and Ti negative
605 anomalies, Hf/Th and Th/Ta ratios) that place it in the domain of island-arc tholeiite series, similar to
606 the low Nb/Y IAT dykes of Peridotite Nappe (Figs. 9c and 9d). Poorly defined $^{40}\text{Ar}/^{39}\text{Ar}$ apparent ages
607 of 40-48 Ma (Mortimer et al., 2018) may be interpreted as minimum ages due to marine alteration
608 and low-temperature Ar loss, and make it a probable differentiated product of the Eocene IAT
609 magmas.. We suggest that these rocks, along with similar, though undated rocks reported from the
610 western New Caledonia margin (Mortimer et al., 2018), were first emplaced at a shallow level of the
611 Peridotite Nappe, or erupted on its top and later collapsed onto the continental slope.

612 ***Andesites from the Intermediate Melange***

613 The Intermediate Melange of northern New Caledonia consists of elongate serpentinite mélangé
614 lenses that contain various rock types, including blocks of serpentinitized harzburgites, dolerites,
615 basalts, cherts and leucocratic rocks (Cluzel, 2020). These lenses, 5 to 50 km long, are pinched within
616 Diahot Terrane schists and display the same greenschist/blueschist to lower eclogite HP-LT
617 metamorphism and foliation attitude. The material of this melange was scraped off from both the
618 Poya Terrane (accreted in front of, and beneath the forearc) and from the mantle wedge
619 (serpentinitized peridotites and dismembered dykes, see Cluzel, 2020).

620 The Haute Néhoué Unit is a large block (ca 1 km-wide) within the melange, located at the
621 northwestern end of an 11 km-long NW-SE trending melange sliver, which contains smaller blocks of
622 Poya Terrane basalts, dolerites, red cherts, felsic rocks and deeply serpentinized harzburgites
623 enclosed in a sheared serpentinite matrix. Haute Néhoué Unit consists of a coherent set of mafic and
624 intermediate rocks in tectonic contact with isoclinally folded Late Cretaceous black cherts and
625 Paleocene micrite of the Diahot Terrane in the south, and with Poya Terrane basalt blocks and Late
626 Cretaceous black cherts to the north (see Cluzel, 2020 for details). It is composed of greenish pillow
627 lavas and medium-grained felsic dykes. Geochemical classification based upon immobile trace
628 elements ratios (Zr/TiO_2 , Nb/Y) (Winchester and Floyd, 1977) delineate one single fractionation trend
629 from andesitic basalt to dacite (diorite) composition for these rocks (Suppl. Fig. S8a). Chondrite-
630 normalized REE patterns display moderate LREE enrichment and flat HREE patterns with moderate
631 positive and negative Eu anomalies denoting some plagioclase accumulation and fractionation
632 respectively (Suppl. Fig. S8b and S8c). Parallel REE and incompatible trace elements patterns suggest
633 origin from a common parental melt via moderate degree of fractionation. The most evolved dyke
634 (medium-grained diorite #NHE10) yielded a U-Pb zircon age of 54.7 ± 0.8 Ma (Cluzel, 2020), indicating
635 that the andesitic sequence is coeval with the earliest dykes of the Peridotite Nappe.

636 ***Inherited zircons***

637 With the exception of the diorite dyke of the Haute Néhoué Unit, in which the occurrence of 23%
638 inherited zircon grains (at 978, 435, 139 and 129 Ma; Cluzel, 2020) signals contamination by crustal
639 rocks or, more probably, clastic sediments (late Cretaceous ?) and despite the large number of
640 zircons analyzed from Lower Eocene magmatic rocks, no zircon older than 55.5 Ma was found in the
641 25 representative dyke samples dated by the U-Pb method. We thus infer that neither crustal rock
642 nor clastic sediment was involved in the genesis of Lower Eocene dykes of Peridotite Nappe.

643

644 ***Nd-Sr, Hf and oxygen isotope geochemistry***

645 Representative samples of the various dyke types were analyzed for Sr and Nd isotopes at the
646 CNRS isotope facility of the Laboratoire Magmas et Volcans, University of Clermont-Ferrand, France)
647 (see appendix for details). Apart for a minor trend toward higher $^{87}Sr/^{86}Sr$ ratios likely due to low-T
648 alteration, all dyke types display Sr-Nd isotope signatures close to Depleted MORB Mantle (DMM). In
649 particular, ϵ_{Nd} values recalculated at 54 Ma vary between 6.33 and 10.12 (Appendix Table S7 and Fig.
650 11). This precludes the involvement of continental crust or sedimentary material, in agreement with
651 the absence of xenocrystic zircons or older zircon cores (see above and Xu et al., 2021). Remarkably,
652 amphibolites of the metamorphic sole yielded similar values ($7.11 < \epsilon_{Nd} < 10.04$; avg.=8.50), which may

653 indicate some genetic relationship (Fig. 11). Nd-Sr isotope ratios of dolerites ($7.41 < \epsilon_{Nd} < 9.36$;
654 $avg.=8.8$) are consistent with their tholeiitic (IAT) composition.

655 Hafnium isotope signature of zircons from gabbro and diorite blocks from the serpentinite sole
656 ($9.4 < \epsilon_{Hf} < 12.9$) (Xu et al., 2021) are also consistent with the origin from a DMM-type source.
657 However, the oxygen isotope composition of zircons ($+4.74\text{‰} < \delta^{18}O < +6.63\text{‰}$) suggest minor
658 involvement of subduction-related aqueous fluids in magma genesis (Xu et al., 2021).

659

660 **Discussion**

661

662 The tectonic and geochemical features of the lower Eocene dyke system raise issues and provide
663 clues on the geodynamic setting of New Caledonia Ophiolite and the shear zones that crosscut it.

664 ***The transform fault setting of HT shear zones questioned***

665 The current paleotransform interpretation of Belep and Bogota shear zones, which are thought to
666 connect laterally with the mantle fabric of Peridotite Nappe (Maurizot et al., 2020c), suggests the
667 occurrence of differently oriented spreading ridge segments (Suppl. Fig. 9a), a common feature of
668 back-arc basins (e.g., North Fiji Basin, Auzende et al., 1995) and could account for oroclinal bending
669 of the northernmost part of Norfolk/New Caledonia Ridge (Northern D'Entrecasteaux Ridge) (Cluzel
670 et al., 2021). However, their geometric relationships do not match those of transform fault system of
671 multiple ridge basins (see Fig. 1). Alternatively, NW-SE (Belep) and NNE-SSW trending (Bogota)
672 dextral shear zones could correspond to C and C' shears respectively, related to regional transcurrent
673 tectonics (e.g., Finch et al., 2020 and references therein; Suppl. Fig. 9b). This interpretation better fits
674 the dextral transcurrent tectonics marked by km-scale "Z-folds" with vertical axes that preceded
675 obduction in the Eocene HP-LT belt and in the Poya Terrane as well (Cluzel et al., 2021). It is worth
676 noting that the occurrence of sheared gabbro dykes re-injected by the same magma, dated at ca 54
677 Ma, which remained undeformed indicates the initiation of such tectonic regime in the upper plate
678 at the beginning of the 55.5-50.0 Ma time interval (Cluzel et al., 2021). Similarly, undeformed
679 pyroxenite dykelets of the Bogota shear zone dated at ca 56 Ma, which postdate ductile motion,
680 were intruded immediately after subduction inception at ≥ 56 Ma. Hence, a syn-subduction
681 transcurrent tectonics interpretation for the development of these high-T shear zones seems more
682 likely. The presence of syn-kinematic orthopyroxenite dikes derived from boninite melts in the
683 Bogota shear zone, reinforces such interpretation. Transcurrent tectonics occurring at subduction
684 inception is consistent with an induced (forced) subduction (Stern, 2004), which may overcome the

685 buoyancy of the hot oceanic lithosphere when intra-oceanic subduction starts at or near a spreading
686 center.

687 The occurrence of high-temperature shear zones now situated a few tens to hundreds of meters
688 above the basal thrust of the ophiolite (i.e. the exhumed roof of the Eocene subduction zone), and
689 reworked supra-subduction dykes in the serpentinite melange, point toward subduction erosion
690 (Cluzel, 2021) and fast uplift along the subduction channel. This interpretation is also supported by
691 the cooling rate calculated for mantle wedge peridotites (Secchiari et al., 2022a) and Lu-Hf garnet
692 age constraints obtained from exhumed eclogites of the Pouebo Terrane (~38 Ma, Taetz et al., 2021).
693 Both lines of evidence point to fast cooling (10-100°C/Myr and 50°C/Myr respectively) and high
694 exhumation rates ($7 \cdot 10^3 \text{m/Myr}$).

695

696 ***Genesis of hornblende-gabbro/diorite dykes; a modeling approach.***

697 With the exception of extreme compositions, all hornblende-bearing dykes share some
698 geochemical features, i.e., LREE/HREE enrichment with moderate (e.g. Gd/Yb up to 3.5, supplem.
699 Table S1), low Y, prominent Nb-Ta negative anomalies, variable Ti negative anomalies and Zr(-Hf)
700 positive anomalies (Fig. 8) that argue for the presence of amphibole, garnet and rutile as residual
701 minerals in their sources. Overall, these geochemical signatures are typical of adakites found in
702 subduction zone settings worldwide. However, the Sr/Y ratios of most Lower Eocene hornblende
703 dikes vary in a wide range (Fig. 12 and Appendix Table S1). Only 30% of the analyzed samples have
704 $\text{Sr/Y} > 40$ and could be therefore recognized as adakites s.s. according to the original definition of
705 Defant and Drummond (1990). Taking into account the less restrictive classification proposed by
706 Richards and Kerrich (2007), up to ~60% of the samples can be considered as adakite-like rocks
707 ($\text{Sr/Y} > 20$) derived from the partial melting of a mafic source. The variable Sr/Y ratios and HREE
708 depletion of gabbro/diorite dykes may be viewed as a consequence of the variable amounts of
709 garnet in their source (e.g., Moyen, 2009).

710 The HT amphibolites of the metamorphic sole of the NC Ophiolite could be viewed as proxy of the
711 slab source of hornblende dike-forming melts. Overall, MORB-type Nd isotope signatures of these HT
712 amphibolites are consistent with the compositions of hornblende gabbro/diorite dikes (Fig. 11).

