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1 **Variscan eclogites from the Argentera-Mercantour Massif (External Crystalline**
2 **Massifs, SW Alps): a dismembered cryptic suture zone.**

3
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13
14
15 **Abstract**

16 We document structural, geochemical, petrological and ⁴⁰Ar/³⁹Ar geochronological data performed on Variscan
17 eclogites from the Argentera-Mercantour Massif, southwestern Alps. Based on high-resolution field mapping,
18 we present new eclogite occurrences and discuss the relationships between eclogites and surrounding
19 migmatites. We recognized for the first time preserved eclogite facies assemblages. Trace elements and REE
20 patterns establish that eclogites protoliths are MORBs contaminated by continental crust. Standard
21 thermobarometry and thermodynamic modeling are consistent with P-T values of 640–740°C for 1.5 ± 0.25
22 GPa, coherent with paleo-geotherms predicted for warm subduction of oceanic crust. We interpret these
23 eclogites as a dismembered cryptic suture zone. ⁴⁰Ar/³⁹Ar dating on amphiboles yields an age of 339.7 ± 12 Ma
24 for eclogite retrogression under amphibolite facies conditions. All these data are combined to link the Argentera-
25 Mercantour Massif in the tectonic framework of both pre-Mesozoic Alpine basement and European southern
26 Variscides.

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28
29 **Keywords:** Argentera-Mercantour, southwestern Alps, Variscan eclogites, protoliths geochemistry,
30 metamorphic evolution, ⁴⁰Ar/³⁹Ar dating.

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32
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38 **Introduction**

39

40 High pressure (HP) metamorphic rocks exposed in mountain belts provide first-order constraints for addressing
41 evolution of orogens (Ernst and Liou 2008) and mafic eclogites are of particular interest because they are records
42 of old subduction zones (Miyashiro 1961; Ernst 1971). In the European Variscan belt, HP metamorphic rocks are
43 widespread (Medaris et al. 1995; Ballèvre et al. 2009; Schulmann et al. 2009; Martínez Catalán et al. 2009;
44 Lardeaux et al. 2014) and have been extensively considered to provide information on pre-continental collision
45 plate tectonics framework (Matte, 2001; Schulmann et al. 2005, 2014; Franke et al. 2017).

46 Mafic eclogites are also exposed in Variscan basement slices preserved in the Alpine belt (Fig. 1 and “Online
47 Resource 10”; Spalla and Marotta 2007; Spalla et al. 2014; Roda et al. 2019 with references therein). In the
48 External Crystalline Massifs (ECMs) of the western Alps, eclogites that retain protolith geochemical signature
49 and pre-alpine HP mineralogy are rare (von Raumer 1984; Latouche and Bogdanoff 1987; Bogdanoff et al.
50 1991; Ferrando et al. 2008; Rubatto et al. 2010). The Argentera-Mercantour Massif (Fig. 1) is the southernmost
51 of the ECMs, within which the metamorphic evolution of mafic eclogites remains poorly constrained. We
52 document here the discovery of preserved omphacite (jadeite contents up to 42 mol%) within the retrogressed
53 eclogites of this Massif. Our goal is to present and discuss structural, geochemical, mineralogical, petrological
54 and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological data on these Variscan HP metamorphic rocks.

55

56 **Geological setting**

57

58 The ECMs belong to the Helvetic-Dauphinois domain of the Alpine belt (Schmid et al. 2004), where Alpine
59 tectonic history has not significantly erased Variscan signatures (Von Raumer et al. 1999, 2009, 2013; Guillot
60 and Menot 2009; Compagnoni et al. 2010; Spalla et al. 2014; Fréville et al. 2018). The Argentera-Mercantour
61 Massif, located at the French-Italian border, is a piece of European continental crust shortened in response to late
62 Alpine collision (Polino et al. 1990; Bogdanoff et al. 2000; Schwartz et al. 2007; Schreiber et al. 2010).

63 Following the first investigations of Faure-Muret (1955), Malaroda and Schiavinato (1958, 1960) and Malaroda
64 et al. (1970), this massif was systematically divided into two main units (Bogdanoff et al. 1991; Rubatto et al.
65 2001; Compagnoni et al. 2010; Carosi et al. 2016) separated by a steeply dipping, NW-SE striking, km-scale
66 shear zone, the « Ferriere-Mollières Shear Zone » (FMSZ, Fig. 2), also known as « Valletta Shear Zone ». The
67 FMSZ, recognized since Faure-Muret (1955), is a dextral transpressive shear zone active in Late Carboniferous

68 times (i.e. 330–310 Ma, Musumeci and Colombo 2002; Corsini et al. 2004; Simonetti et al. 2018) and
69 significantly re-activated during Alpine times (i.e. 27–21 Ma, Corsini et al. 2004; Sanchez et al. 2011). HP/HT
70 metamorphic rocks are described in the western and eastern complexes (Faure-Muret 1955; Malaroda et al.
71 1970), which consist of Variscan high-grade migmatites derived from meta-sediments (Bortolami and Sacchi
72 1968; Blasi 1971; Malaroda et al. 1970) or from meta-granitoids (Bogdanoff and Ploquin 1980; Bogdanoff et al.
73 1991). Upper Ordovician to Lower Silurian ages have been obtained for magmatic protoliths (Rubatto et al.
74 2001).

75 The main pre-Alpine structure is a pervasive steeply dipping regional-scale foliation defined by a shape
76 preferred orientation of biotite, feldspar and sillimanite, and by alternating quartz + feldspars + biotite ± garnet-
77 bearing leucosomes and biotite-rich melanosomes.

78 In the eastern unit post-kinematic cordierite-bearing leucocratic diatexites are widespread (Blasi and Schiavinato
79 1968). At the regional scale, coarse-grained undeformed cordierite-bearing melts clearly crosscut the main
80 foliation. In both complexes centimeter-scale cordierite-rich leucosomes intersect and progressively erase the
81 main foliation in biotite and sillimanite-bearing migmatites. Two main partial melting events can be inferred: the
82 first syntectonic episode (M1) developed in equilibrium with biotite, garnet and sillimanite, and the second (M2),
83 clearly post kinematic, in equilibrium with cordierite, at lower pressure. The age of 323 ± 12 Ma obtained by
84 U/Pb on zircons dates the last regional migmatitic event (Rubatto et al. 2001).

85 Within the eastern complex, a large granitic stock (Fig. 2) intersects the foliation (Boucarut 1967; Ferrara and
86 Malaroda 1969). This granite is dated at 292 ± 10 Ma by Ferrara and Malaroda (1969) and in the range 299–296
87 Ma by Corsini et al. (2004), and belongs to the late Carboniferous Fe-rich granitoids (Debon and Lemmet 1999)
88 in the EMCs of the western Alps.

89 The exhumation of the Argentera-Mercantour Variscan crust is sealed by impure sandstones and black shales,
90 rich in upper Pennsylvanian vegetal fossils, filling intra-mountain basins bordered by strike-slip faults (Corsini
91 and Faure-Muret 1946, 1951; Faure-Muret and Fallot 1955).

92 During Alpine collision the Argentera-Mercantour Massif was metamorphosed under upper greenschist facies
93 conditions (Corsini et al. 2004; Lardeaux 2014a; Filippi et al. 2019). Alpine deformation is concentrated in
94 ductile shear zones marked by alternating fine grained recrystallized quartz + albite and muscovite + chlorite ±
95 titanite-rich layers (Sanchez et al. 2010, 2011).

96
97

98 **Eclogites in the Argentera-Mercantour Massif: Localized or diffuse occurrences?**

99

100 In the Ibero-Armorican arc and French Massif Central, mafic eclogites are localized in specific horizons
101 (Leptyno-Amphibolite Complexes) in association with ultramafic rocks and HP felsic rocks (Martínez Catalán et
102 al. 2007; Ballèvre et al. 2009; Faure et al. 2009; Lardeaux 2014b). These horizons are localized at the base of the
103 allochthonous nappes thrust upon low-pressure units (Matte and Burg 1981; Ledru et al. 1989, 2001; Lardeaux
104 et al. 2001; Berger et al. 2010). In Sardinia or Maures-Tanneron Massif (southern European Variscides) eclogites
105 are recognized as lenses dispersed within migmatites (Carmignani et al. 1994; Franceschelli et al. 2007; Cruciani
106 et al. 2011; Schneider et al. 2014).

107 In the Argentera-Mercantour Massif, the presence of rare eclogites was first reported by Faure-Muret (1955) and
108 other limited occurrences identified later (Malaroda et al. 1970; Latouche and Bogdanoff 1987; Colombo et al.
109 1994; Rubatto et al. 2001). Ages of 459 ± 4 Ma and 486 ± 7 Ma have been obtained by U-Pb dating on zircons
110 for protoliths of HP mafic rocks (Rubatto et al. 2001, 2010). We discovered a significant number of other
111 occurrences (Fig. 2) with specific characters: First, eclogites occur as poorly deformed lenses, centimetric to
112 metric thick, within surrounding migmatites (Fig. 3). Second, boudins of eclogites can be associated with mafic
113 HP granulites and/or serpentinites. Up to now, an association with HP felsic granulites remains unknown. Third,
114 eclogites are severely retrogressed under amphibolite facies conditions: preserved eclogite facies assemblages
115 are exceptional (Fig. 4). Last, eclogites occur in both western and eastern units in all types of migmatites, i.e.
116 within biotite-sillimanite-garnet banded M1 migmatites as well as within leucocratic cordierite bearing M2
117 migmatites, but also within migmatitic orthogneisses and migmatitic amphibolites.

118 Typical field relationships are illustrated by two lithological and structural maps (Figs. 5 and 6).

119 In this massif, mafic granulites have been studied in detail (Ferrando et al. 2008; Rubatto et al. 2001, 2010), but
120 petrological and structural analyses on eclogites remain scarce, probably because deciphering preserved eclogite
121 facies assemblages is rare.

122

123

124 **Geochemistry of eclogites**

125

126 Analytical conditions and results are given in “Online Resource 1”.

127

128 **Major elements**

129

130 Our samples, displaying mid ocean ridge basalt (N-MORB) average composition (McKenzie and O'Nions
131 1991), show a weak depletion in CaO and TiO₂ and an enrichment in K₂O. On the Al₂O₃ vs. TiO₂ diagram
132 (Konzett et al. 2012), they tend to lie in the basalt field and for a few samples in plagioclase-rich cumulates (Fig.
133 7), suggesting that their protoliths were basalts and gabbros.

134 In the Total Alkali versus Silica diagram (Online Resource 2) SiO₂ content ranges from 40.51 to 52.01 wt% and
135 our sample compositions are mainly in the field of basalts, while in the AFM diagram, they plot mainly in the
136 fields of tholeiitic, and for a restricted number of samples in the calc-alkaline volcanic series (Fig. 7).

137 However, major elements are mobile during metamorphism (Winchester and Floyd 1977). Taking into account
138 that surrounding migmatites have undergone dehydration reactions coeval with retrogression of eclogites, the
139 latter could have experienced chemical contamination from outside (see Štípská et al. 2014 for example).
140 Consequently K₂O enrichment from surrounding migmatites is possible. We therefore consider the least mobile
141 elements during alteration and metamorphism (i.e. HFS elements such as Sc, Y, Zr, Hf, Ti, Nb and Ta, or REEs)
142 when using chemical diagrams for protoliths characterization.

143

144 **Trace elements and REE**

145

146 The eclogites have typical N-MORB compositions in the Zr-Ti-Nb-Y (Pearce 1996) or in the Y-Nb-Zr
147 (Meschede 1986) diagrams, except for two samples displaying within-plate chemical signatures and
148 corresponding to a foidite and a microbasalt (Fig. 8a, see also Online Resource 2).

149 In discriminant diagrams (Fig. 8b, and Online Resources 3,4 and 5) sample compositions plot in the field of
150 tholeiitic basalt and MORB, whereas a few of them show affinity with calc-alkaline to alkaline basalts. This can
151 be interpreted as the consequence of various tectonic contexts of protoliths formation or as the result of chemical
152 contamination during protoliths formation and/or evolution. To face these uncertainties, the best tools are the
153 diagrams proposed by Pearce (2008) or the chondrite normalized Th vs. Nb (normalization according to Sun and
154 McDonough 1989) proposed by Saccani (2015), within which the effects of crustal contamination on fractional
155 crystallization and on the MORB-OIB array can be depicted. In these diagrams, a significant number of our
156 samples plot outside the MORB-OIB array suggesting that these rocks have been contaminated by crustal
157 components (Fig. 8c, see also Online Resource 6).

158 The Nb/La ratio is regarded as an index of crustal contamination of magmas (Thompson et al. 1983). For our
159 eclogites, this ratio ranges between 0.20 and 1.71 with a mean value of 0.7. For comparison, MORB and OIB
160 typically show values of 0.93 and 1.3 respectively (Sun and McDonough 1989), while continental flood basalts
161 display values of 0.14–1.4 (Thompson et al. 1983). This indicates that our samples have undergone crustal
162 contamination.

