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1 **Tracing metallic pre-concentrations in the Limousin ophiolite-derived rocks and Variscan**
2 **granites (French Massif Central)**

3

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20 **Abstract**

21 The French Massif Central (FMC) is part of the West European Variscan belt. The economic and
22 sub-economic ore deposits from the Limousin area are the consequence of intensive fluid-rock
23 interactions related to a lithospheric delamination process that occurred during the Neo-Variscan
24 Period (320–290 Ma). This study focuses on the fingerprints of fluid-rock interactions in the
25 Limousin ophiolite-derived rocks and granitic massifs, in order to trace the source of trace
26 elements transported by fluids involved in the formation of the Variscan ore deposits (including

27 Sn, W, Au and U deposits). A series of compatible and incompatible metallic trace elements (Ag,
28 As, Au, Bi, Li, Co, Cr, Cu, Ga, Ge, Ni, Pb, Sb, Sc, Sn, Ti, V, W) were analysed in a series of
29 basic (metagabbros, basic dykes), ultrabasic (serpentinites) and felsic (granites and gneisses)
30 rocks, as well as in the mineral assemblages, in order to determine their relationships with the
31 Limousin ore deposits.

32 Serpentinites and their mineral assemblages (serpentine, olivine, amphibole, chlorite and
33 spinel) show high concentrations in compatible transition metals (Sc, V, Cr, Co, Ni, Zn, Cu) and
34 low abundances of incompatible metals (Ge, Ga, Mo, Sn, Sb, W). Associated amphibolites
35 (metagabbros and metamorphosed dykes) show high contents of compatible metals, though
36 lower than ultrabasic rocks. Amphibole and, to a lesser extent, titanite, are the main hosts for
37 metallic trace elements. Abundances in incompatible metals are generally low. Amphibole has
38 low Sn and Sb contents, while higher Sn contents were measured in titanite. In a UHP zoisite-
39 eclogite cropping out close to one of the ophiolite-derived massifs, zoisite, rutile, omphacite and
40 retrograde amphibole show high concentrations of incompatible elements (Sn, Sb, W, As, Ga,
41 Ge, Mo and Bi). However, it has been shown that high-temperature and low-temperature
42 hydrothermal alteration that have affected the basic and ultrabasic protoliths is c. 100 Ma older
43 than those that were responsible for the formation of ore deposits. For this reason and as well
44 because of their low W and Sn abundances, the ultrabasic and basic rocks cannot be
45 considered as source for the Limousin W–Sn ore deposits.

46 In granite and gneisses, the metallic trace elements are mainly hosted by biotite and
47 muscovite. Quartz and feldspars only show minor amounts of Sc, Ga, Ge, Sn and Pb. The micas
48 contain low amounts of Ge, As, Sb, Bi and variable amounts of Sc, Cu and Ga. Besides, micas
49 display significant Sn and W contents. Variations in Sn and W in micas allow tracing the pre-
50 mineralisation concentrations in the granitic massifs: there is a clear geographical correlation
51 between the Sn and W composition of micas and the importance of the W–Sn ore deposits,
52 arguing for a local source of pre-concentrations. Granites from the Vaulry W–Sn ore deposit

53 display the highest W and Sn concentrations, while granites hosting the Au deposits in the
54 southern part of the Limousin (St-Yrieix, Cheni, Laurières) show the smallest concentrations.

55

56 **Keywords**

57 Trace elements; Metallic elements concentrations; serpentinites; metagabbros; granite; Variscan

58 belt; LA-ICP-MS

59 **1. Introduction**

60 The French Massif Central (FMC) belongs to the internal zone of the West European Variscan
61 belt (e.g. Matte, 2001; Lardeaux et al., 2014) (Fig. 1a). It results from a continent-continent
62 collision between the Gondwana and Laurussia (formed of Laurentia, Baltica and Avalonia) that
63 has started during the Devonian (c. 355 Ma), after the closure of oceanic domains and collision
64 of continental fragments derived from the disruption of the northern Gondwana margin (Eo-
65 Variscan stage) during the Early Palaeozoic (Cambrian-Silurian) (e.g. Ledru et al., 1989; Franke,
66 1992; Faure et al., 2009; Lardeaux et al., 2014; von Raumer et al., 2015; Stephan et al., 2019).
67 Different models, partly diverging, were developed for the evolution of the Variscan orogen and
68 Palaeozoic geodynamics. Major oceanic domains were separating the continental domains. In
69 particular, the Rheic ocean that opened in the Cambrian–Ordovician along the continental
70 margin of northern Gondwana was one of the most important oceanic domains during the
71 Palaeozoic (e.g. Matte, 2001; Pin et al., 2001; Stampfli et al., 2002, 2011; Ballèvre et al., 2009;
72 Faure et al., 2009, 2017; Nance et al., 2010; Romer and Kroner, 2013; von Raumer et al., 2013,
73 2015, 2016; Lardeaux et al., 2014; Franke et al., 2017; Stephan et al., 2019).