713 Melting modeling was therefore attempted using the trace-elements composition of the BABB-type
714 amphibolites of the metamorphic sole (Cluzel et al., 2012b; Fig. 13a) and partition coefficients of
715 Bédard (2006) modified by Zhang et al. (2013) (Appendix Table S8). The best fit results were obtained
716 with a protolith containing 5 wt% garnet, 70 wt% amphibole, 20 wt% plagioclase, 3% titanite and 1%
717 rutile for an equilibrium melting rate of 20-40%. Owing to the variability of amphibolites of the

718 metamorphic sole, end-member compositions were used in the calculation (Hi and Lo in Fig. 13a).
719 The majority of the hornblende-diorite/gabbro dikes fall in the range of model melts (Fig. 13b), with
720 some variability of LILEs and Ti contents though. While Ti may be controlled by retention of rutile and
721 its different proportions in the source rocks, the prominent Rb (and K, not shown)
722 variability/depletion in the investigated hornblende dykes most likely results from post-solidus
723 (hydrothermal) elemental mobility.

724 A significant number of hornblende diorite/gabbro dikes (as listed in Supplementary Table S1)
725 exhibit higher MgO (8.4-13.2 wt%) than typical adakites (MgO = 3-6 wt%, Defant and Drummond,
726 1990; Richards and Kerrich, 2007), or than melts with intermediate silica contents experimentally
727 obtained from dehydration melting of amphibolite (Rapp and Watson, 1995). Such chemical features
728 are frequently coupled with high compatible trace element contents (e.g. 300 ppm < Cr < 3500 ppm),
729 which point to interaction of slab-derived melts with an ultramafic component (Rapp et al., 1999;
730 Martin et al., 2005). Mixing with a peridotite-derived component, either through melt/rock
731 interaction (Rapp et al., 1999; Prouteau et al., 2001) or melting of a peridotite source metasomatized
732 by silicic slab melts (Martin et al., 2005) may partly obscure/overprint pristine adakitic signatures,
733 lowering initially high Sr/Y and La/Yb ratios.

734 Notably, most hornblendites and anorthosites strongly diverge from the major element
735 composition of hornblende gabbro/diorites. However, they generally retain similar trace element
736 (e.g. Nb, Ta, Ti negative anomalies) and Sr-Nd isotope signatures. These rocks have mineralogical and
737 textural features such as hornblende cracks filled with plagioclase, hornblendite/anorthosite layering
738 (Appendix Supplem. Figs. S3e and S3f) and the occurrence of comagmatic anorthosite and
739 hornblendite enclaves within hornblende-gabbro dykes (Appendix Suppl. Figs. S3g and S3h), which
740 suggest that these rocks formed by the segregation of solid mineral phases from crystal mushes
741 originating from similar melts as hornblende diorite/gabbros dykes.

742 ***The switch to IAT and the thermal regime of subduction***

743 The late occurrence (~ 50Ma) of tholeiitic dykes with subduction signatures (IAT) coeval with the end
744 of slab melt production suggests a change in the thermomechanical regime of the subduction zone.
745 Slab cooling after subduction inception took about 12 Ma to evolve from the HT-LP gradient close to
746 the spreading ridge (56 Ma) to the HP-LT gradient recorded by eclogites at 44 Ma (Spandler et al.,
747 2005). Petrochronological studies performed on the Semail Ophiolite, which share some analogies
748 with New Caledonia, have shown that the intra-oceanic subduction zone remained warm for about 5
749 My after subduction inception (Soret et al., 2022), a time span comparable to the duration of slab
750 melting in NC. After that, eclogitization of the mafic crust increased its density and provoked the

751 acceleration of subduction, slab steepening and corner flow. Meanwhile, slab dehydration that
752 followed the cessation of slab melting (e. g., Mibe et al., 2011) induced hydrous melting of the
753 uplifted asthenosphere and generated IAT dykes in the upper plate.

754 ***Evolution of melts generated at subduction inception***

755 The prominent diversity of dykes generated at subduction inception suggests the occurrence of at
756 least five different melt series:

- 757 – “Direct” slab melts formed by low-degree partial melting of the HT amphibolites
758 (hornblende-gabbro/diorite series) variably influenced by an ultramafic component.
759 Hornblendites and anorthosites represent end-members of solid-phase separation within
760 magmatic conduits;
- 761 – Melts derived from moderately depleted mantle wedge peridotites modified by slab melts
762 (CE-boninite), which in turn, generated gabbro-cumulates
- 763 – Melts generated by hydrous melting of mantle wedge peridotites modified by CE-boninite
764 melt (hornblende-pyroxenites, websterites);
- 765 – Tholeiitic melts derived from “fertile” mantle sources that intruded an already cooled host
766 rock (IAT dolerites);
- 767 – Andesites formed by hydrous melting of undepleted peridotite with some sediment
768 contribution.

769 The geochemistry of the pyroxenite dykes shows that they were likely generated from hydrous
770 melting of moderately depleted harzburgite previously modified by slab melt inputs, similar to the
771 parental magma of the CE-boninite. The local New Caledonia harzburgites are too depleted to be the
772 melting residue of these melts (Cluzel et al., 2016; Secchiari et al, 2022b) indicating a different
773 mantle source. The volumetrically predominant hornblende-gabbro/diorite dykes are likely to have
774 been mostly slab-sourced and were intruded during the whole episode of forearc magmatism (55.5
775 Ma to 50 Ma). The end of this episode of slab melting that recorded slab cooling at ~50 Ma coincided
776 with the occurrence of mantle-sourced IAT dolerites. Therefore, the ophiolite of New Caledonia
777 belongs to a minor group of supra-subduction ophiolites in which the magmatic activity is dominated
778 by adakite-like rocks that resulted from partial melting of the subducting slab (Furnes and Dilek,
779 2017).

780 Magmatic rocks that erupted or intruded in the forearc soon after subduction inception in the
781 interval 56-50 Ma generally do not display any discernable evolution through time (e.g., Mg# and
782 Sr/Y; other elemental ratios not presented) (Suppl. Figs. S10a and S10b); except an increase of the

783 Th/Ta ratio and negative Nb (and Ta) anomaly (Suppl. Figs. S10c and S10d), which both record an
784 increasing influence of amphibole in the source rock through time. There is only a small difference in
785 the timing of dyke intrusion along strike the island. Dyke emplacement started at ~54 Ma on average
786 in the northern (west coast) massifs vs. ~ 53Ma in Massif du Sud (Supplem. Fig. S11a). Such
787 diachronism suggests that subduction was not only oblique at start, but also progressed southward in
788 a fan-like fashion due to the bending of the retreating slab (Supplem. Fig. S11b). This feature is
789 consistent with the Eocene anticlockwise rotation of New Caledonia recorded by paleomagnetic data
790 (Cluzel et al., 2021).

791 Subducted sediments did not contribute to the genesis of slab melts because the basalts of the
792 lower plate that recrystallized to generate the amphibolites of the metamorphic sole, source of slab
793 melts, were too young to carry sediments. When sediments transported by the lower plate (i.e., the
794 Norfolk Ridge) finally entered the subduction zone, the HP-LT thermal regime was already
795 established and led to eclogite crystallization instead of partial melting.

796 The origin of andesitic pillow lavas erupted on the sea floor above the subduction zone at
797 54.0 ± 1.0 Ma (Cluzel, 2020) is an unusual feature at hot subduction inception, i.e. before the corner
798 flow is established. It could be postulated that lower Eocene andesites resulted from the
799 differentiation of IAT magma; however, their REE and trace elements contents are drastically
800 different (Fig. 9 and Suppl. Fig. S8); in addition, the IAT dykes have significantly higher Hf/Th ratios
801 (>3) that rule out contamination by continental crust or clastic sediments. It is suggested that
802 andesites originated from the melting of a previously metasomatized mantle source uplifted locally
803 due to transcurrent tectonics (transtension) in a subsidiary forearc basin. The occurrence of inherited
804 zircon grains in the cogenetic diorite dyke (Cluzel, 2020) signals contamination by crustal rocks
805 crosscut by the ascending magma or, more likely, subducted Late Cretaceous clastic sediments,
806 which are a significant component of the HP-LT belt (Diahot Terrane).

807 The dating and geochemistry outlined in this study show that melting occurred in a variety of
808 different sources over a relatively short period of time with a large predominance of slab melts.
809 Although the context is somewhat different (intra-oceanic subduction vs. active margin), these
810 results meet in part the conclusions of Kimura et al. (2014) based upon the late Cenozoic volcanism
811 of SW Japan associated with the subduction of the young Shikoku Basin. Accordingly, forearc magma
812 diversity is due to the interaction of slab melts with diverse sources, as also proposed by Yu et al.
813 (2020) for the west Philippine ophiolites, where the first magmatic products vary from normal to
814 enriched MOR-type basalts, with no boninite occurrence. This finding does not fit the 'subduction
815 initiation rule' (SIR; Whattam and Stern, 2011), which was established from the western Pacific Izu-

816 Bonin-Mariana (IBM) subduction. In the IBM system and some Tethyan-type ophiolites (e.g. Dilek and
817 Thy, 2009; Shervais et al., 2019, 2021; Coulthard et al., 2021), forearc basalts erupted first, followed
818 by boninites before the transfer of the magmatic activity into the volcanic arc itself. Similarly, the
819 progressive magmatic evolution of the NC ophiolite does not follow the typical sequence predicted
820 for the supra-subduction type ophiolite by Dilek and Furnes (2011), i.e. N-MORB, IAT and boninites,
821 exemplified by the Jurassic Albanides-Hellenides ophiolites (Saccani et al., 2018).

822 Differences between the IBM system and the sequence of events in the Early Eocene to the east
823 of New Caledonia indicate different subduction initiation mechanism. The IBM arc is thought to have
824 been established from the onset of subduction in a “cold” old Western Pacific plate. In contrast,
825 Eocene subduction in New Caledonia started near a spreading center and involved young, hot and
826 buoyant oceanic lithosphere (Cluzel et al., 2012a, 2020). Hence, the conditions necessary to generate
827 forearc basalts at subduction onset were not met. Tholeiitic basalt dykes (IAT) were emplaced 6 myr
828 after subduction inception, when the down-going plate was sufficiently cold for eclogites to form,
829 which would have increased the slab bulk density and associated steepening. This may have resulted
830 in the incipient retreat of the slab, allowing mantle upwelling and generation of IAT. Afterwards, the
831 magmatic activity was probably transferred into the volcanic arc itself (the Loyalty Arc) and magma
832 generation in the forearc ceased.

833 This evolution hypothesis challenges the correlation of NC and IBM subduction zones proposed by
834 many authors (e.g. Whattam et al., 2008, Mortimer et al. 2018, Agranier et al., 2023). In detail, these
835 subduction zones differ on several key points: i) subduction inception in NC and IBM was only
836 roughly synchronous, with NC subduction starting before 56 Ma; i.e., circa 4-5 myr before IBM
837 subduction initiation at 51-52 Ma (Reagan et al., 2013; Ishizuka et al., 2018); ii) NC subduction was
838 hot and shallow-dipping at the beginning and may be considered as forced subduction. In contrast,
839 IBM involved old and cold western Pacific lithosphere, was steeply dipping and likely spontaneous
840 (i.e. gravity-driven) ; iii) IBM subduction zone was (and is still) plunging continent-ward (i.e., W-ward),
841 while NC subduction originally plunged ocean-ward (i.e., N- or NE-ward after mid-Eocene
842 anticlockwise rotation).