163 Contamination is also suggested by low and variable Zr/Hf ratios (mean value of 34.11) and Nb/Ta ratios (mean
164 value of 12.47) in comparison with those of N- and E-MORB (respectively 36 and 17 according to Sun and
165 McDonough 1989). Moreover, MORB and OIB have uniform value of 25 ± 5 for their Ce/Pb ratio whereas the
166 mean value for continental crust is around 4 (Hofmann et al. 1986). The Ce/Pb value of our samples is in the
167 range of 0.03-15.7 (with a mean value of 5.12) and thus compatible with crustal contamination of MORB
168 basalts.

169 According to Hofmann (2003), the Nb/U and the Pb/Nd ratios are indicators for crustal contamination. Nb/U
170 ratio of 47 ± 11 is typical for MORBs, while the mean value of continental crust is 8 (Rudnick and Fountain
171 1995). This ratio varies from 2 to 47.8 with a mean value of 12.6 for the studied eclogites. The Pb/Nd ratios are
172 of 0.04, 0.07, 0.08, and 0.63 for N-MORB, E-MORB, OIB, and continental crust respectively (Hofmann 2003).
173 Our eclogites present a Pb/Nd ratio ranging between 0.14 and 1.81 with a mean value of 0.41. This confirms
174 crustal contamination.

175 The total REE content indicates that most samples show an enrichment, 10–40 times greater than average N-
176 MORB. The REE patterns of the Argentera-Mercantour eclogites are presented in Fig. 9 and the patterns of OIB,
177 N-MORB and E-MORB are shown for comparison.

178 Primitive mantle normalized trace element pattern shows contrasting trends: a first group of samples show
179 enrichment in LREE with respect to HREE, whereas the second is characterized by slightly depletion,
180 compatible with E-MORB and N-MORB trends, respectively. Some samples are characterized by strong Th
181 depletion, similar to average N-MORB values, whereas a few others have Th positive anomaly. All samples
182 show positive Rb, Pb and U anomalies and trends with Sr and Ba depletions, and most of them Nb, Ta and Mo
183 depletions. These trends are relevant markers for contaminated basaltic sources (Donnelly et al. 2004). Finally,
184 the $(La/Sm)_N$ ratio is highly variable (0.42-3.09), indicating various degrees of fractionation (Carlson 2003) .

185

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188 **Petrography of eclogites**

189

190 The Argentera-Mercantour eclogites experienced severe retrogression and eclogite facies assemblage is rarely
191 preserved. The cores of the boudins consist of fine-grained rocks with reddish millimeter to centimeter-sized
192 garnets but their margins are totally transformed into garnet-free amphibolites.

193 The peculiar features of these eclogites are the occurrence of clinopyroxene-plagioclase symplectitic domains
194 and of plagioclase-brown/green amphibole symplectitic coronas around garnets (Fig. 10a, b, c, d and f). The
195 development of clinopyroxene-plagioclase symplectites results from omphacite destabilization and is a
196 discontinuous precipitation explained in terms of decreasing pressure with or without temperature increase
197 (Boland and Van Roermund 1983; Joanny et al. 1989, 1991). Some of our samples show an evolution of the
198 symplectite texture, with thinner clinopyroxene lamellae developed at the expense of the coarser ones (Fig. 10b).

199 Garnet occurs as grains up to 5 millimeter in size containing inclusions of quartz, amphibole, rutile, zircon and in
200 some samples small sized brown biotite, epidote, ilmenite, apatite and monazite (Fig. 10e). In one sample (Lac
201 Niré), an inclusion of omphacite, associated with amphibole, was observed within garnet.

202 Amphibole, the most abundant phase, occurs in different textural sites: inclusions within garnets, large green to
203 light-brown amphibole grains in the matrix, brown amphibole rims near the plagioclase-clinopyroxene
204 symplectites, discontinuous droplets in association with plagioclase within coronas around garnets, large
205 poikiloblastic green-brown grains, in association with biotite, fully replacing the symplectites or invading totally
206 the garnets grains, and pale green amphibole filling late veins.

207 Plagioclase occurs together with clinopyroxene in symplectites replacing omphacite, but also together with the
208 dark-brown amphibole rimming garnet or amphibole 1, close to the symplectites or together with droplets of
209 amphibole around garnet, and in late veins associated with titanite, chlorite, white mica, biotite and pale green
210 amphibole.

211 In the matrix, rutile grains are usually rimmed by ilmenite. Quartz is common and biotite and K-feldspar are
212 interstitial with respect to garnet, clinopyroxene and amphibole 1 and 2. In the most retrogressed samples,
213 chlorite is developed at the expense of biotite and amphibole as well as in the fractures of garnet.

214

215

216

217

218 **Mineral chemistry**

219

220 Mineral analyses were performed on 3 samples (Lac Long, Lac Niré and Pas des Ladres) in the French part of
221 the massif, and 3 samples (upper Valasco Valley) in the Italian part. Analytical conditions and results are
222 presented in “Online Resource 7”.

223 Garnet from French samples is almandine-rich (almandine 49–56 mol%) and spessartine-poor with pyrope,
224 grossular contents ranging from 20 to 32 and 17 to 24 mol% respectively. In the diagram of Coleman et al.
225 (1965), garnets are distributed between the fields of B and C eclogite types (Fig. 11b). The zoning pattern is
226 limited to the mineral rims, with an increase in almandine content correlated to a decrease of pyrope and
227 grossular contents.

228 In Valasco Valley, garnet is Fe-rich (almandine 61–54 mol%) with low spessartine (3–6 mol% at core, 1–3
229 mol% at rim), pyrope and grossular contents 8–12 mol% and 21–32 mol%, respectively. They plot in the field of
230 type C eclogites (Fig. 11b) showing a constant pyrope content and grossular enrichment from core to rim. Indeed
231 Mn and Fe²⁺ decrease from 0.16 to 0.01 and 1.83 to 1.70 apfu, respectively, whereas Ca slightly increases from
232 0.75 to 0.96 apfu, from core to rim.

233 Clinopyroxene was identified in different sites. In the Lac Niré sample, the clinopyroxene inclusion within
234 garnet is omphacite with jadeite content around 32 mol%. In Lac Niré as well as in Pas des Ladres eclogites,
235 clinopyroxene 1 is replaced by clinopyroxene 2 + plagioclase symplectites. In the coarser lamellas, the jadeite
236 component is 42 mol%, while the thinner ones have jadeite contents of 15–27 mol% (Fig. 11a).

237 In Valasco Valley, clinopyroxene 1 is not preserved, but its composition has been estimated, by mass balance
238 calculations of symplectite domains, at a jadeite content from 24 to 38 mol%. Pyroxene in symplectite (Cpx 2) is
239 always diopside with jadeite content < 5 mol% in all localities.

240 Plagioclase lamellae from the symplectites are oligoclase (anorthite 20–32), while plagioclase forming coronitic
241 shells around garnet in equilibrium with amphiboles is andesine (Fig. 11c). Plagioclase in veins with actinolite is
242 anorthite 8–13.

243 Amphiboles, from Pas des Ladres and Lac Niré are calcic (Fig. 11d) according to the classification of Leake et
244 al. (2004) and Hawthorne et al. (2012). The amphiboles in inclusions within garnets are magnesio-hornblendes
245 with Mg content of about 0.22 to 0.31 apfu. The amphiboles from coronas around garnet and the amphiboles
246 replacing clinopyroxene of the symplectites, or invading garnets are magnesio-hornblendes with Mg content of
247 about 0.30 to 0.38 apfu.

248 In Valasco Valley, amphibole 1 and 2 are mostly pargasite. Amphibole 3a is ferro-pargasite, amphibole 3b is
249 mostly ferro-edenite or magnesio-hornblende. Amphibole 4 is actinolite. Amphibole 1 and 2 have Al^{IV} between
250 1.53–2.04 apfu and Al^{VI} between 0.37 and 0.74 apfu. Ti content in amphibole 1 varies between 0.12 and 0.20
251 whereas amphibole 2 is Ti-richer (from 0.20 to 0.31 apfu). Na content decreases from 0.55 apfu in amphibole 1
252 to 0.49 apfu in amphibole 2.
253 Amphibole 3a, in rims around garnet, is particularly Al-rich (Al_{tot} from 2.55 to 3.74 apfu, clustering at 3.06
254 apfu), with XMg < 0.5 and Ti lower than 0.14 apfu. Amphibole 3b, mainly replacing clinopyroxene, has Al_{tot}
255 between 0.75 and 2.07 apfu and Ti lower than 0.07 apfu. Amphibole 4 has Al_{tot} < 0.70 apfu and Ti < 0.04 apfu
256 everywhere.

257

258 **Metamorphic history and P-T evolution**

259

260 The Argentera-Mercantour eclogites preserve evidences of a first eclogitic stage documented by the occurrence
261 of almandine-rich garnet and by the preservation of omphacite (jadeite 42 mol%) within symplectites and as
262 inclusion within garnet (jadeite 32 mol%). Mineral inclusions in garnet suggest that the stable mineral
263 assemblage during this stage was garnet + omphacite + calcic amphibole + quartz + rutile + zircon ± epidote ±
264 biotite.

265 A post-eclogitic stage is evidenced by the development of clinopyroxene (jadeite 27-5 mol%)-plagioclase
266 symplectites at the expense of omphacites and plagioclase-calcic amphibole-ilmenite coronas around garnets.
267 Rutile is replaced by ilmenite in the matrix. Garnet rims are in equilibrium with amphibole-plagioclase coronas
268 containaing in some samples also diopside. This last assemblage is typical for granulite-facies conditions.

269 An amphibolite-facies stage is constrained by the replacement of symplectitic clinopyroxenes by amphiboles and
270 the development of poikiloblastic amphiboles, in association with biotites and ilmenite, at the expense of all
271 previous minerals.

272 A late stage is documented by formation of micro-cracks filled with albite, actinolite, titanite and chlorite and by
273 replacement of biotites and calcic amphiboles by chlorite. The replacement of ilmenite by titanite can also be
274 related to this late stage (greenschist facies conditions).

275 A first approach for estimation of eclogitic P-T conditions is the use of experimentally determined domains for
276 the observed mineralogy. For samples with chemical compositions of N-MORB, we use the grid determined by
277 Schmidt and Poli (1998). With H₂O in excess, the association garnet-omphacite-amphibole-quartz is stable in the

278 range 600–700°C for 1.4–2.4 GPa. This estimation is in agreement with the field of amphibole bearing eclogite
279 facies for basaltic compositions (Ernst and Liou 2008)

280 Thermobarometry is a second approach and Fe²⁺-Mg exchange between garnet and clinopyroxene is well
281 adapted for eclogites. The calibration established by Krogh-Ravna (2000) integrates the effects of Ca and Mn
282 components and requires clinopyroxene with a sodic content in the range 0–0.51 to work. Because of severe
283 retrogression, this thermometer was only applied in the Lac Niré sample using the preserved inclusion of
284 omphacite (Na content of 0.33) within garnet. Application of this geothermometer, for pressure from 1.2 to 1.4
285 GPa, showed temperatures of 710 - 720°C.

286 The garnet–hornblende Fe-Mg thermometer (Graham and Powell 1984), applied to Lac Long samples showing
287 inclusions of magnesio-hornblende with rational grain boundaries with garnet and in equilibrium with quartz and
288 rutile, gives a range of 640–690 °C.

289 However, the reliability of Fe-Mg exchange thermometry is dependent on uncertainties in estimation of the Fe³⁺
290 /Fe^{tot} ratios in minerals (Carswell et al. 2000). As established by Štípská and Powell (2005), if Fe³⁺ content is not
291 taken into account for garnet-clinopyroxene thermometry the calculated temperatures are overestimated,
292 therefore the proposed values (710–720°C) are considered as maximum temperatures.

293 We used also the Zr-in-rutile thermometry (Watson et al. 2006, Thomkins et al. 2007) relevant for quartz and
294 zircon-bearing rocks. In the Lac Niré eclogite, the Zr content in rutiles in inclusion within eclogitic garnet is
295 between 221 and 458 ppm. Application of the Zr content in rutile thermometer for pressure of 1.2 and 1.5 GPa
296 showed temperatures of 634–694 and 646–706°C respectively.

297 The minimal pressure of the eclogitic stage is estimated at 1.4–1.6 GPa using the maximal values of the jadeite
298 content (32–42 mol%) of omphacites (Gasparik and Lindsley 1980; Holland 1990).

299 Ti-content in amphibole thermometer (Otten 1984), using the core compositions of in amphibole 1 grains, gives
300 temperatures of 680–715°C in Valasco Valley eclogites. Ernst and Liu (1998) calibration suggests T= 700–
301 740°C, using the same data, for which Al-content indicates P = 1.1–1.7 GPa.

302 In summary, application of various thermo-barometers, to all the studied samples, yields to a P-T range of 640–
303 740°C and 1.1–1.7 GPa (Online Resource 8).