74 The Moldanubian Zone (or allochthonous domain; Fig. 1a) includes a series of basic and
75 ultrabasic magmatic rocks, with an oceanic affinity. Geochronological data indicate that these
76 rocks formed during the Cambrian-Ordovician, and are contemporaneous with to the opening of
77 the Rheic ocean (for a review of ages, see von Raumer et al., 2013, 2015). In the Western
78 Variscan belt, these ophiolite-derived rocks were interpreted as the remnants of a narrow ocean
79 between Gondwana and Armorica (Matte, 2001), the "Galicia-South Brittany-Moldanubian" or
80 "Medio-European" ocean (Matte, 2001; Faure et al., 2005; Lardeaux et al., 2014), or the
81 remnants of a late-Cambrian active margin setting along the northern margin of the Gondwana
82 (von Raumer et al., 2015). Other studies suggest that the ophiolite-derived rocks recognised in
83 the Moldanubian zone correspond to the oriental branch of the Rheic ocean (e.g. Stephan et al.,
84 2019).

85 Two main orogenic events are recognised in the Western Variscan belt: 1) Middle
86 Devonian subduction of the oceanic crust under Gondwana and Laurussia; and 2) Early
87 Carboniferous continental collision between Laurussia and Early-Palaeozoic Gondwana-derived
88 continental terranes (e.g. von Raumer et al., 2013; Kroner and Romer, 2013; Lardeaux et al.,
89 2014; Faure et al., 2017; Franke et al., 2017). The FMC has recorded a polycyclic evolution
90 during the Variscan orogeny corresponding to six main tectonic-metamorphic events (e.g. Faure
91 et al., 2009; Lardeaux et al., 2014): 1) Late Silurian high-pressure metamorphism (D_0 ; c. 415
92 Ma); 2) Early Devonian top-to-southwest nappe displacement, migmatisation and exhumation of
93 HP rocks (D_1 ; 385–380 Ma); 3) Late Devonian to Early Carboniferous top-to-northwest shearing
94 and medium pressure–medium temperature metamorphism (D_2 ; 360–350 Ma); 4) Visean top-to-
95 south ductile shearing and onset of synorogenic extension in the northern FMC (D_3); 5) Early
96 syn-orogenic extensional tectonism controlled by NW–SE stretching (D_4); 6) Late Carboniferous
97 late-orogenic extensional tectonism controlled by NNE–SSW stretching (D_5).

98 The FMC is composed of five units resulting from the piling of a series of nappes during
99 the Meso-Variscan period (Devonian–Early Carboniferous; c. 350–360 Ma; D_2 event) (Ledru et
100 al., 1994; Faure et al., 2009): 1) the Upper Allochthon (called Thiviers-Payzac Unit or Gartempe
101 Unit in the North Limousin), a group of low-grade Palaeozoic terranes of limited extent; 2) the
102 Middle Allochthon (Upper Gneiss Unit; UGU; Ledru et al., 1989; Dubuisson et al., 1988; Faure et
103 al., 2009), a composite terrane formed by the "leptyno-amphibolite complex" (LAC) (e.g.
104 Santallier et al., 1988), an association of paragneisses, leptynites and amphibolites of medium to
105 high grade Barrovian metamorphism, with migmatised metagreywackes and relicts of (ultra)high-
106 pressure (UHP) eclogites ($P \sim 2.9$ GPa, $T \sim 660^\circ\text{C}$; 412 ± 5 Ma; Berger et al., 2010) and HP
107 granulites (Groupe I granulites; Pin and Vielzeuf, 1983); 3) the Lower Allochthon (Lower Gneiss
108 Unit; LGU; Ledru et al., 1989; Faure et al., 2009), consisting of metasediments (paragneisses,
109 micaschists, metashales and metagreywackes) and Late Proterozoic–Early Cambrian and
110 Ordovician orthogneisses; 4) the "Parautochthon", made of metasediments and metagranites;

111 and 5) the Southern Palaeozoic Fold-and-Thrust Belt, corresponding to continental
112 margin/platform series (e.g. Arthaud, 1970; Demange, 1994). Serpentinite and metagabbros
113 massifs of the Limousin form a thrust sheet of ophiolite-related rocks within the Middle Allochthon
114 zones, i.e. intercalated between the UGU and LGU (Girardeau et al., 1986; Dubuisson et al.,
115 1989; Berger et al., 2005).