843 **Conclusion**

844 Lower Eocene dykes in the New Caledonian Peridotite Nappe dominantly originated from slab
845 melts generated by low-pressure melting of high-temperature amphibolites. There was no sensible
846 evolution of magmatic activity through time except an increase of amphibole signature in the source,
847 probably by metamorphic disequilibrium of clinopyroxene. All lithotypes occurred throughout the
848 period except IAT dolerites that appeared in the final stage and intruded already cooled mantle

849 wedge rocks. Slab melts formed at relatively low pressure (garnet-poor source rock) from the still hot
850 and shallow-dipping slab; in contrast, IAT appeared by the end of the forearc magmatic activity and
851 record the change into progressively cooler and more steeply dipping subduction zone (Fig. 14a).
852 Most slab melts were emplaced directly and cooled relatively slowly, while some interacted with the
853 mantle wedge peridotites and indirectly generated CE-boninite. In turn, CE-boninites and coeval
854 melts interacted with peridotites and were responsible for the formation of websterite intrusive
855 rocks (Figs. 14b and 14c). The melts were generated in the forearc of a shallow-dipping and hot
856 subduction zone, which originated near a spreading ridge at ca 56 Ma, progressively cooled, and then
857 plunged more steeply at ca 50 Ma (Fig. 14a). The bulk of forearc magmatism occurred during that
858 short time interval. Such an evolution is at odds with the forearc basalt-then-boninite sequence
859 suggested by the SIR; thus it seems that the sequence eruption in hot and shallow (forced)
860 subduction zones such as the north- or northeast-dipping Eocene subduction of New Caledonia
861 differs markedly from that in colder subduction zones such as the Izu-Bonin-Marianna system.

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870

871 **Data availability**

872 The datasets generated during and/or analysed during the current study are available in the [NAME]
873 repository, []

874

875 **References**

876

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1277 **Figure captions**

1278

1279 Fig. 1: Geological sketch map of New Caledonia. BSZ: Belep Shear Zone, HC: Humboldt Corridor, BTF:
1280 Bogota Transform Fault. Inset, LHR: Lord Howe Ridge, NR: Norfolk Ridge, LR: Loyalty Ridge, HP:
1281 Hikurangi Plateau.

1282

1283 Fig. 2: Total Alkali-Silica diagram (Le Bas et al., 1986; Le Maitre et al., 2002) to show the diversity in
1284 composition of Lower Eocene dykes. The rock nomenclature is mainly field-based; some
1285 discrepancies may be due to element mobility.

1286

1287 Fig. 3: Plum pyroxenite intrusion. (a) Geological map redrawn from the online geological map of New
1288 Caledonia, (www.georep.nc); (b) and (c) Tectonic sketch of transcurrent (dextral) emplacement of
1289 Plum intrusion and satellite dykes. (b) First step, transtensional intrusion of the main body. (c)
1290 Second step, passive rotation and emplacement of new shear-parallel dykes.

1291

1292 Fig. 4: Whole-rock geochemical features of Plum and Ouassé pyroxenites, Népoui boninite and
1293 boninite-series felsic dykes (a) Chondrite-normalized REE diagram (Evensen et al., 1978) for Plum
1294 pyroxenite (a), Ouassé pyroxenite (c) and boninite-series dykes (e); the dotted pattern represents the
1295 dated sample ($^{40}\text{Ar}/^{39}\text{Ar}$). (b) MORB-normalized REE and trace elements spiderdiagram (Sun and
1296 McDonough, 1989) for Plum pyroxenite (b), Ouassé pyroxenite (d) and boninite-series dykes (f).
1297 Dotted patterns refer to dated samples (U-Pb zircon). REE and REE-Tr diagrams for Népoui boninite
1298 (Cluzel et al., 2016) and modeled melts for pyroxenites of Plum (Secchiari et al., 2022b) and Ouassé
1299 (Ferrari, 2021) are shown for comparison.

1300

1301 Fig. 5: $^{40}\text{Ar}/^{39}\text{Ar}$ data for the amphiboles of the Plum samples; PL-AM7 pyroxenite and PL-AM10
1302 hornblende-gabbro dyke. (a, b) show the age spectra, K/Ca and radiogenic yield diagrams obtained
1303 by the incremental heating experiments and (c,d) are the isochron diagrams. The isochron ages are
1304 the preferred cooling ages of the samples.

1305

1306 Fig. 6: Geological map of Bogota Peninsula (redrawn from the online geological map of New
1307 Caledonia, www.georep.nc)

1308

1309 Fig. 7: $^{40}\text{Ar}/^{39}\text{Ar}$ degassing diagram for the amphiboles of Ouassé pyroxenite dykelets.

1310

1311 Fig. 8: Chondrite-normalized REE diagrams (Evensen et al., 1978) and MORB-normalized REE and
1312 trace elements spiderdiagrams (Sun and McDonough, 1989) for hornblende-bearing dykes,
1313 anorthosites and granites: a-b, hornblendites; c-d hornblende-rich gabbros/diorites, e-f,
1314 "anorthosites", g-h granites. (a) and (b) dashed patterns represent dated samples (Ar/Ar); (c) and (d)
1315 blue patterns represent dated samples (U-Pb zircon); (e) and (f), dashed patterns represent dated
1316 samples (U-Pb zircon); (g) and (h) plain lines type 1 granites, dashed lines type 2 granites.

1317
1318 Fig. 9: Whole-rock geochemical features of dolerite dykes. (a) Classification diagram of Winchester
1319 and Floyd (1977), most samples plot in the field of andesitic basalts; the white triangle represents the
1320 sample #DMBE 3 (IAB-like) and the purple dot represents the andesite dredge sample from IPOD
1321 prograd. (b) Chondrite-normalized REE diagram (Evensen et al., 1978); the bold dotted pattern
1322 represents #SIR 8 dated at 50.4 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$). (c) Ternary Hf-Th-Ta "discriminant" diagram (Wood
1323 1980; Vermeesch, 2006) most samples plot in the island-arc tholeiite (IAT) domain. (d) MORB-
1324 normalized REE and trace elements spiderdiagram (Sun and McDonough, 1989) same patterns as (b).

1325
1326 Fig. 10: $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum diagram diagram for the dolerite SIR 8 (Si Reis nickel mine, Boulinda
1327 Massif).

1328
1329 Fig. 11: Nd-Sr isotope ratios of selected Lower Eocene dykes and HT amphibolites of the
1330 metamorphic sole

1331
1332 Fig. 12: Sr/Y vs. Y discrimination diagram (Defant and Drummond, 1990) to show that about 60% of
1333 dykes have low Y contents and Sr/Y ratios >20 and thus correspond to adakite-like rocks (Richards
1334 and Kerrich, 2007. Note that only about 20% of dykes plot in the domain of adakite sensu stricto,
1335 these likely were issued from a source containing a significant amount of garnet.

1336
1337 Fig. 13: Slab melt modeling. (a) REE-Tr spiderdiagrams of HT amphibolites of the metamorphic sole
1338 (Cluzel et al., 2012b) taken as a protolith for slab melt modeling (Zhang et al., 2013). (b)
1339 Spiderdiagrams of modeled melts through 20-40% partial melting of amphibolites (green array)
1340 compared with the array of hornblende-diorite/gabbro dykes (hatched) and dated dykes (black lines).

1341
1342 Fig. 14: (a) simplified evolution of the Eocene subduction zone, from inception to slab steepening. (b)
1343 sketch model of shallow-dipping hot subduction to show the location of high-temperature
1344 amphibolites of the metamorphic sole and relationships between oblique subduction, development

1345 of shear zones and exhumation of the mantle wedge. (c) sketch diagram of fluid and melt circulations
1346 in the mantle wedge between 55 and 50 Ma.

1347

1348 **Supplementary figures (Appendix)**

1349

1350 Suppl. Fig. S1: columnar section of New Caledonia Ophiolite, mainly based on the Montagne des
1351 Sources sequence (Massif du Sud).

1352

1353 Suppl. Fig. S2: Field pictures. (a) moderately sheared hornblende-gabbro dyke crosscutting websterite
1354 (Plum intrusion); (b) foliated websterite (Plum intrusion); (c) HT sheared harzburgite with shallow-
1355 dipping stretching lineation (Ouassé Bay, Bogota Shear Zone); (d) pyroxenite dykelets crosscutting
1356 sheared harzburgite (Ouassé Bay); (e) boudinaged pyroxenite (white arrow, Ouassé Bay). Smyl=
1357 mylonitic foliation.

1358

1359 Suppl. Fig. S3: Field pictures. (a) anastomosed hornblendite dykes (N'go Pass); (b) weathered
1360 hornblende-gabbro dyke crosscutting Plum pyroxenite; (c) mylonitized hornblende-gabbro dyke
1361 (Plum); (d) mylonitized hornblende-gabbro dyke with eye-shaped hornblendite porphyroclast (Plum);
1362 (e) incipient hornblendite/anorthosite layering (composite dyke, Ouen Island); (f) isoclinally folded
1363 anorthosite/hornblendite internal layering, white arrow: anthophyllite rim around a peridotite ghost
1364 enclave (gabbro dyke, Ouen Island); (g) hornblendite enclaves in a hornblende-gabbro dyke (Ouen
1365 Island); (h) co-magmatic anorthosite and hornblende-rich enclaves in the same dyke (Ouen Island).

1366

1367 Suppl. Fig. S4: MgO-SiO₂ and MgO-TiO₂ diagrams for boninite classification (Pearce and Reagan,
1368 2019). Abbreviations: BA= basaltic andesite, LOTI= low-Ti basalts, SHMB= high-Mg basalts, HMA=
1369 high-Mg andesite, BADR= basalt-andesite-dacite-rhyolite suite, LSB= low-Si boninites, HSB= high-Si
1370 boninites.