304 Thermodynamic calculations were performed with the free energy minimization program THERIAK/DOMINO
305 (De Capitani and Petrakakis 2010, version 2015). Pseudosections were calculated for two samples from Valasco
306 Valley and Lac Long (Fig. 12). Compositions were chosen starting from the whole rock analysis, considering
307 H₂O in excess, and not considering the presence of a melt, as the mafic boudins are enclosed in migmatitic felsic

308 rock, but without evidence of internal melting. For Valasco Valley, we have removed K₂O from the composition,
309 as biotite and K-feldspar occur as interstitial grains or filling late veinlets suggesting that boudins were
310 contaminated during country rocks partial melting. For Lac Long sample, because biotite was observed in
311 equilibrium with eclogitic garnet, K₂O was considered in the system. Finally, we have evaluated the effect of O
312 and H₂O on our pseudosection, as suggested by Guiraud et al. (2001) and Rebay et al. (2010) to evaluate their
313 effect on the stability and preservation of assemblages. We used the following mineral a-x relations: plagioclase
314 (Holland & Powell 2003), garnet (White et al. 2007), ilmenite (White et al. 2000), amphibole (Diener et al.
315 2007), clinopyroxene (Green et al. 2007), orthopyroxene (White et al. 2002), biotite, (White et al. 2007) and
316 white mica (Coggon and Holland 2002). The thermodynamic database is ds55 (Holland and Powell 1998).

317 Valasco Valley pseudosection is represented in Fig. 12a where the observed assemblage garnet, amphibole 1,
318 omphacite, rutile and quartz, is stable in a field that spans from 625 to 740 °C and 1.2 to 1.75 GPa. This penta-
319 variant field is limited by the appearance of plagioclase towards high temperatures, the disappearance of
320 amphibole towards higher pressure and temperatures, the appearance of a jadeitic pyroxene at higher pressures
321 and lower temperatures, the presence of clinozoisite at lower temperatures and the presence of titanite towards
322 lower temperatures. The compositions of clinopyroxene and amphibole in the field are close to those detected in
323 Valasco Valley eclogites. In addition thermobarometric estimates performed on eclogite facies minerals from
324 Valasco Valley rocks perfectly fit with the HP assemblage stability field. Addition of Fe₂O₃ would shift epidote
325 stability towards higher temperature, up to close the penta-variant field we are interested in for around 10 mol%
326 of FeO₁ as Fe₂O₃. Conversely, plagioclase-in and amphibole-out lines would be affected only by minor shifting
327 towards higher temperatures and pressures, respectively, of the order of four tens of Celsius degrees and less
328 than 0.1 GPa. A similar configuration would result from taking into account zoisite instead of clinozoisite in a
329 Fe₂O₃-free system.

330 In the Lac Long pseudosection (Fig. 12b) the assemblage amphibole, omphacite, garnet, rutile, quartz and biotite
331 is stable in a P-T range of 600 to 780°C and 1.25 to 1.80 GPa, in agreement with thermobarometric estimates.
332 The proposed field is bounded by plagioclase at high temperature, amphibole at high pressure, phengite at low
333 temperature, and titanite and clinozoisite at low pressure. If we consider zoisite instead of clinozoisite, the lower
334 temperatures field boundary would be shifted to T at around 690°C, still in agreement with other P-T
335 determinations made for this assemblage. Addition of Fe₂O₃ would shift both epidote stability towards higher
336 temperature, and phengite towards higher temperature and lower pressure, until the closing of the penta-variant

337 field of interest for slightly more than 10 mol% of FeO_i as Fe₂O₃. The effects on the other field boundaries are
338 negligible.

339

340 **Geochronological constraints on eclogite facies metamorphism**

341

342 Ages of eclogite facies metamorphism previously obtained in the eclogites from the ECMs, as well as from other
343 segments of the European Variscan belt, are not well constrained (Von Raumer et al. 2009; Paquette et al. 2017,
344 Lotout et al. 2018). The geochemical signatures of the studied eclogites are totally consistent, particularly
345 regarding multi-elements abundance patterns, with those obtained for HP mafic granulites sampled in the
346 Argentera-Mercantour Massif (Rubatto et al. 2001, 2010) suggesting that these HP mafic granulites could be
347 considered as markers of the incipient stage of eclogites retrogression. In such a case the age of eclogite facies
348 metamorphism is probably close to that obtained for HP granulite facies, i.e. 340 ± 4 Ma (Rubatto et al. 2010) by
349 U/Pb zircon dating. So far, we have not enough zircons from eclogites to try to get an age of the eclogite facies
350 metamorphism. However, we performed ⁴⁰Ar/³⁹Ar dating on amphiboles from these eclogites. We selected a
351 sample from the Lac Long eclogites. This sample (LL09-12J) is closely associated with the (LL13) sample used
352 for the previously presented pseudosection and has a similar geochemistry. In this (LL09-12J) sample,
353 magnesio-hornblendes are in textural equilibrium with plagioclase, ilmenite and garnet. The inferred age
354 corresponds thus to the re-equilibration of eclogites under amphibolite facies conditions.

355 Analytical conditions and results are available as “Online Resource 9”. Amphibole single grain of Lac Long
356 sample (LL09-12J) yields a plateau age (Fig. 13) at 339.7 ± 12 Ma corresponding to 93.77 % Ar released and to
357 four steps. The inverse isochron for the plateau steps gives a concordant age at 339.8 ± 12 Ma (MSWD = 2.17).
358 The plateau-age at 339.7 ± 12 Ma is considered as the best age estimate for the record of amphibolite facies
359 imprint.

360

361 **Discussion and conclusions**

362

363 **Origin of eclogites protoliths**

364

365 Trace elements and REE patterns establish that the Argentera-Mercantour eclogites are MORBs contaminated by
366 continental crust components. Crustal contamination can be the consequence of contamination of basaltic

367 liquids, i.e. eclogite protoliths, prior to HP metamorphism or contamination of eclogites by surrounding
368 migmatites during or after their emplacement.

369 In the first hypothesis, the multi-element diagrams of incompatible elements display Pb, U and Large Ion
370 Lithophile Element (LILE) enrichments, and also Nb and Ta depletions. These patterns are characteristic of arc
371 lavas and back-arc basalts (Hofmann 2003; Labanieh et al. 2012). In that case, the geochemical signatures of the
372 studied eclogites are more compatible with those of supra-subduction zone basalts rather than those oceanic
373 domains.

374 Considering the second hypothesis, during high temperature hydrothermalism as well as during high-temperature
375 crustal melting, significant K, Pb, U and more generally LILEs influxes by crust-derived fluids migration can be
376 envisaged (Michard and Albarède 1986; Pattison 1991; Klinkhammer et al. 1994; Štípská et al. 2014). In such a
377 case, the eclogite protoliths could be standard MORBs later contaminated during orogenic processes.

378

379 **Paleo-gradient and context for eclogites formation**

380

381 The eclogite-facies conditions were quantified by independent methods as comparison between natural and
382 experimentally determined mineral assemblages, well calibrated independent thermobarometers and
383 thermodynamic modeling. The obtained results are consistent with temperatures comprised between 640 and
384 740°C for pressure of 1.5 ± 0.25 GPa.

385 To infer their geodynamic significance, we compare these P-T estimates with metamorphic gradients (Miyashiro
386 1961; Ernst 1971; Spear 1993) and with geotherms proposed by Cloos (1993) for cold and warm subductions
387 (Fig. 14).

388 The eclogitic stage fits the conditions predicted by the the different models calculated for warm subduction of
389 oceanic crust and therefore they represent the traces of a suture zone.

390

391 **The Argentera-Mercantour Massif: record of a dismembered suture zone**

392

393 The Argentera-Mercantour eclogites are witnesses of an oceanic crust formed in a back-arc basin or in an
394 oceanic domain. Whatever is the selected pre-orogenic configuration, these rocks are the exhumed remnants of
395 subducted oceanic crust, therefore localizing a paleo-suture zone transposed after continental collision in a
396 migmatitic complex. However this paleo-suture zone is totally dismembered and eclogites are mixed with

397 various types of migmatites. The original suture is totally overprinted by mass transfer of partially molten
398 continental crust, giving rise to a “diffuse cryptic suture zone” in the sense of Schulmann et al. (2014).

399 The mechanism leading the extensive dismembering of the suture zone remains unclear and different models can
400 be considered. We can envisage dismembering, during continental collision, of previously stacked ophiolitic unit
401 in response to vertical transfer (i.e. extrusion) of hot partially molten orogenic crust during ongoing shortening.
402 This results in large-scale gravity driven migmatitic dome-like structures favouring mixing of eclogites and
403 migmatites as proposed in Variscan Bohemian Massif (Štípská et al. 2004; Schulmann et al. 2005, 2009, 2014)
404 or in the Maures-Tanneron Massif (Schneider et al. 2014, Gerbault et al. 2018). We can also consider
405 dismembering of subducted crust during oceanic subduction and/or incipient collision leading the formation of a
406 “hot subduction channel” composed of subducted, partially molten, upper crustal rocks and lithospheric hydrated
407 mantle rocks as proposed by et al. (2008). The result is a “ hot subduction mélange”, localized at the plates
408 interface, producing fast exhumation of mafic HP rocks within molten meta-sediments.

409

410 **The Argentera-Mercantour eclogites in the southern European Variscides framework**

411

412 Considering field relationships between eclogites and surrounding migmatites, their geochemistry, and
413 petrological evolutions, the Argentera-Mercantour eclogites show remarkable similarities with Sardinian
414 eclogites (Giacomini et al. 2005; Franceschelli et al. 2007; Cruciani et al. 2015, 2017; Fig. 14), with eclogites
415 from the Maures-Tanneron Massif (Bard and Caruba 1982; Briand et al. 2002; Corsini and Rolland 2009;
416 Schneider et al. 2014) and finally with Corsican eclogites (Palagi et al. 1985; Lardeaux et al. 1994; Faure et al.
417 2014).

418 The European Variscan belt is the result of tectonic accretion of Gondwana derived micro-plates colliding with
419 Laurussia during the Devonian (Torsvik et al. 1990, 1996; Tait et al. 1997; Matte 2001; Von Raumer et al. 2002;
420 Cocks and Torsvik 2011; Stampfli et al. 2013). The continuity of this process led to the formation of a huge
421 collision belt and to the amalgamation of the Pangea supercontinent in the late Paleozoic (Franke 1989, 2000;
422 Matte et al. 1990; McKerrow et al. 2000; Domeir and Torsvik 2014). Among the various micro-plates involved
423 in this evolution, paleomagnetic constraints combined with geochronological data and regional geologic
424 investigations have identified, in the southern Variscides, the so-called Maures-Esterel-Corsica-Sardinia
425 (“MECS”) block (see Edel et al. 2014, 2018). If we consider that the occurrences and the petrochemical
426 characteristics of eclogites are critical for deciphering the main suture zones in orogenic systems, the similarities

427 between HP metamorphic rocks recognized in Argentera-Mercantour, Sardinia, Maures-Tanneron and Corsica
428 suggest that the southernmost ECMs possibly linked with the MECS block. Such a link is also consistent with
429 various paleo-plates tectonic reconstructions proposed by Stampfli et al. (2013). In the framework of Gondwana
430 derived micro-plates these authors distinguish, during upper Devonian times, northern blocks (Iberian Massif,
431 Armorican Massif, French Massif Central, Bohemian Massif), located close to the Rheic (or Rheno-Hercynian)
432 ocean and southern blocks (Sardinia, Corsica, ECMs,...) located close to the Paleo-Tethys. In such a scheme, the
433 ages of eclogite facies metamorphism are different with a decrease of geological ages of HP metamorphism from
434 northern to southern blocks. Geochronological data are therefore first-order constraints to discuss these models.
435 In the MECS block, late Cambrian-Ordovician to lower Silurian protolith ages of HP rocks are constrained by
436 U/Pb zircon dating (Innocent et al., 2003; Palmeri et al. 2004; Giacomini et al. 2005, 2008; Rossi et al. 2009;
437 Schneider et al. 2014; Cruciani et al. 2017). This time span is coherent with ages of 459 ± 4 Ma and 486 ± 7 Ma
438 obtained in the Argentera-Mercantour Massif (Rubatto et al. 2001, 2010) but also with as all protoliths ages
439 obtained in the ECMs (Von Raumer et al. 2002; Schultz et al. 2004). In the MECS block, the timing of crustal
440 anatexis is constrained in the range 345-310 Ma (Di Vincenzo et al. 2004; Giacomini et al. 2005, 2008; Demoux
441 et al. 2008; Rossi et al. 2009; Corsini et al. 2010; Faure et al. 2014; Li et al. 2014; Cruciani et al. 2015; Oliot et
442 al. 2015). The age of 323 ± 12 Ma obtained for crustal partial melting in the Argentera-Mercantour Massif
443 (Rubatto et al. 2001) as well as our age at 339.7 ± 12 Ma for eclogite re-equilibration under amphibolite facies
444 conditions are in agreement with these results. A characteristic signature of continental collision in the MECS
445 block is the development of a transpressive regime in the range 330-310 Ma leading the development of ductile
446 strike-slip shear zones in Sardinia (Elter et al. 1990; Carosi and Oggiano 2002; Carosi et al. 2004, 2009) and
447 Maures-Tanneron Massif (Corsini and Rolland 2009; Corsini et al. 2010). In the Argentera-Mercantour Massif,
448 the FMSZ transpressive shear zone active in the range 330-310 Ma (Simonetti et al. 2018) is a record of a similar
449 tectonic regime.