116 Neo-Variscan stage (Late Carboniferous-Early Permian) is characterised by extensive
117 granitic magmatism and succession of extensional tectonics, characterised by NW–SE
118 stretching and NE–SW shortening (D₄ and D₅ events; Faure et al., 2009). Strike-slip motions
119 resulted in E–W folding of the piled nappes and formation of major crustal-scale shear zones
120 (Arthaud and Matte, 1975). Two exhumation stages associated with two extensional stages
121 were recorded: 1) The first exhumation stage occurred at c. 335 Ma, i.e. at the onset of the syn-
122 convergent extension that have followed the end of the nappe stacking (Faure and Pons, 1991;
123 Burg et al., 1994; Faure, 1995); 2) The second exhumation stage results from the late-orogenic
124 extension and collapse of the belt, at c. 310–305 Ma (Le Carlier de Veslud et al., 2013). A
125 widespread basic magmatism evidenced as lamprophyre dykes and related to the “basin and
126 range” extensional tectonics, also occurred during the Neo-Variscan stage (e.g. Vallance et al.,
127 2004). Medium- to low-pressure granulite facies metamorphism due to a major thermal anomaly
128 during the late stage of the Variscan orogeny led to the formation of migmatites, anatectic
129 granites and granulites at the basis of the crust (Groupe II granulites; Pin and Vielzeuf, 1983).

130 The French Massif Central (FMC) contains numerous Variscan ore deposits, some of
131 them being known since the gallo-roman times (Au, Ag, Pb, Sn) (Cauuet, 1999; Marignac and
132 Cuney, 1999). Particularly, the FMC hosts important economic to subeconomic deposits of Au,
133 U, W, Sn, Li, F, Ba, Ta, kaolin and coal, as well as minor deposits of Sb, Be, Pb, Zn, Cu and Mo
134 (Cathelineau et al., 1990; Marignac and Cuney, 1999). Most of the Variscan ore deposits were
135 formed during the late stage of uplift (i.e. during the Neo-Variscan stage) (Boiron et al., 1990,
136 1992, 2003; Marignac and Cuney, 1999; Alexandrov et al., 2000; Cuney et al., 2002).

137 In the Northern Limousin, important W-Sn, Au and U ore deposits were found (e.g.
138 Boiron et al., 1989, 1990, 2003; Cuney et al., 1990; Scaillet et al., 1996a, 1996b; Raimbault,
139 1998; Le Carlier de Veslud et al., 2000, 2013; Essarraj et al., 2001, Vallance et al., 2001, 2004;
140 Harlaux et al., 2018). While most of the W-Sn deposits in the FMC are hosted by metasediments
141 (Marignac and Cuney 1999), several minor W-Sn deposits in the Northern Limousin are hosted
142 by rare metal peraluminous granites (RMG) (Vallance et al., 2001). Emplacement of the RMG at
143 shallow depth (< 5 km) was associated to an intensive fracturing, which behaved as pathways
144 for the mineralising fluids (Le Carlier de Veslud et al., 2000). Mineralised structures are vein-
145 hosted and result from the percolation of different hydrothermal fluids during delamination
146 process of the lower lithosphere related to exhumation of the Variscan belt (Vallance et al.,
147 2001; Le Carlier de Veslud et al., 2013). Medium- to low-pressure granulite-facies
148 metamorphism at the basis of the lower crust was also source of fluids, whose migration and
149 transport were enhanced by decompression and extension processes during exhumation (Le
150 Carlier de Veslud et al., 2000; Vallance et al., 2001).

151 Most U deposits correspond to fillings of faults developed in the basement and are mainly
152 hosted by granites (e.g. Cuney et al., 1990; Cathelineau et al., 1990, 1993; Marignac and
153 Cuney, 1999). U deposits result from multi-stage processes: 1) brittle fracturing associated to
154 late magmatic shear zones during the Late Carboniferous magmatism triggered the release of
155 highly fractionated uranium-rich melts and fluids, which crystallised in the late magmatic shear
156 zones; 2) U was mobilised during hydrothermal processes related to Early Permian extensional
157 tectonics (Cuney et al., 1990; Cathelineau et al., 1990; Marignac and Cuney, 1999). Mixing of
158 reducing fluids derived from former overlying Permian sedimentary basins (half-grabens) and
159 meteoric oxidising fluids (Dubessy et al., 1987) was responsible for U leaching and re-
160 precipitation in fracture zones at low-temperature ($160 \pm 180^\circ\text{C}$: Lespinasse and Cathelineau
161 1990; El Jarray 1993).