1371

1372 Suppl. Fig. S5: (a) Terra-Wasserburg U-Pb zircon diagram for RPYR 6 hornblendite; (b) ⁴⁰Ar/³⁹Ar age
1373 spectrum diagram for VOH 1 hornblendite

1374

1375 Suppl. Fig. S6: Terra-Wasserburg U-Pb zircon diagrams of twelve selected hornblende
1376 gabbros/diorites.

1377

1378 Suppl. Fig. S7: Terra-Wasserburg U-Pb zircon diagrams of one anorthosite (a) and four selected
1379 granites (b).

1380

1381 Suppl. Fig. S8: Whole-rock geochemical features of Haute Néhoué andesites. (a) Classification
1382 diagram of Winchester and Floyd (1977), most samples plot in the field of andesite. (b) Chondrite-
1383 normalized REE diagram (Evensen et al., 1978); the bold dashed pattern in a and b represents #NHE
1384 10 diorite dyke dated at 54.7 ± 0.8 Ma (U-Pb zircon). (c) MORB-normalized REE and trace elements
1385 spiderdiagram (Sun and McDonough, 1989)

1386

1387 Suppl. Fig. S9: Tectonic sketch to illustrate the two possible interpretations of high-temperature
1388 shear zones of Peridotite Nappe. 3a: transform fault model; 3b: C-C' shears model

1389

1390 Suppl. Fig. S10: Evolution of some elemental ratios during the 57-48 Ma period. Note the absence of
1391 evolution of Mg# and Sr/Y (most other ratios not presented); in contrast, there is a neat increase of
1392 Th/Ta ratio and Nb negative anomaly through time.

1393

1394 Suppl. Fig. S11: (a) age-latitude relationships of Lower Eocene dykes based upon conventional U-Pb
1395 zircon ages only. Note the small age difference of ca. 1 myr between south-easternmost and north-
1396 westernmost dykes. Errors in latitude have been arbitrarily taken as 0.01° ; (b) sketch diagram of the
1397 Eocene subduction zone shortly after inception to account for southward age progression of forearc
1398 activity (for clarity, curvature at that stage is somewhat exaggerated).

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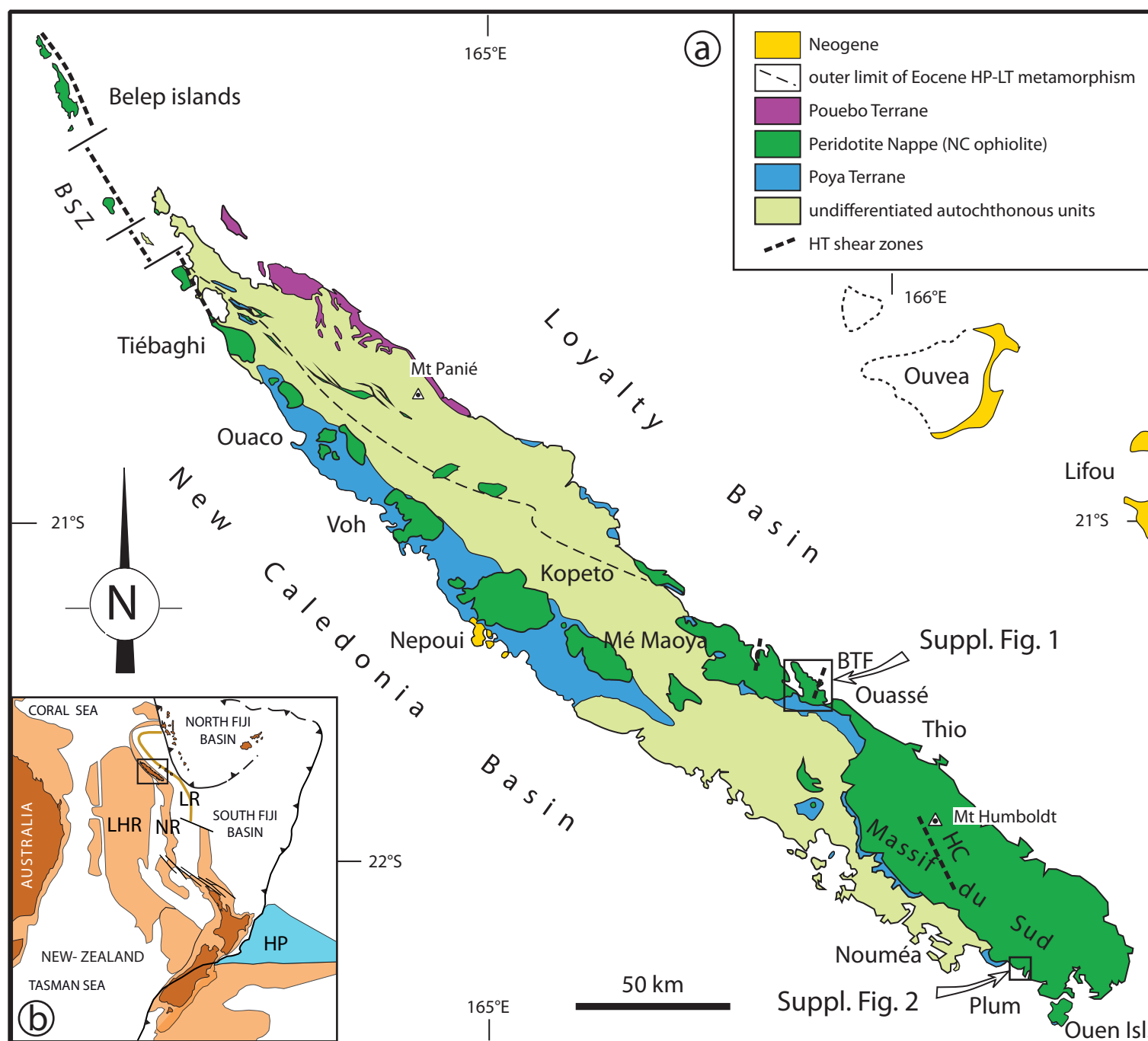
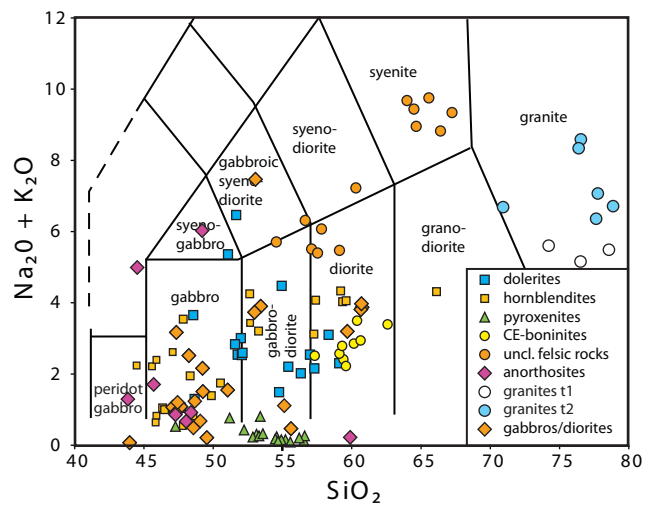
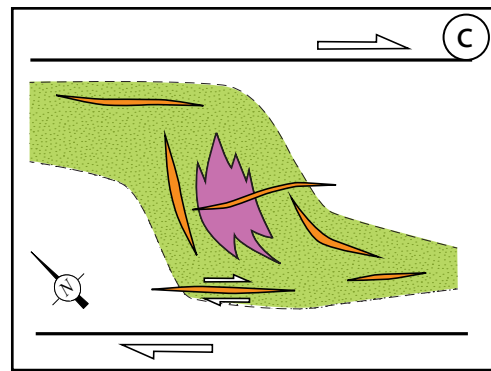
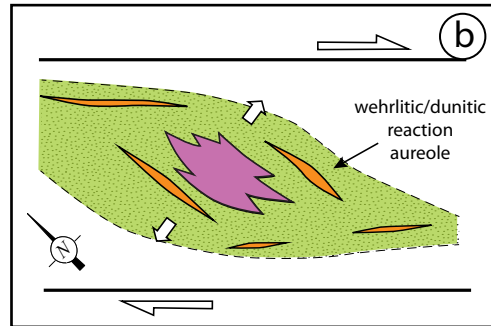
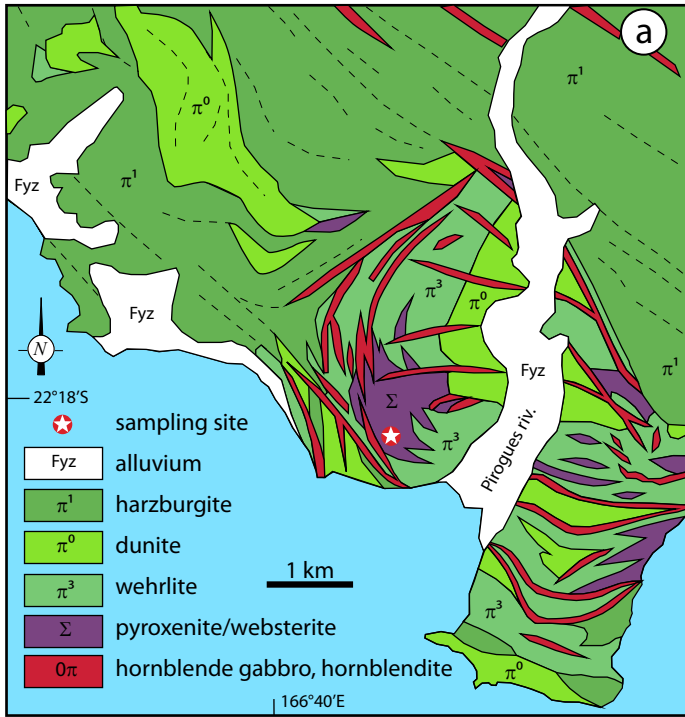
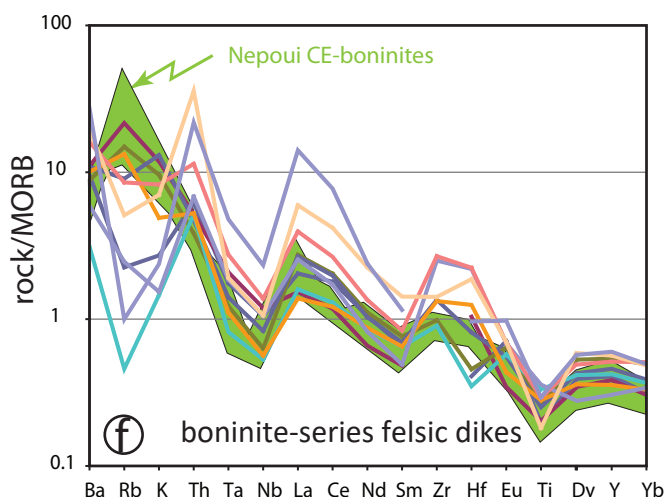
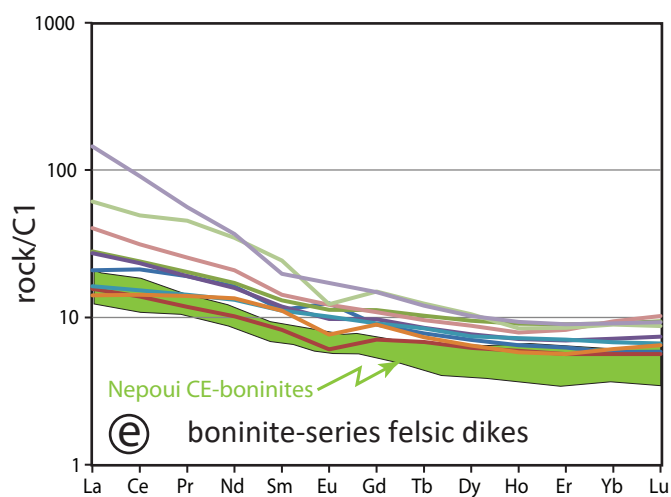
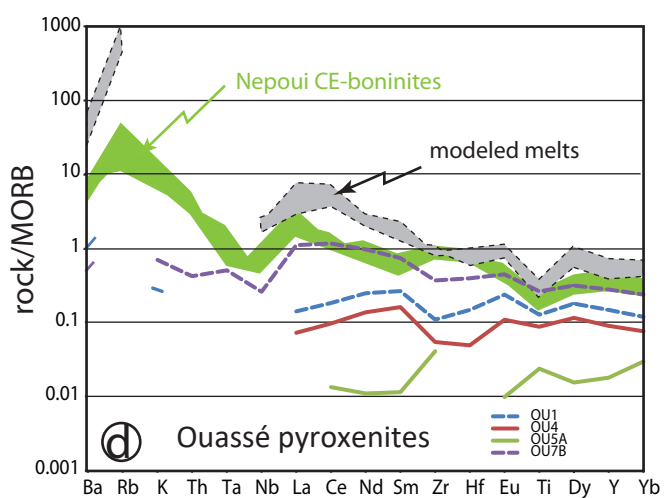
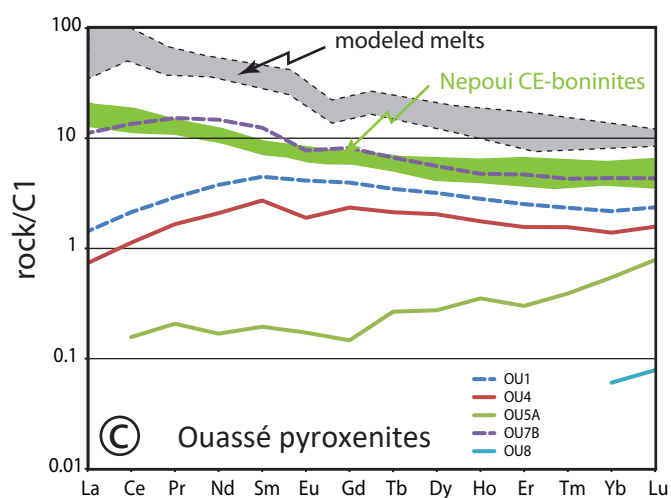
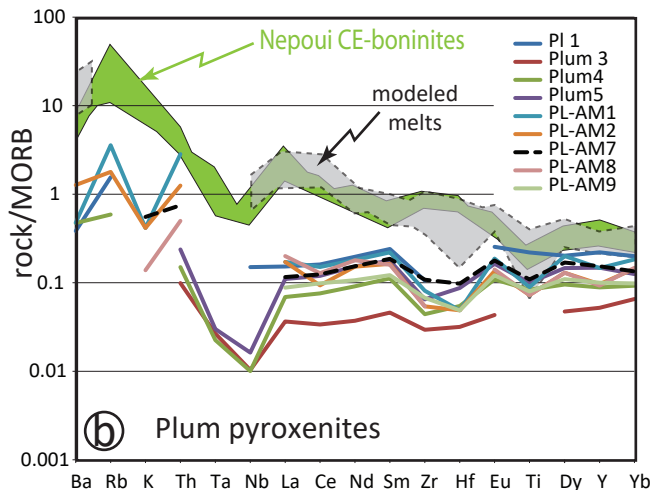
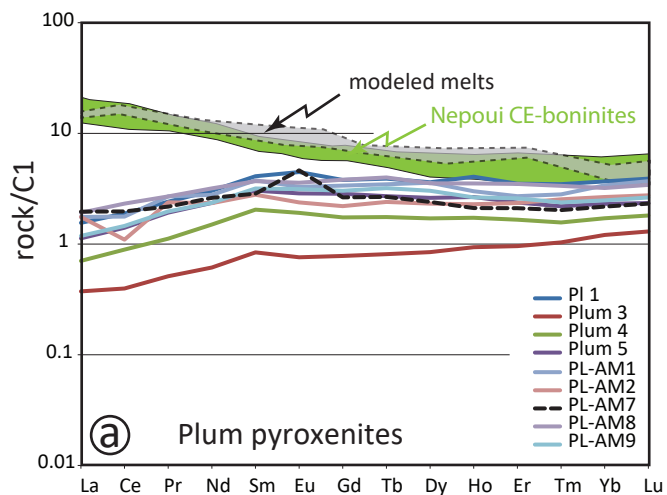