450 However, the ages of eclogite facies metamorphism remains poorly constrained in the MECS block as well as in
451 the ECMs (Von Raumer et al. 2009; Schneider et al. 2014; Cruciani et al. 2015; Spalla et al. 2014; Regorda et al.
452 2019) and new radiometric datings are necessary to discriminate between the different geodynamic scenarios.
453 The U/Pb zircon age obtained on HP granulites in Corsica at ca. 360 Ma is still debated (Li et al. 2014), therefore
454 the only available geochronological constraints are the age of HP mafic granulites at ca. 340 ± 4 Ma (Rubatto et
455 al. 2010) and the the time span of 344-330 Ma. for retrogression of eclogites under amphibolite facies.

456 In the framework of the Alpine belt, the identification of which and how many Variscan suture zones (e.g. Matte
457 1986, 2001; Franke 2000; Faure et al. 2005; Guillot et al. 2009; Lardeaux et al. 2014) occur in the Alpine chain
458 is still debated (Regorda et al. 2019). The evolution of this Palaeozoic lithosphere in terms of geodynamic
459 scenario has been rarely explored (e.g. von Raumer et al. 2003, 2013; Spiess et al. 2010). The discussion on the
460 provenance of the Alpine basement units in terms of palaeogeography and influence of Palaeo-Tethys evolution
461 on the Variscan dynamics is still open. For this purpose, the comparison of the thermo-mechanical evolution of
462 the Argentera-Mercantour eclogites with those of other Variscan eclogites from the Alps is a key starting point
463 to shed light on their geodynamic significance (Fig. 14 and “Online Resource 10”). PT conditions inferred from
464 eclogites and HP-granulites of Helvetic Domain indicate a thermal state higher than that of warm subduction
465 zones (Cloos 1993), mainly compatible with the initial unperturbed geotherm or higher, with the only exception
466 of Valasco Valley and Lac Niré rocks. The P-T conditions of the latter lie on the warm subduction geotherm,
467 fully comprised in the Franciscan metamorphic gradient, like some of the Sardinian eclogites. For the first time
468 this indicate that oceanic subduction-related metamorphism developed also in the Helvetic Variscan crust. This
469 makes the investigated rocks more similar to the Variscan eclogites from other Alpine structural domains, which
470 mainly show a thermal state between warm and cold subduction gradients or locally colder, as the case of some
471 Austroalpine eclogites from Oetztal, Tonale and Silvretta nappes.

472 The previously presented constraints demonstrate the coherence of the Variscan geological signatures depicted
473 in the Argentera-Mercantour Massif and those of the MECS block, which share the same tectono-metamorphic
474 evolution. Hence we envisage that the southernmost ECM is a part of the MECS block in the framework of the
475 European Variscan belt. To test such a hypothesis it will be now critical to obtain high-resolution chronological
476 constraints on the age of eclogite facies metamorphism, particularly in Argentera-Mercantour and Maures-
477 Tanneron Massifs.

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916 **Figure captions**

917

918 Fig. 1: The Argentera-Mecantour Massif in the Alpine arc (after Spalla et al., 2014; Regorda et al., 2019). Black
919 stars: Eclogites known in the pre-Alpine basements from different alpine zones; star labels individuate Alpine
920 and Sardinian eclogites as listed in “Online Resource 10”. Inset shows the Alpine arc in the framework of the
921 European Variscan belt. Inset labels: Arm - Armorican Massif; BCBF - Bristol Channel-Bray Fault; BM -
922 Bohemian Massif; Ca - Cantabrian terrane; Cib - Central Iberian; Co - Corsica; FMC - French Massif Central;
923 GSBS - Galicia-Southern Brittany Suture; MT - Maures-Tanneron Massif; OM - Ossa Morena; Py - Pyrenees;
924 Sa - Sardinia; Si - Sicilian-Apulian basements; SP - South Portuguese Zone; WL - West Asturian-Leonese.

925

926 Fig. 2: Geological map of the Argentera-Mercantour Massif (after Faure-Muret, 1955, Malaroda et al. 1970,
927 Corsini et al., 2004, Carosi et al. 2016 and Gosso et al. 2019). FMSZ - Ferrere-Mollières Shear Zone. CG –
928 Central Granite. 1 Eclogite occurrences and sampling localities with yellow stars; 2 Migmatitic paragneisses; 3
929 Migmatitic orthogneisses and metavolcanics; 4 Amphibolites; 5 Syn-tectonic diorites; 6 leucocratic cordierite-
930 bearing anatexites; 7 Amphibolite to greenschistfacies mylonites; 8 Granitoids; 9 Carbonaceous schists,
931 conglomerates, and siliciclastic sandstones (late Carboniferous); 10 -Permian to Eocene sequences; 11 late- to
932 post-Alpine faults; 12 Penninic Front; 13 Greenschist-facies mylonites. White contours labelled a and b locate
933 the maps of Fig. 5 and 6, respectively.

934

935 Fig. 3: Eclogitic boudins. (a) Col de Fenestre– Pas des Ladres (France) eclogitic boudin in metatexites; (b)
936 Balma Ghiliè valley (Italy) eclogite within paragneisses; (c) Monte Matto (Italy) boudin of eclogite in cordierite
937 bearing diatexites; (d) Lac Niré (France) eclogite within paragneisses; (e) Lac Long eclogitic boudin in
938 metatexites.

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940 Fig. 4: Retrogressed eclogites. (a) Lago Brocan (Italy) coronas of amphibole around garnets; (b) Lac Niré
941 (France) eclogite with plagioclase around garnets.

942

943 Fig. 5: Geological map of the high Gordolasque valley at the Italian-French border. (a) Structural map.
944 (b) Detailed map of eclogite boudin outcropping north of the Lac Long and cross-section.

945

946 Fig. 6: Structural map of upper Valasco Valley (Italy). In the inset detailed form-surface map located with a
947 black star in the structural map, close to Lago Inferiore di Valscura.

948

949 Fig. 7: Geochemistry of Argentera-Mercantour eclogites in the Al₂O₃ vs. TiO₂ diagram after Konzett et al. 2012
950 (7a) and in the AFM diagram after Irvine and Baragar, 1971 (7b).

951

952 Fig. 8: Chemical compositions of eclogites in the Y-Nb-Zr (8a after Meschede, 1986), the Th/Yb versus Ta/Yb
953 (8b after Pierce, 1982) and in the N-MORB normalized Th-Nb (8c after Saccani, 2015) diagrams. Fields hatched
954 in grey represent available chemical compositions of Sardinian eclogites (see text for discussion and references).

955

956 Fig. 9: Primitive mantle normalized REE (a) and LILEs HFSEs (b) patterns of eclogites. N-MORB, E-MORB
957 and OIB compositions after Sun and McDonough (1989).

958

959 Fig. 10: Photomicrographs of eclogites. (a) Clinopyroxene-plagioclase symplectite (crossed polars, lac Niré,
960 France). Progressive decrease of lamellae size from center to rim; (b) Evolution of the symplectite, from coarse
961 clinopyroxene lamellae (with Jadeite content of 42 mol%) to thinner lamellae (with jadeite content of 15
962 mol%), crossed polars, Pas des Ladres (France). Gt - Garnet; Pl - Plagioclase; Amp - Amphibole. (c) Re-
963 equilibrated eclogites with diopside-plagioclase symplectites. Relict amphibole 1 green cores are rimmed by
964 brown Ti rich amphibole 2 that together with plagioclase occurs in coronas around garnets (plane polarized light,
965 Lago Brocan, Italy); (d) Detail on the plagioclase-amphibole 2 coronas around garnet; (e) Inclusions of quartz,
966 amphibole 1 and rutile in garnet (BSE image, Brocan Lake, Italy); (f) Diopside-plagioclase symplectite replacing
967 omphacite (BSE image, sample 0878, Valasco Valley, Italy).

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Fig. 11: Composition of clinopyroxenes (a), garnets (b), feldspaths (c) and amphiboles (d) from Argentera eclogites. In all the diagrams minerals from Lac Niré, Lac Long and Pas de Ladres eclogites are represented by stars. Italian eclogites are circles. Different colors have been used for minerals from Valasco Valley: in the diagram 11(a) of Morimoto (1988) black and grey circles represent the recalculated compositions of omphacites and clinopyroxene 2 respectively. In comparison, omphacite of Lac Niré appears in black stars. In Na/Na+Ca vs Al/Al+Si diagram 11(d), stars locate the end-member compositions labelled as follows: ACT - Actinolite, BAR - Barroisite, ED - Edenite, GL - Glaucophane; KTP - Katophorite, PRG - Pargasite, SAD - Sadanagaite, TS - Tschermakite, WIN - Winchite.

Fig. 12: P-T pseudosections calculated from Valasco Valley (a, CARA884) and Lac Long (b, LL13). The red line divides the P-T field of plagioclase bearing assemblages, towards higher temperatures, from assemblages without plagioclase, towards higher pressures. The shaded light blue fields represent the P-T conditions determined using classic geothermobarometry for the eclogite-facies assemblage. Univariant lines are thick. For pseudosection (a), assemblages in the numbered fields are as follow: 1 - garnet, 2 clinopyroxenes, rutile, quartz; 2 - garnet, amphibole, 2 clinopyroxenes, rutile, quartz; 3 - garnet, amphibole, 2 clinopyroxenes, rutile, clinozoisite, quartz; 4 - garnet, 2 amphiboles, 2 clinopyroxenes, rutile, quartz; 5 - garnet, 2 amphiboles, clinopyroxene, clinozoisite, rutile, quartz; 6 - garnet, amphibole, clinopyroxene, titanite, clinozoisite, rutile, quartz; 7 - garnet, amphibole, clinopyroxene, titanite, clinozoisite, quartz; 8 - garnet, amphibole, titanite, clinozoisite, quartz; 9 - garnet, amphibole, clinopyroxene, titanite, quartz; 10 - garnet, amphibole, clinopyroxene, titanite, rutile, quartz. For pseudosection (b), numbered fields match with the following assemblages: 1 - garnet, white mica, 2 amphiboles, clinopyroxene, quartz, rutile; 2 - garnet, white mica, 2 amphiboles, clinopyroxene, quartz, rutile, clinozoisite; 3 - garnet, biotite, white mica, amphibole, 2 clinopyroxenes, quartz, rutile; 4 - garnet, biotite, amphibole, 2 clinopyroxenes, quartz, rutile; 5 - garnet, biotite, white mica, amphibole, clinopyroxene, quartz, rutile; 6 - garnet, biotite, white mica, amphibole, clinopyroxene, quartz, rutile, clinozoisite; 7 - garnet, biotite, amphibole, clinopyroxene, quartz, rutile, titanite; 8 - garnet, biotite, amphibole, quartz, rutile, sphene; 9 - garnet, biotite, amphibole, clinopyroxene, quartz, rutile, clinozoisite; 10 - plagioclase, garnet, biotite, amphibole, quartz, rutile; 11 - plagioclase, garnet, biotite, amphibole, quartz, rutile, titanite; 12 - garnet, biotite, clinopyroxene, quartz, rutile; 13 - plagioclase, garnet, biotite, clinopyroxene, quartz, rutile. Some very small fields have not been numbered.

Fig. 13: $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum diagram performed in amphibole from Lac Long sample.

Fig. 14: P-T conditions (violet square) of studied eclogites compared with metamorphic gradients (after Miyashiro, 1961; Ernst, 1971; Spear, 1993) and geotherms after Cloos (1993): (1) near spreading ridge or volcanic arc, (2) normal gradient of old plate interior, (3) warm subduction zones, and (4) cold subduction zones. Vo represents the initial unperturbed geotherm. (a) P-T conditions of Argentera eclogites compared with eclogites from Helvetic-Dauphinois zone and Sardinian eclogites. (b) P-T conditions of Argentera eclogites compared with eclogites from Penninic and Austro-Alpines zones. Box labels refer to locations in Figure 1 and descriptions in "Online Resource 10".