162 Au deposits are mostly fault-sealed by quartz, hosted by metamorphic series such as

163 micaschists, orthogneiss, metasedimentary rocks (arkosic sandstones, limestones), as well as
164 by granites or volcano-sedimentary rocks (Cuney et al., 1990; Boiron et al., 1989, 2003; Bouchot
165 et al., 1989, 1997; Essarraj et al., 2001; Marignac and Cuney, 1999; Vallance et al., 2004). Au
166 was deposited in second-order faults associated to the main shear zones, active during the
167 Variscan uplift (Boiron et al., 2003; Vallance et al., 2004). Au deposits were formed by H₂O -
168 CO₂ (± CH₄ and N₂) fluids in equilibrium with the metamorphic rocks and then diluted with
169 surface fluids that have percolated during exhumation (Boiron et al., 2003).

170 In this study, we have analysed the compatible and incompatible metallic trace elements
171 by LA-ICPMS in a series of basic and ultrabasic rocks from the Limousin thrust sheet of
172 ophiolite-derived rocks, and from a series of granites as well as in the mineral assemblage in the
173 different host rocks. The main purpose consists of identifying the main hosts for the metallic
174 trace elements and determining their behaviour and budget in non-mineralised rocks, in order to
175 trace the source of trace elements transported by fluids involved in the Variscan ore deposits, at
176 different scales.

177

178 **2. Geological setting and studied samples**

179 ***2.1. Limousin thrust sheet of ophiolite-derived rocks***

180 The Limousin thrust sheet of ophiolite-derived rocks occurs as a 25 km long and 1–5 km wide
181 belt of basic and ultrabasic rocks, segmented by Variscan and late Carboniferous faults. It
182 belongs to the upper part of the Middle Allochthon and consists of remnants of an ancient
183 oceanic crust (Santallier, 1981; Dubuisson et al., 1989; Berger et al., 2005, 2006) probably
184 emplaced during the Cambro-ordovician rifting (Berger et al., 2010). Main lithologies are
185 characteristic of a lherzolite-harzburgite ophiolite type and include diopside-bearing
186 harzburgites, harzburgites, dunites, wehrlites, as well as troctolites, amphibolitic (meta)gabbros
187 and basic dykes of mid-ocean ridge basalt affinity (Berger et al., 2005, 2006). The ultrabasic
188 rocks are highly serpentinised (Dubuisson et al., 1989), with only rare relicts of olivine, spinel or

189 pyroxene. Serpentinisation of ultrabasic rocks and amphibolitisation of the basic rocks result
190 from intensive seafloor hydrothermal alteration under low-P conditions (~0.2 GPa) and
191 decreasing temperature from high-T late-magmatic conditions to greenschist–zeolite
192 metamorphic facies (Berger et al., 2005).

193 In this study, we have investigated the metallic trace element composition of a series of
194 samples from La Flotte, Le Cluzeau and Saint-Laurent (St-Laurent) ophiolite-derived massifs
195 (Fig. 2; Table 1): six serpentinites (LAU-1, LAU-2, CLUZ-6, FLOT-1, FLOT-4, FLOT2a and
196 FLOT-2b) and four amphibolites (metamorphosed basic dykes CLUZ-1, CLUZ-1a; metagabbros
197 CLUZ-4 and CLUZ-5) (El Korh et al., in press). Serpentinites from St-Laurent (LAU1, LAU2)
198 display an assemblage of Mg-rich serpentine, chlorite forming veins, spinel, amphibole
199 (composition from tremolite to Mg-hornblende) and rare relicts of olivine. These samples are
200 comparable to the highly serpentinised dunites described by Dubuisson et al. (1989).
201 Serpentinites from La Flotte (FLOT1, FLOT2a, FLOT2b) and Le Cluzeau (CLUZ6) contain
202 serpentine that have partially to totally replaced olivine and pyroxene, as well as large lenses of
203 chlorite + amphibole (tremolite to Mg-hornblende), spinel, iron oxides and sulphides (Figs. 3a–c).
204 These samples may correspond to the serpentinised harzburgites or troctolites described by
205 Dubuisson et al. (1989) and Berger et al. (2005). The major element composition of serpentinites
206 is typical of ultrabasic rocks, with low SiO₂, TiO₂, Al₂O₃, Fe₂O₃, FeO, CaO, Na₂O and K₂O
207 abundances, and a high MgO content. Serpentinites from La Flotte (FLOT1, FLOT2a, FLOT2b)
208 and Le Cluzeau (CLUZ6) show slightly higher Al₂O₃, Fe₂O₃ and FeO abundances, and a lower
209 XMg than samples from St-Laurent. Amphibolites (basic dykes CLUZ-1 and CLUZ-1a and
210 metagabbros CLUZ-4 and CLUZ-5) consist of green-brown amphibole (mainly Mg-hornblende),
211 plagioclase, chlorite and iron oxides (Fig. 3d). Amphibole often shows evidence of alteration
212 along grain rims and fractures. Amphibolites have a basic composition, with higher SiO₂, Al₂O₃,
213 CaO and Na₂O contents and a lower MgO content than serpentinites (see also El Korh et al., in
214 press).