Fig. 1. Geological sketch map of New Caledonia. BSZ: Belep Shear Zone, HC: Humboldt Corridor, BTF: Bogota Transform Fault. Inset, LHR: Lord Howe Ridge, NR: Norfolk Ridge, LR: Loyalty Ridge, HP: Hukurangi Plateau.







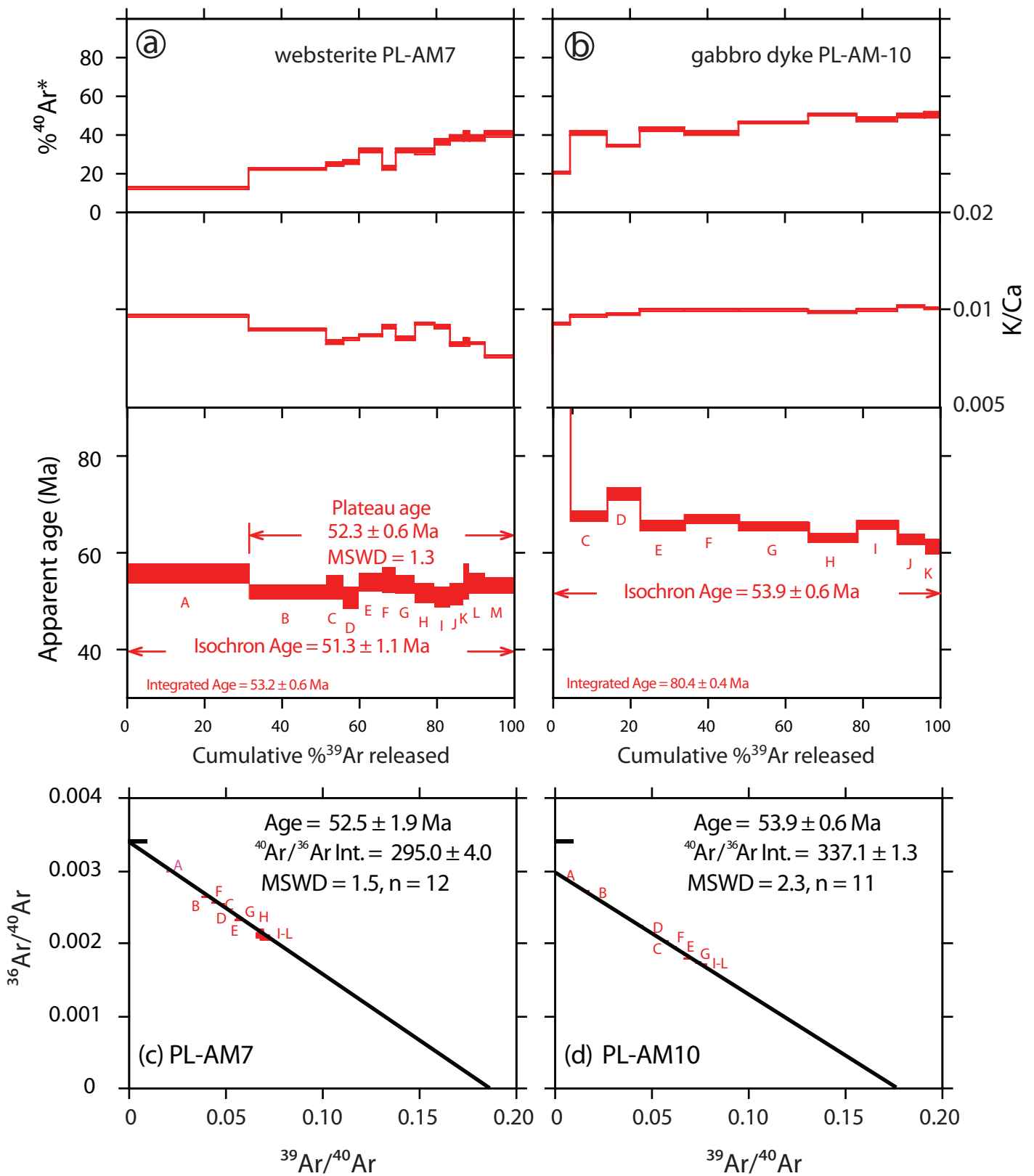
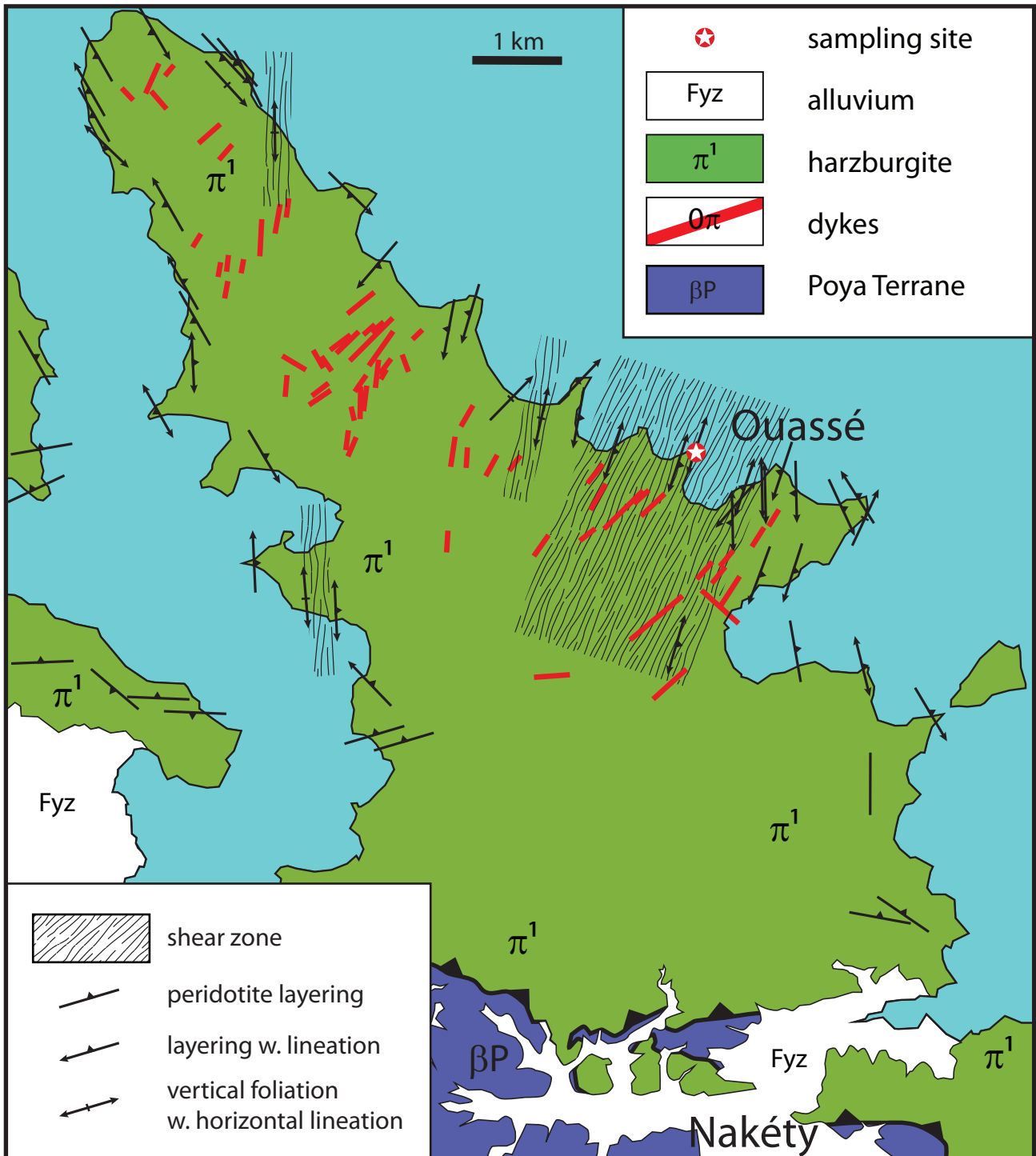