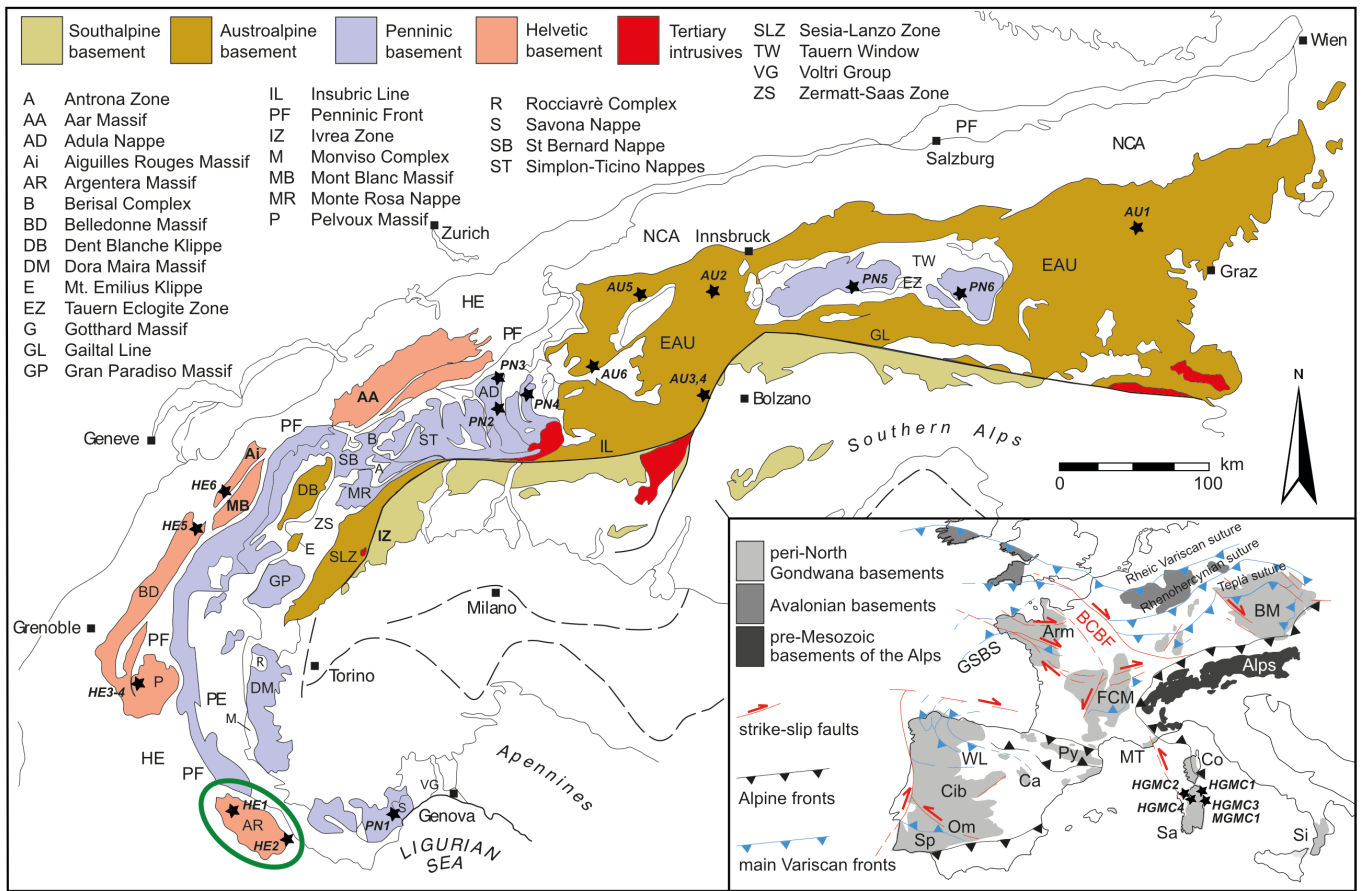


FIGURE 1

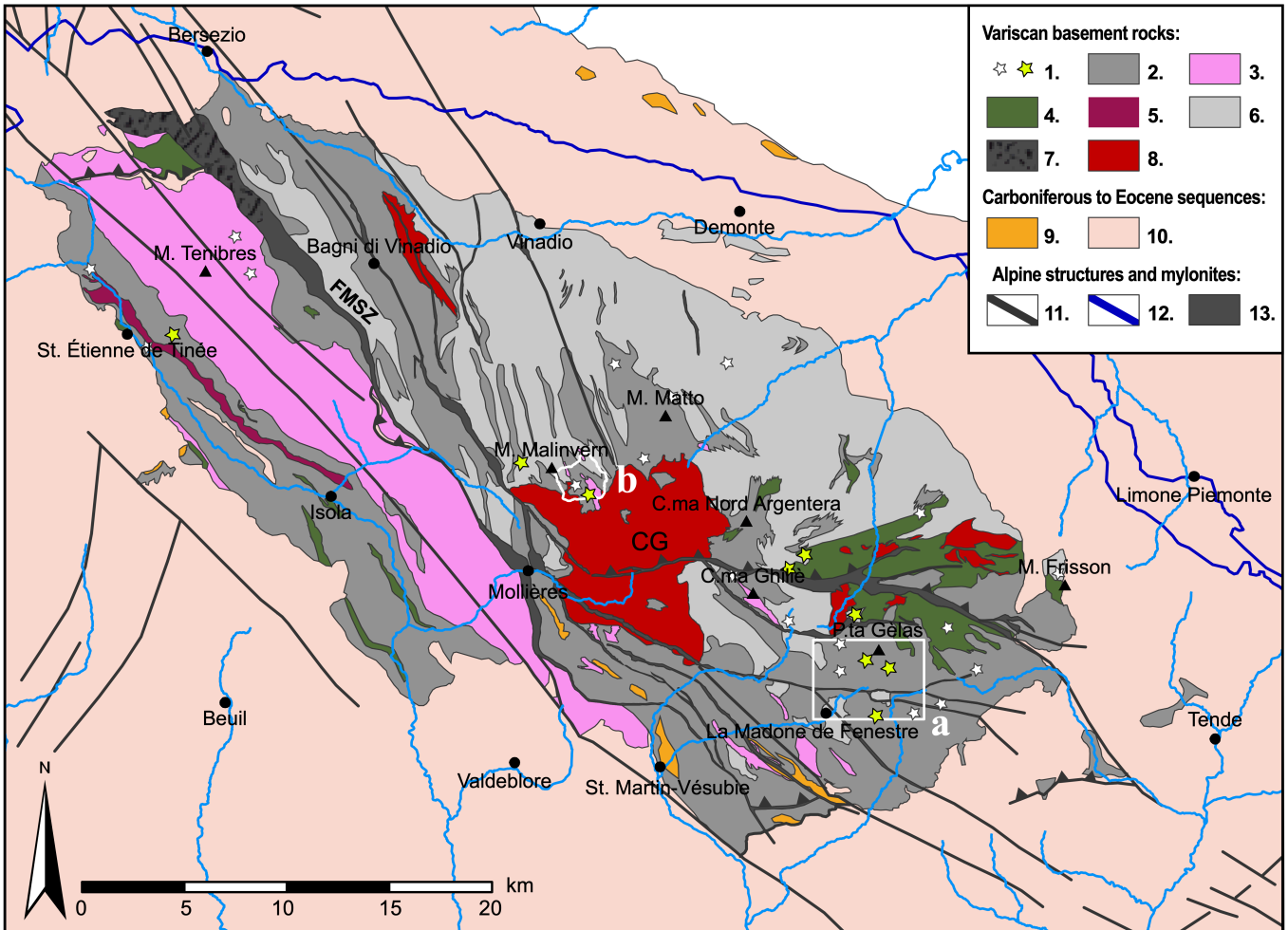


FIGURE 2

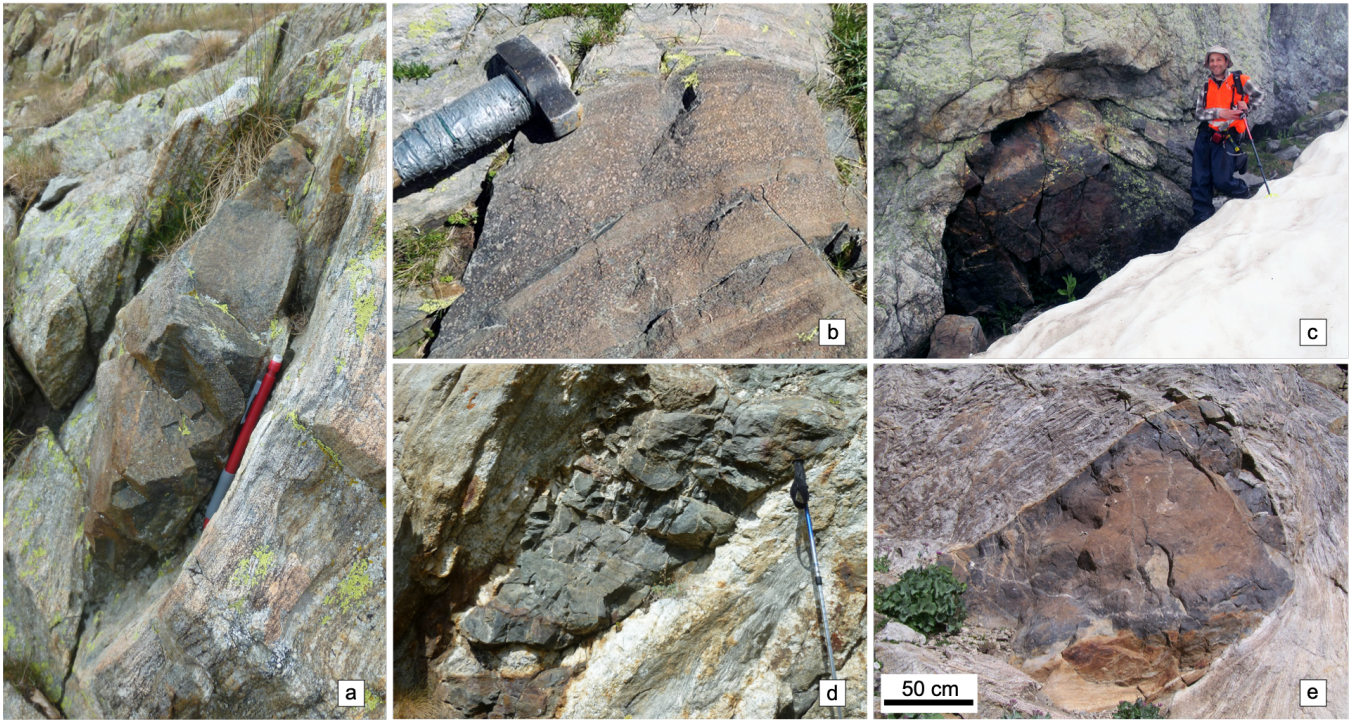


FIGURE 3

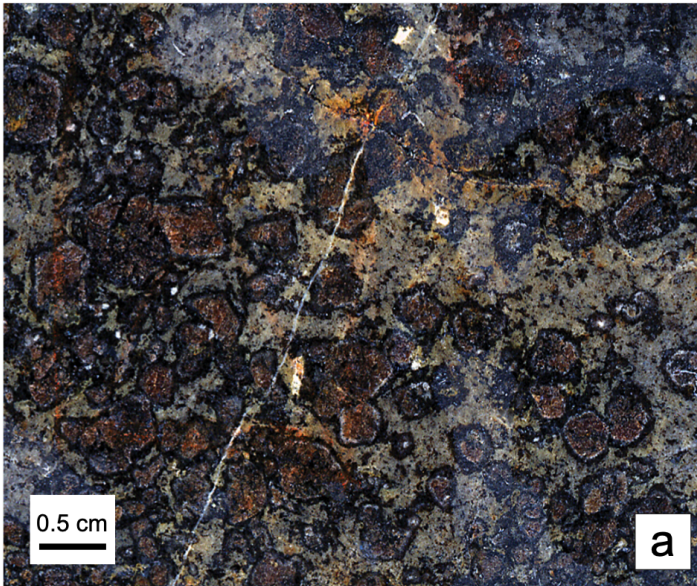
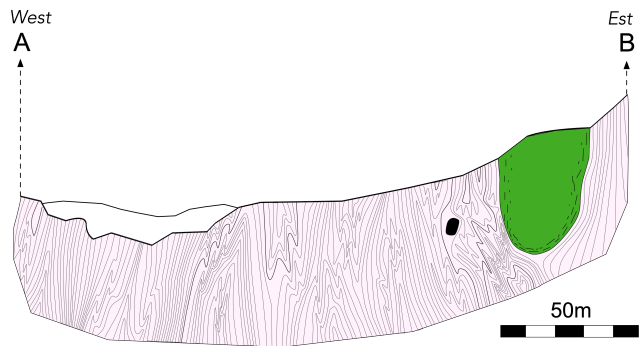
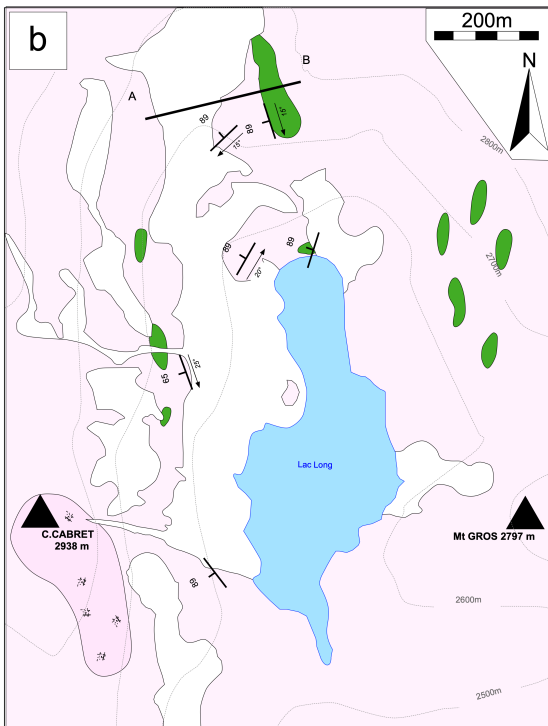
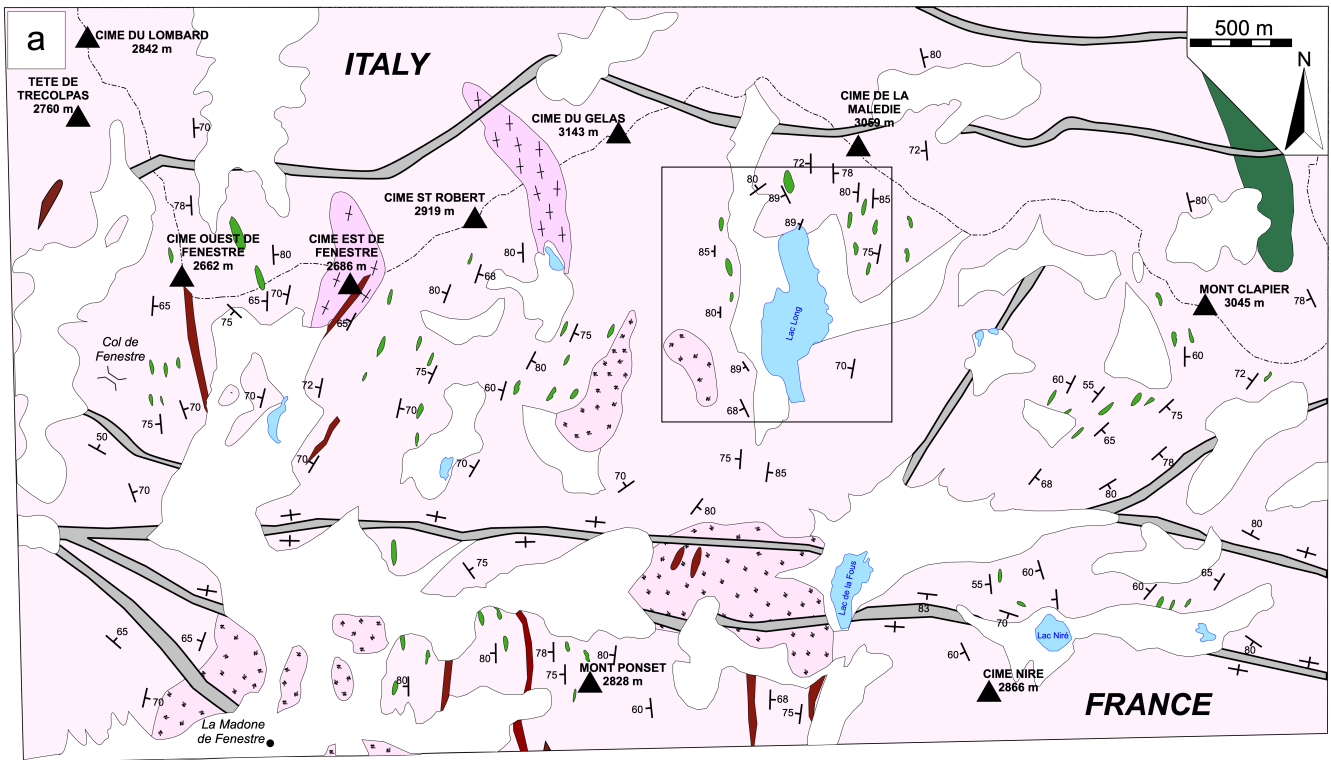


FIGURE 4






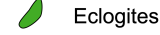




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|---|---------------------------------------|---|------------------|
|  | Late surficial covers |  | Amphibolites |
|  | Migmatitic paragneisses (metatexites) |  | Eclogites |
|  | Cordierite-bearing diatexites |  | Granitic dykes |
|  | Migmatitic orthogneisses |  | Alpine Mylonites |

FIGURE 5

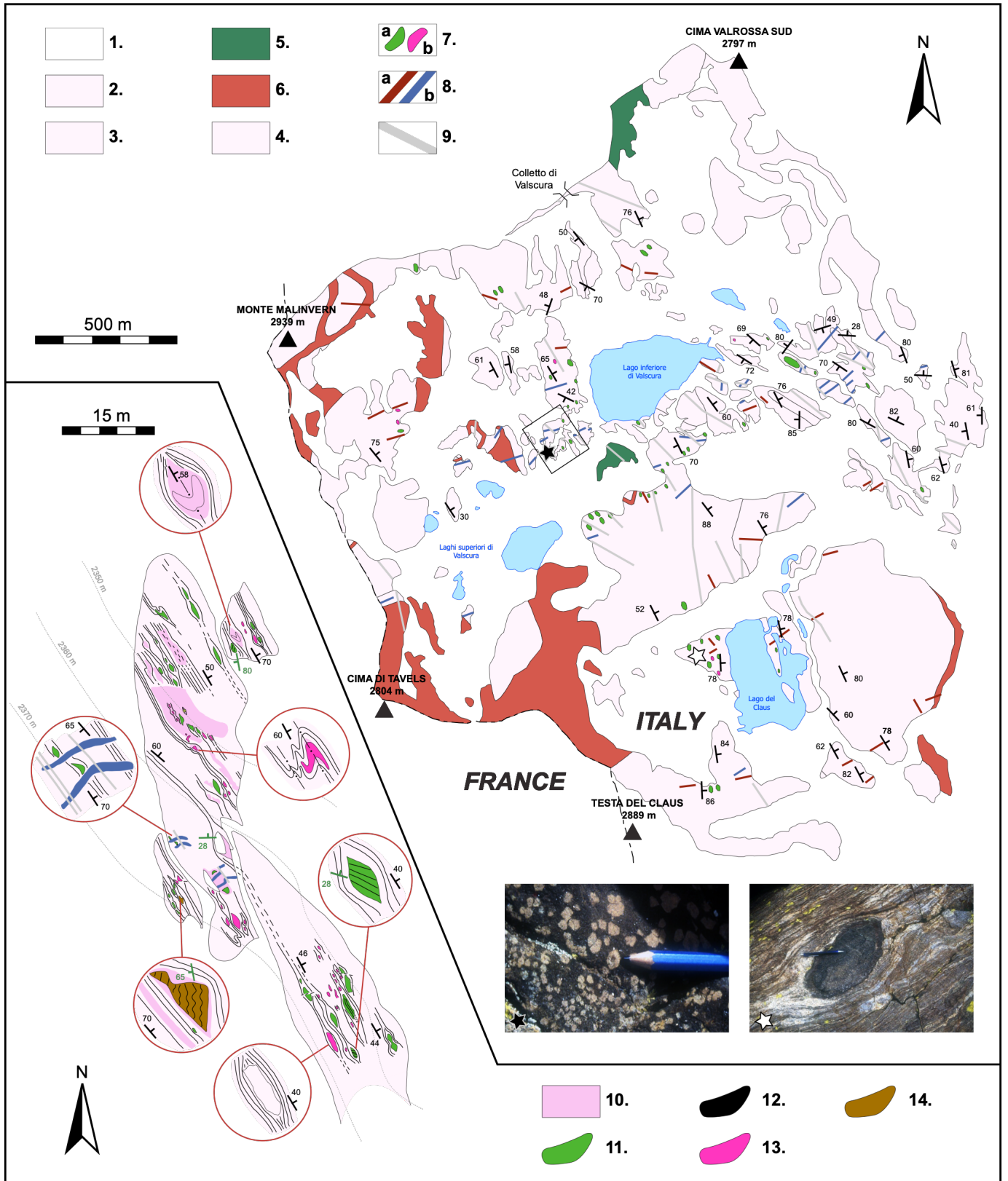


FIGURE 6

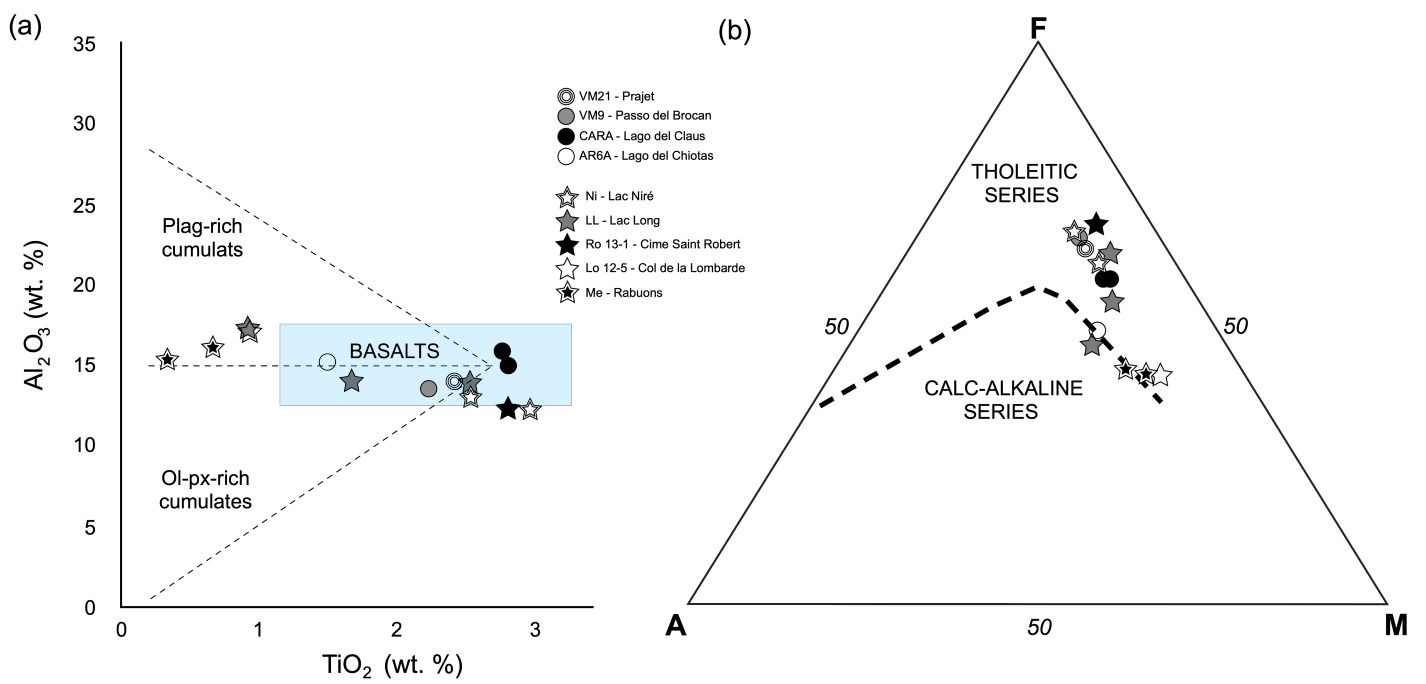


FIGURE 7

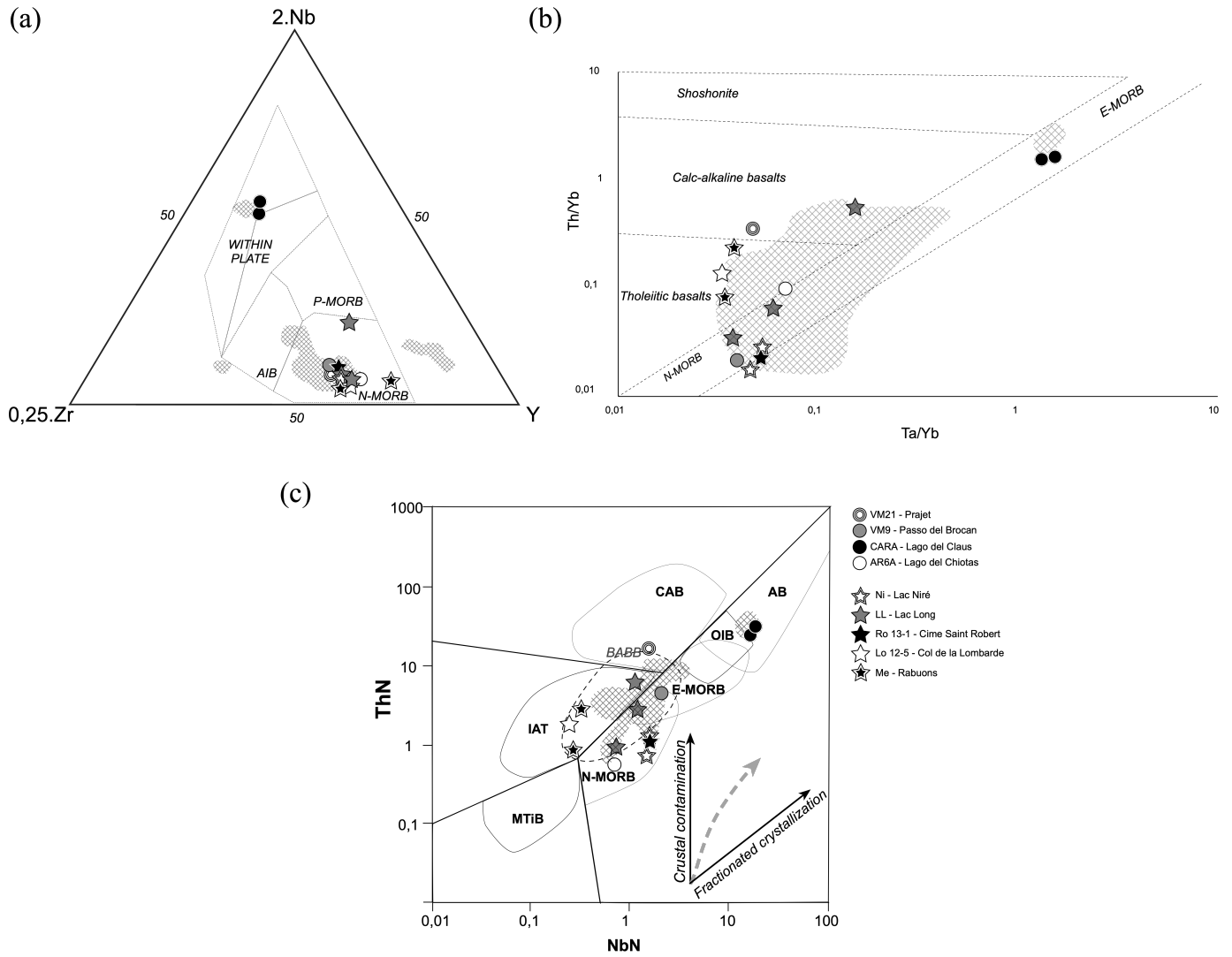


FIGURE 8

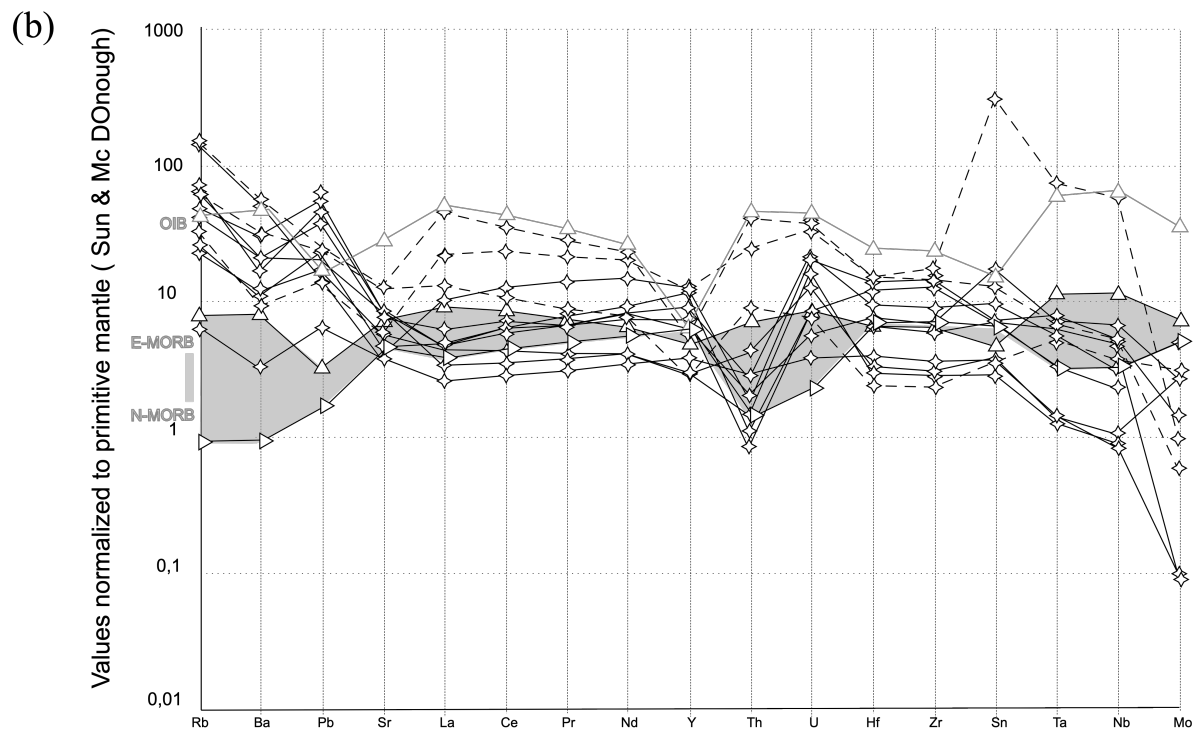
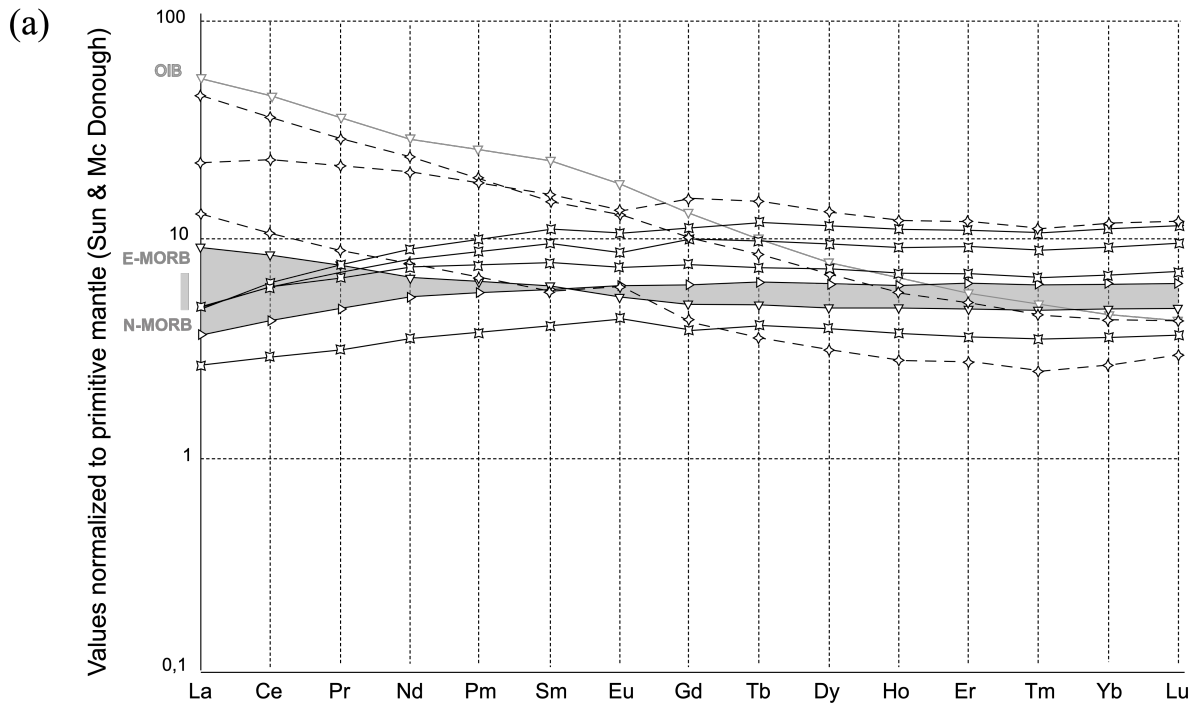