Fig. 3. ⁴⁰Ar/³⁹Ar data for the amphiboles of the Plum samples; PL-AM7 pyroxenite and PL-AM10 hornblende-gabbro dyke. (a, b) show the age spectra, K/Ca and radiogenic yield diagrams obtained by the incremental heating experiments and (c,d) are the isochron diagrams. The isochron ages are the preferred cooling ages of the samples.



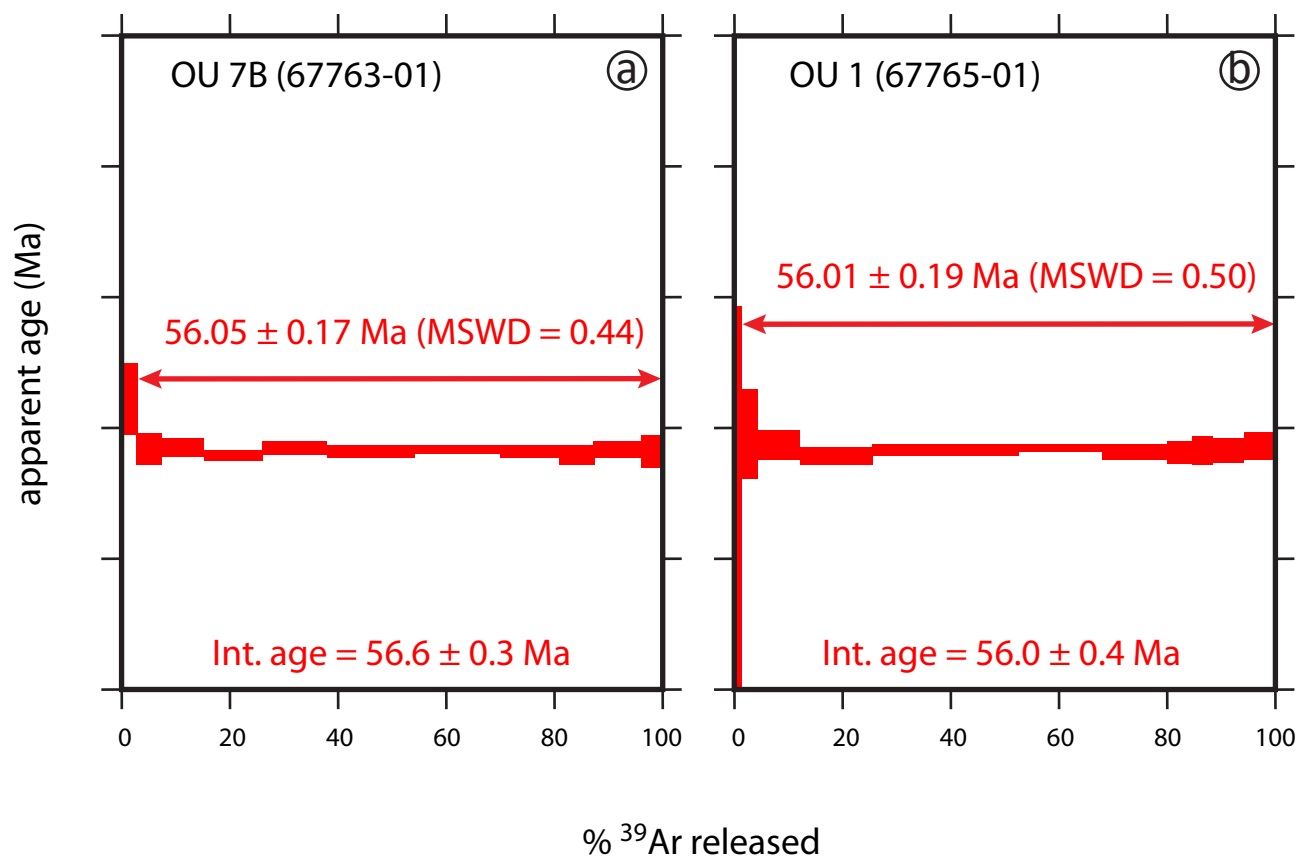
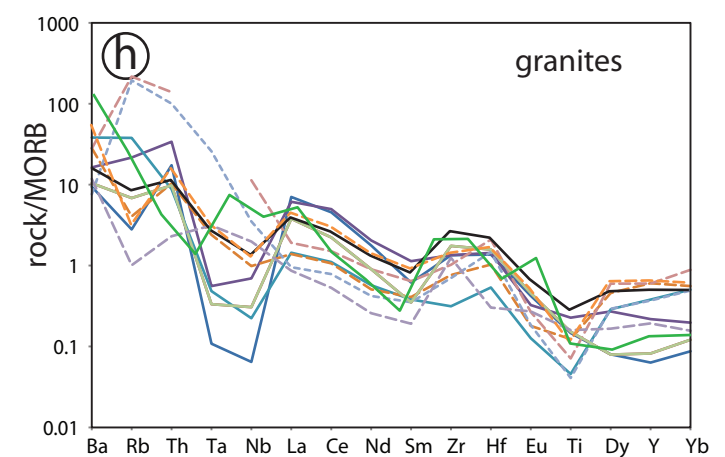
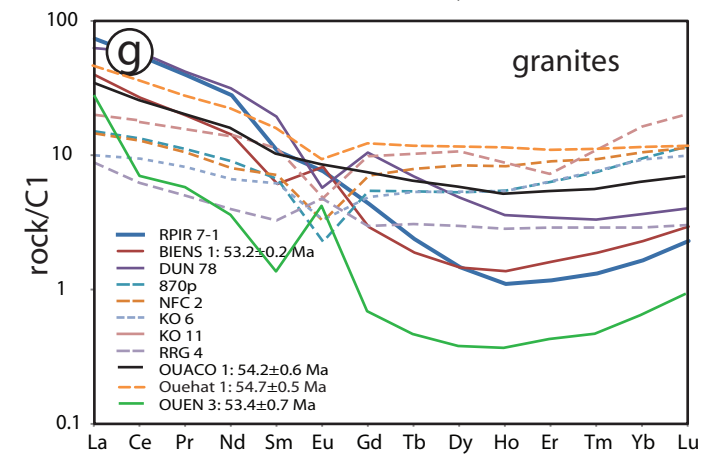
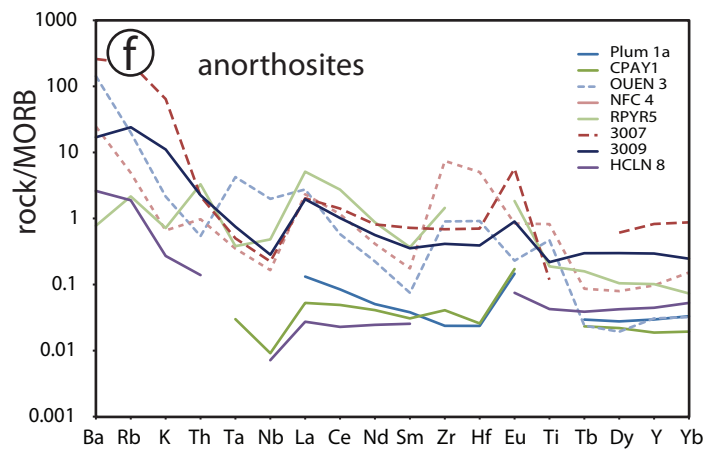
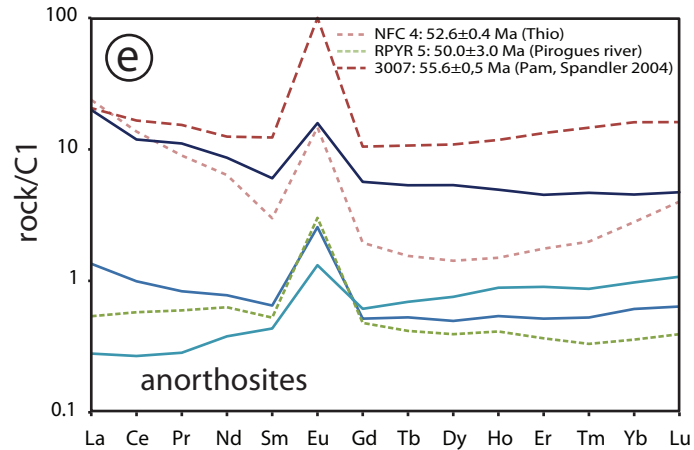
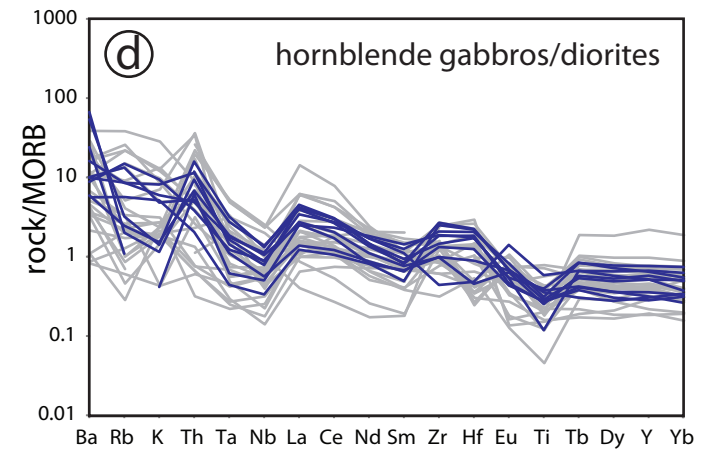
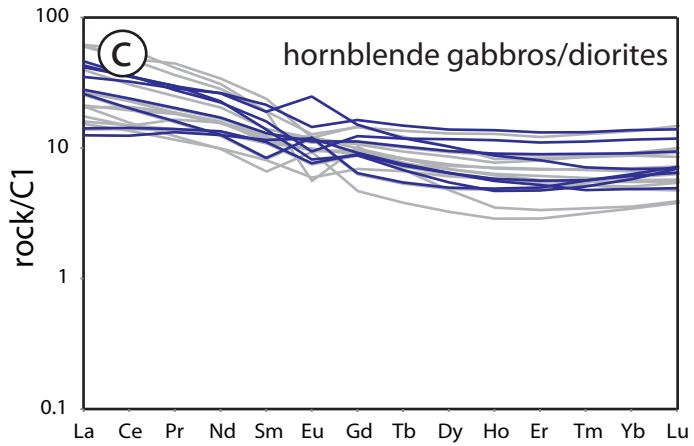
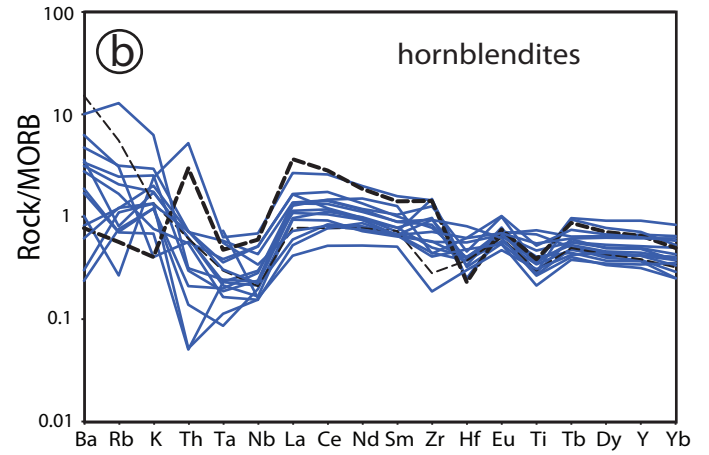
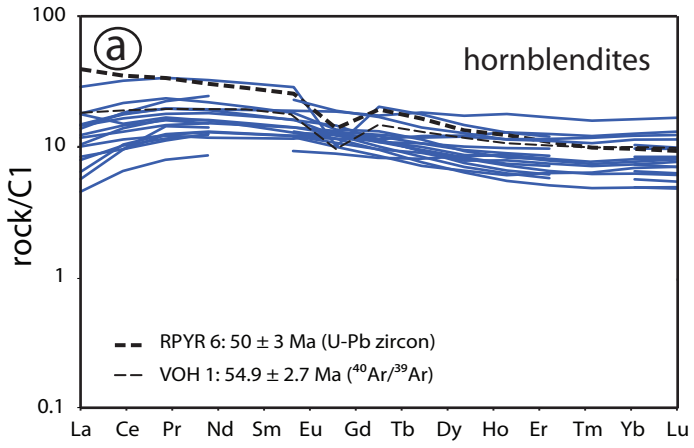