FIGURE 9

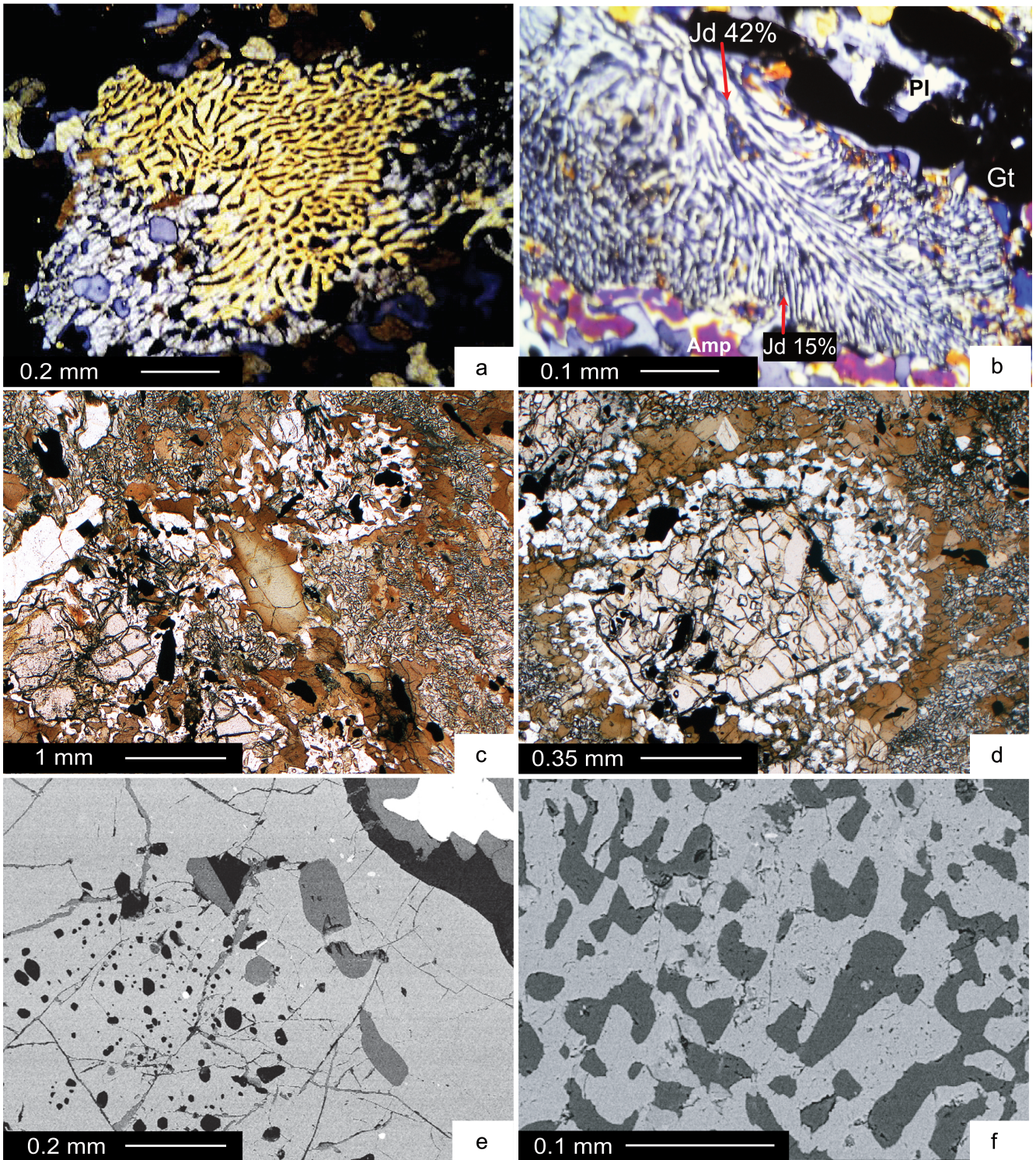


FIGURE 10

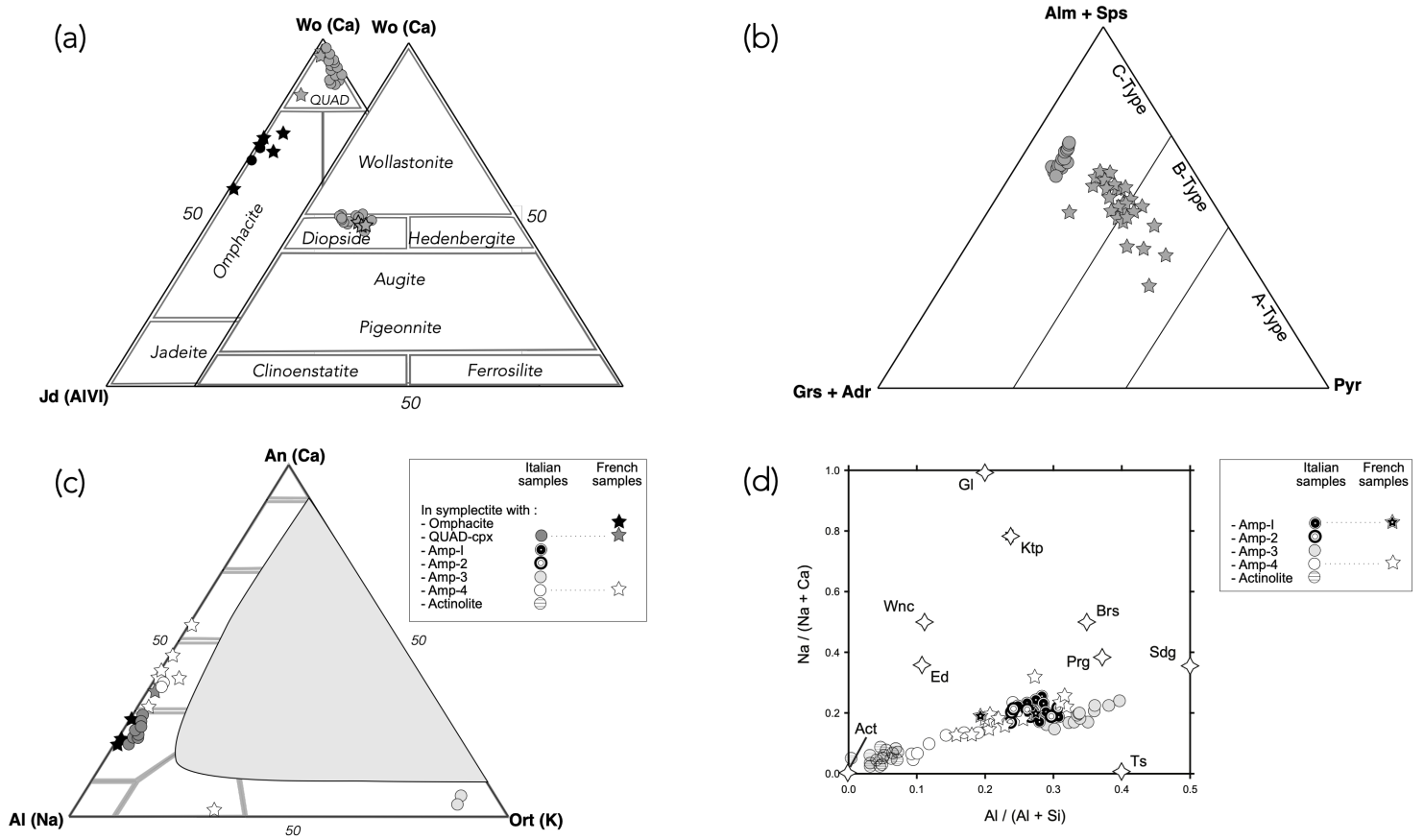


FIGURE 11

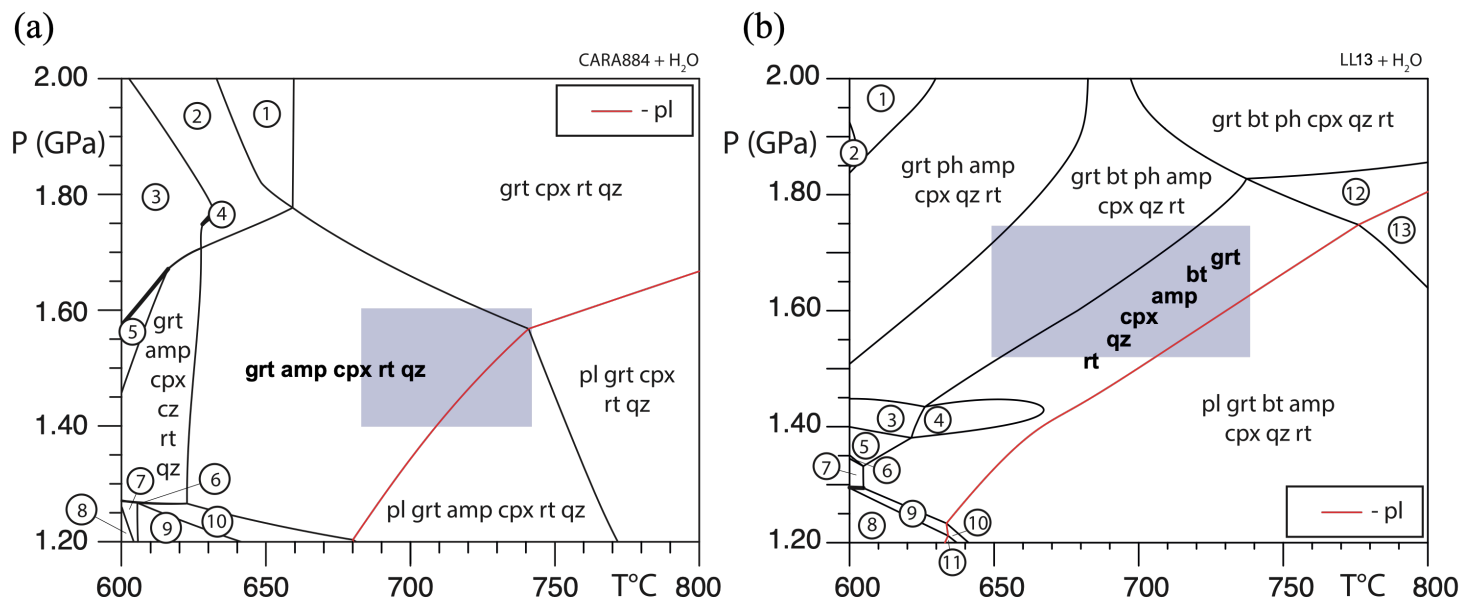


FIGURE 12

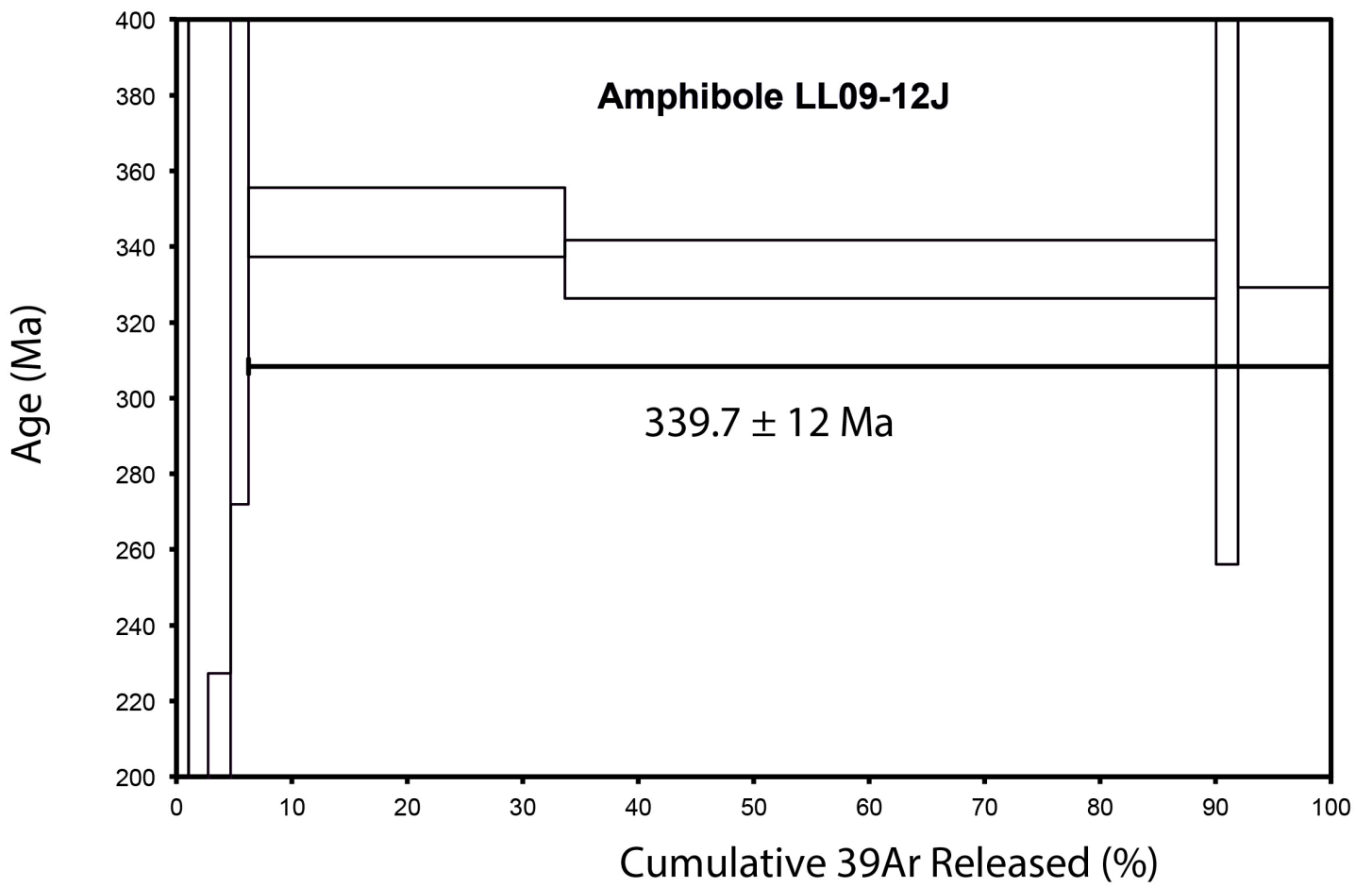


FIGURE 13

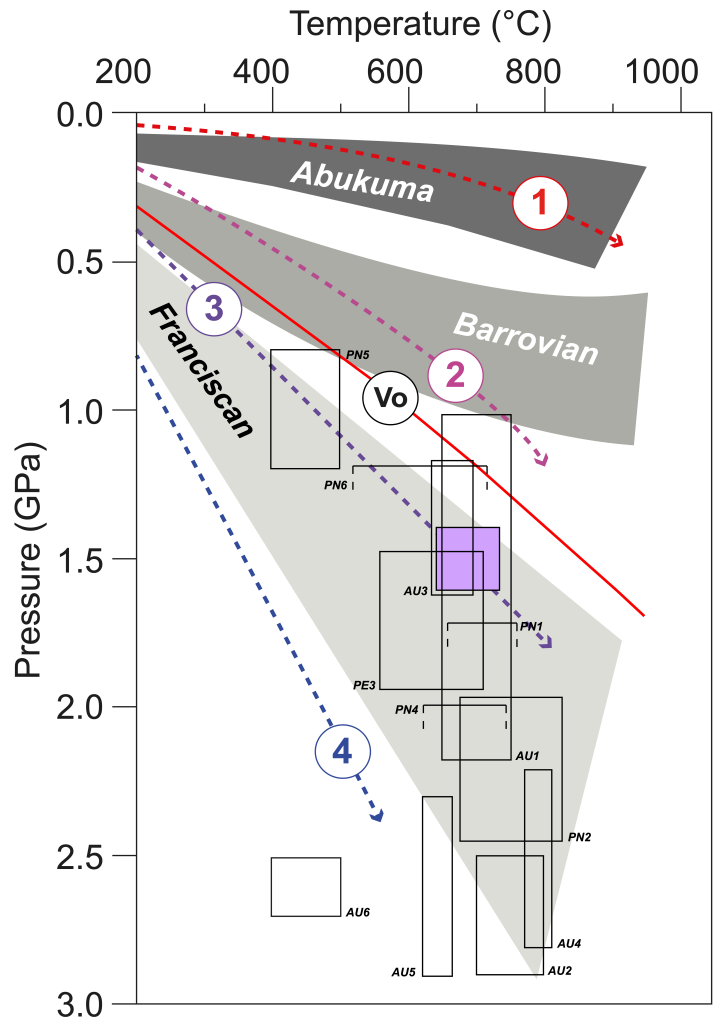
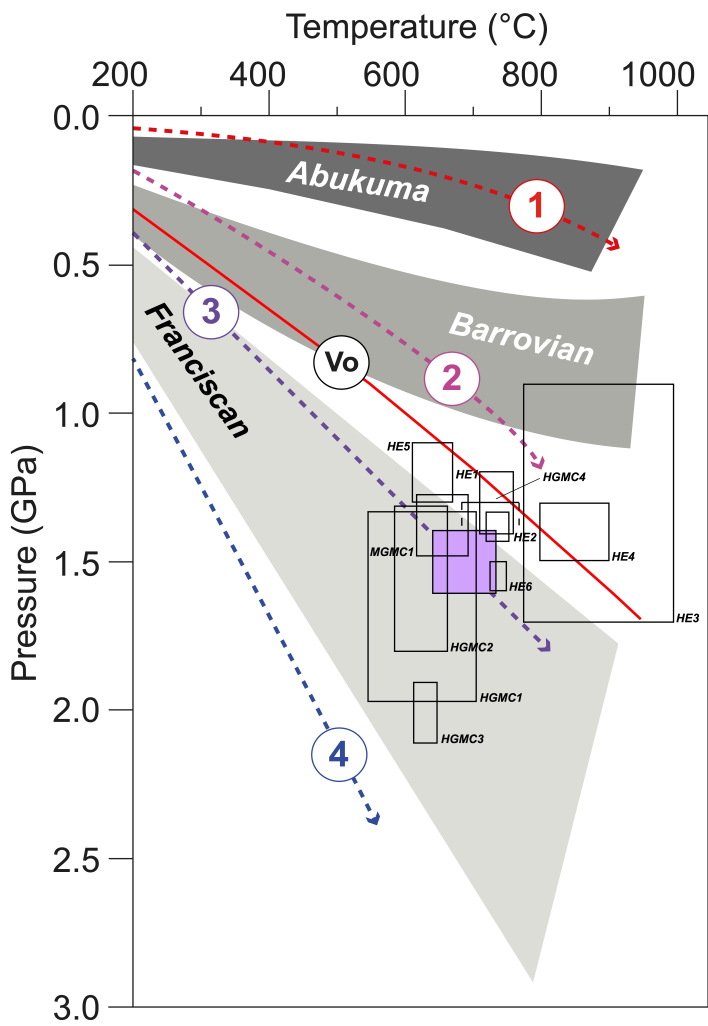


FIGURE 14