Fig. 4: $^{40}\text{Ar}/^{39}\text{Ar}$ degassing diagram for the amphiboles of Ouassé pyroxenite dykelets.



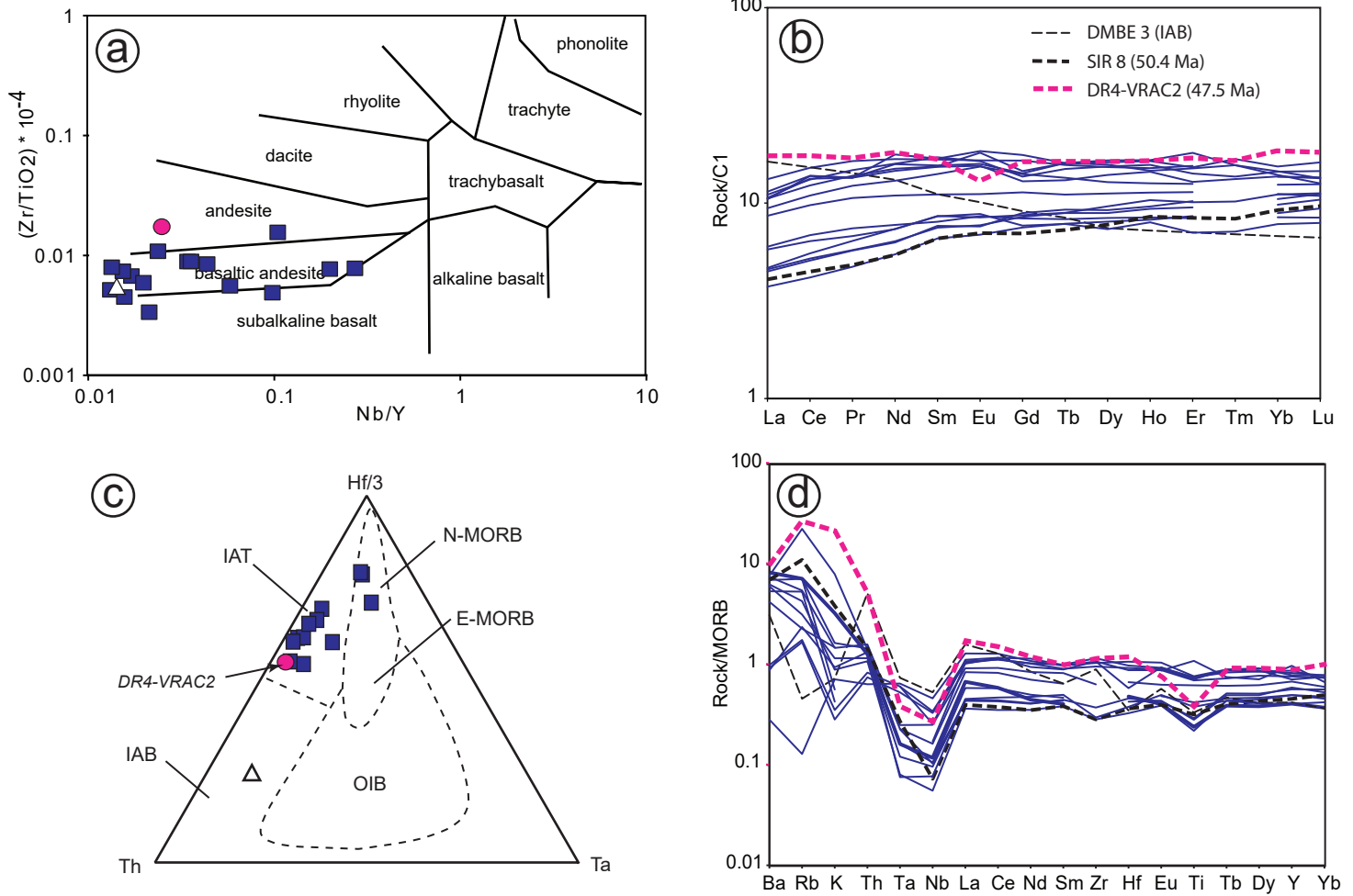


Fig. 6. Whole-rock geochemical features of dolerite dykes. (a) Classification diagram of Winchester and Floyd (1977), most samples plot in the field of andesitic basalts; the white triangle represents the sample #DMBE 3 (IAB-like) and the purple dot represents the andesite dredge sample from IPOD program. (b) Chondrite-normalized REE diagram (Evensen et al., 1978); the bold dotted pattern represents #SIR 8 dated at 50.4 Ma ($^{40}Ar/^{39}Ar$). (c) Ternary Hf-Th-Ta "discriminant" diagram (Wood 1980; Vermeesch, 2006) most samples plot in the island-arc tholeiite (IAT) domain. (d) MORB-normalized REE and trace elements spiderdiagram (Sun and McDonough, 1989) same patterns as (b).

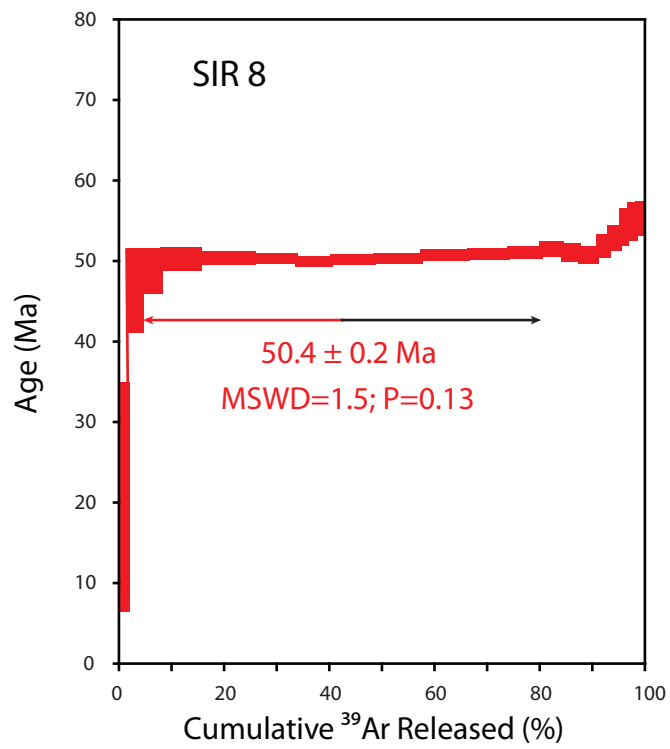
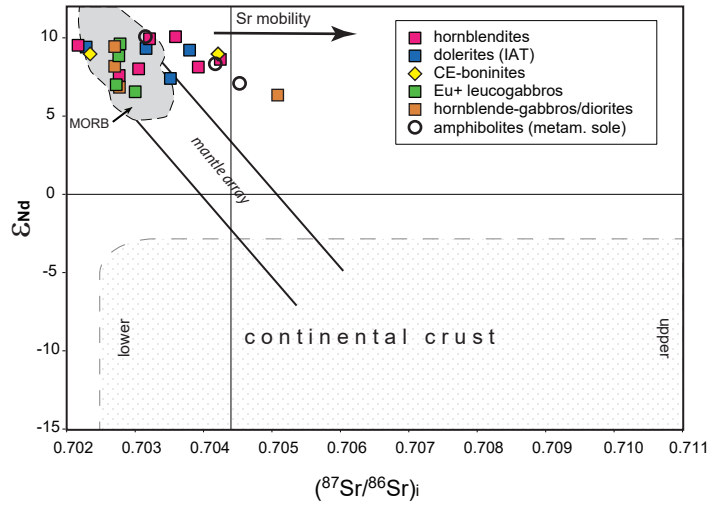
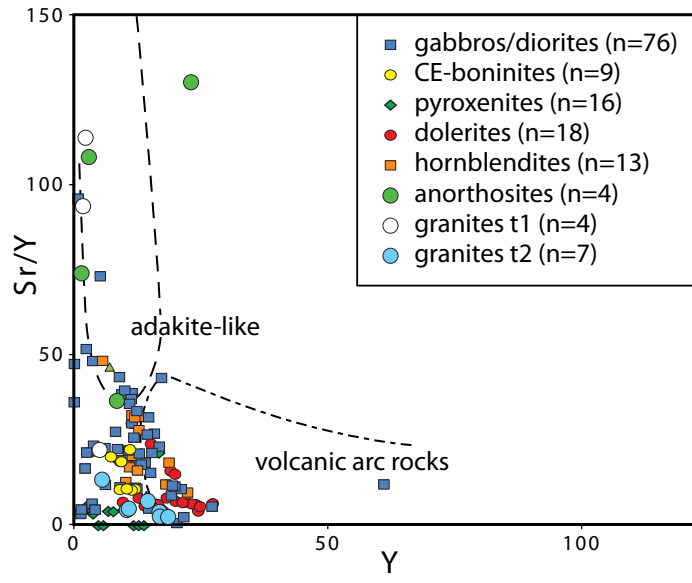


Fig. 7. $^{40}\text{Ar}/^{39}\text{Ar}$ degassing diagram for the dolerite SIR 8 (Si Reis nickel mine, Boulinda Massif).





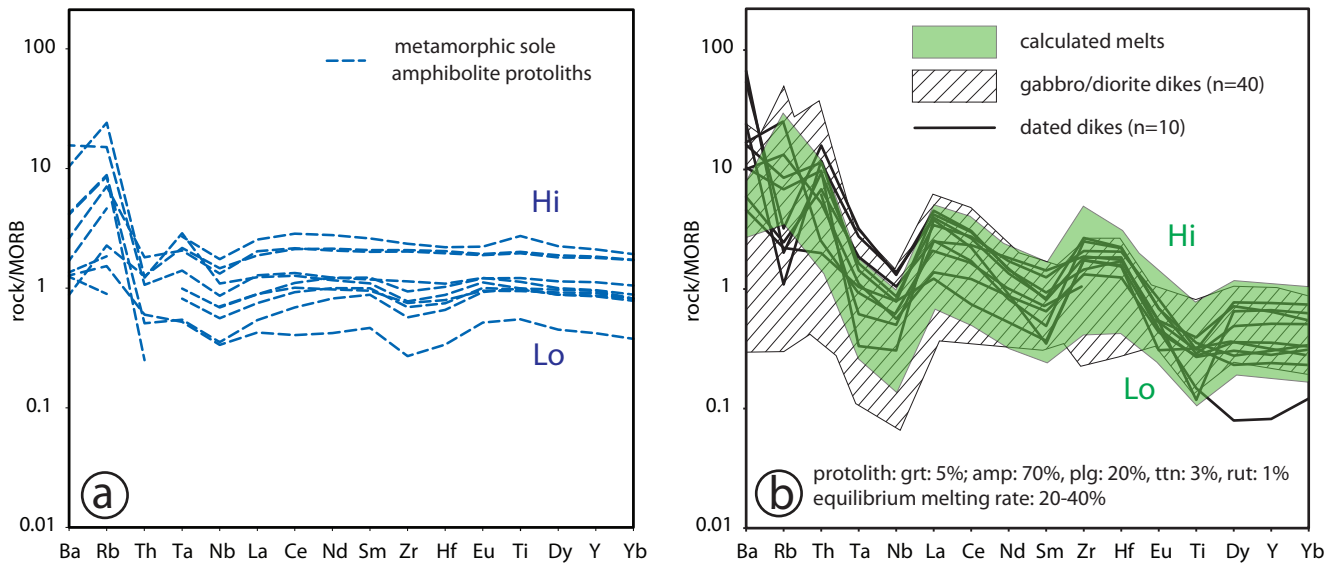


Fig. 8: Slab melt modeling. (a) REE-Tr spiderdiagrams of HT amphibolites of the metamorphic sole (Cluzel et al., 2012) taken as a protolith for slab melt modeling. (b) Spiderdiagrams of modeled melts through 20-40% partial melting of amphibolites (green array) compared with the array of hornblende-diorite/gabbro dykes (hatched) and dated dykes (black lines).

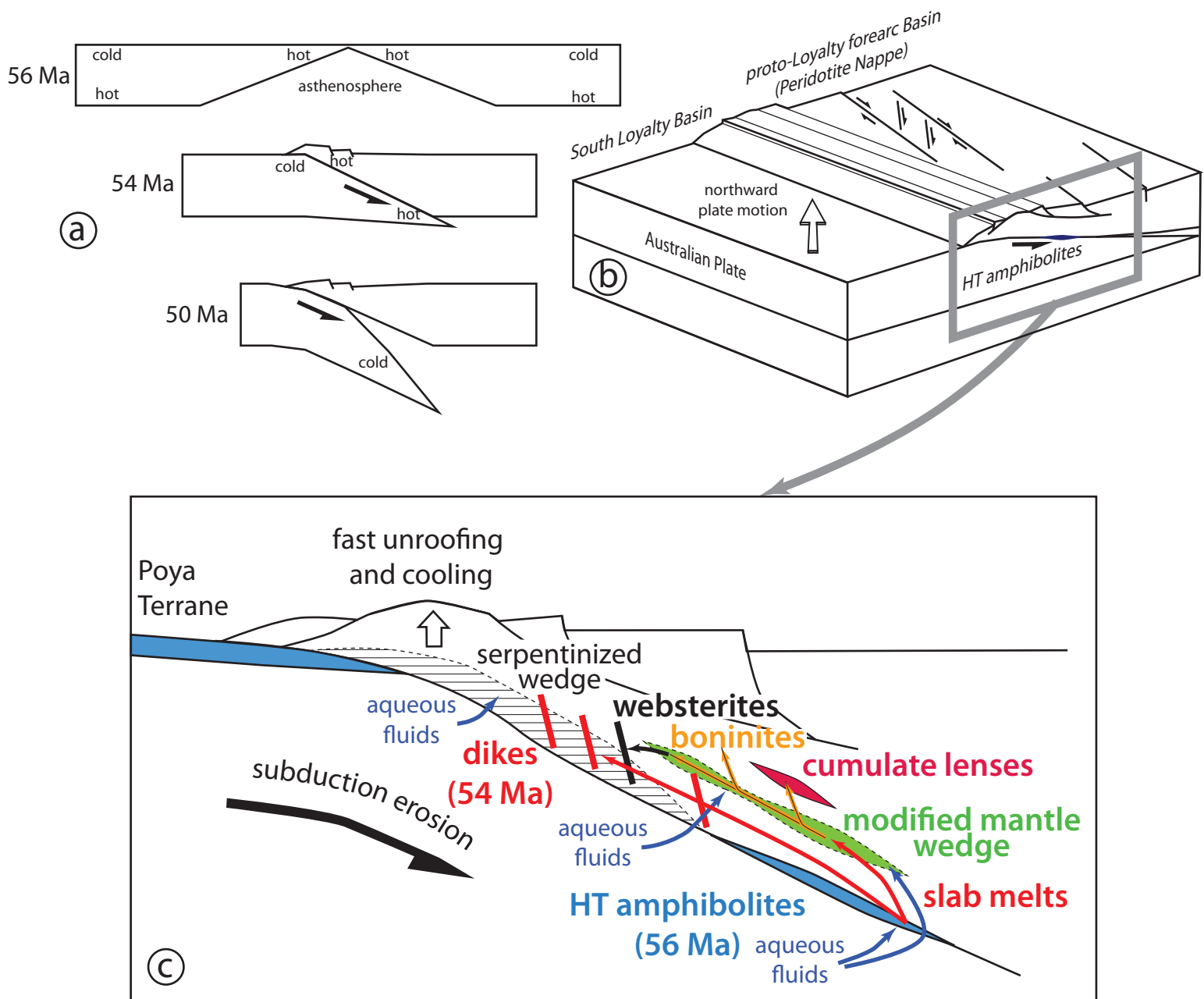
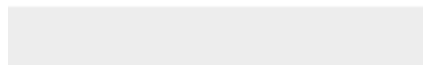


Fig. 14: (a) simplified evolution of the Eocene subduction zone, from inception to slab steepening. (b) sketch model of shallow-dipping hot subduction to show the location of high-temperature amphibolites of the metamorphic sole and relationships between oblique subduction, development of shear zones and exhumation of the mantle wedge. (c) sketch diagram of fluid and melt circulations in the mantle wedge between 55 and 50 Ma.



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Abstract

New Caledonia Ophiolite is crosscut by coarse to medium grained pyroxenites and hornblende gabbros/diorites dykes intruded between 55.5Ma and 50Ma (U-Pb zircon and $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende), while finer-grained dolerites of tholeiitic affinity are younger (50-47 Ma). Production of hornblende-gabbros/diorites was modelled by moderate degree (20-40%) of partial melting of the HT amphibolites of the metamorphic sole. End-member compositions, hornblendites and anorthosites, resulted from solid-state phase segregation of crystal mushes within tectonically active magmatic conduits. Cascade reactions of slab melts with mantle wedge peridotites successively formed clinoenstatite-boninite magmas, which fed gabbronorite cumulate lenses at the mantle-crust transition, in turn clinoenstatite-boninite melts reacted with peridotites to form websterites. The youngest magmas of tholeiitic affinity, appeared about 6 Ma after subduction inception when the cooler subducting slab plunged more steeply. Incipient slab retreat allowed corner flow, triggering low pressure hydrous melting of the uplifted asthenosphere. The early stages of forearc magmatism were closely associated with transcurrent shear zones, which recorded oblique subduction inception. The lower Eocene tectonic and magmatic features of the New Caledonia ophiolite witness the existence of a north- or northeast-dipping hot (forced) subduction zone in the Southwest Pacific, which appears notably distinct from the slightly younger west-dipping Izu-Bonin-Marianna cold (spontaneous) subduction system.