215

216 **2.2. UHP eclogite**

217 Trace elements were also measured in one UHP zoisite-eclogite (sample L04-143; from Berger
218 et al., 2010). It crops out 500 m away from the St-Laurent ophiolite-derived massif, in contact
219 with a leucocratic orthogneiss lens (UGU), within the LGU. Formation of the magmatic
220 precursors of UHP eclogites in the Limousin has been dated between 489 and 475 Ma (U–Pb;
221 zircon) (Berger et al., 2010). Peak pressure and temperature conditions have been estimated to
222 be 660°C and 2.9 GPa and occurred at c. 412 ± 5 Ma (U–Pb age; zircon), i.e. earlier than the
223 middle Devonian continent-continent collision (Berger et al., 2010). The studied zoisite-eclogite
224 L04-143 is composed of garnet, zoisite, omphacite, and rutile. Partial retrogression is evidenced
225 by the formation of amphibole around omphacite and zoisite and ilmenite rimming rutile (for the
226 detailed description of the sample, see Berger et al., 2010).

227

228 **2.3. Granites and gneiss**

229 We have also examined the mineral trace element composition in a series of granites and gneiss
230 from the Vaulry (leucogranites VAU2 and VAU6; Vallance et al., 2001) W–Sn metallic district,
231 from the Saint-Yrieix (St-Yrieix) Au district (St-Yrieix: orthogneiss GNEISS-SY-1; Laurières:
232 leucogranite Lau10e, Essarraj et al., 2001; Cheni: leucogranite Che10b, Vallance et al., 2004),
233 and from the Saint-Sylvestre (St-Sylvestre) (IGAR744, 820A; El Jarray, 1993) and La Marche
234 (Ay.MAR; Ayt Ougougdal, 1994) leucogranites (U districts) (Figs. 3e and f).

235

236 **2.3.1. Vaulry W-Sn district**

237 The Vaulry W-Sn district is hosted by the Blond RMG pluton (Cuney et al., 1994; Raimbault,
238 1998; Le Carlier de Veslud et al., 2000; Vallance et al., 2001; Alexandrov et al., 2002). It
239 constitutes a late generation of peraluminous granites in the North-western part of the FMC
240 (Améglio et al., 1998). The Blond granite forms a 17 km long and 5 km wide, E–W elongated

241 pluton, which is located at 30 km west of Limoges (France) (Chèvremont et al., 1992). The age
242 of magmatic emplacement is estimated to be c. 313 Ma, i.e. during the late stages of the
243 Variscan magmatism ($^{40}\text{Ar}/^{39}\text{Ar}$ age on muscovite; Alexandrov et al., 2002). Intrusion of the
244 Blond granite results from the activity of the Oradour strike-slip fault, which is interpreted as the
245 root of the granite, along the SW boundary of the pluton (Soufi, 1988; Améglio et al., 1998). To
246 the North, the Blond granite intruded into the UGU, while to the south it intruded into the late
247 Devonian Cieux-Vaulry and Glane biotite-rich granites, which were emplaced into the LGU
248 (Soufi, 1988; Améglio et al., 1998; Vallance et al., 2001). Thus, it is suggested that the Blond
249 pluton is located at the boundary between the UGU and the LGU (Vallance et al., 2001). The
250 studied granites (VAU2 and VAU6) are mainly composed of quartz, K-feldspar partially
251 sericitised, plagioclase, biotite and muscovite, and contain zircon inclusions in biotite (Fig. 3e).

252 W-Sn deposits are mainly located along the eastern boundary of the Blond pluton and
253 result from fluid percolation in the northern border of the Blond pluton (Le Carlier de Veslud et
254 al., 2000; Vallance et al., 2001). They were interpreted to be synchronous to the rare metal
255 leucogranite emplacement (Alexandrov et al., 2000; Cuney et al., 2002). However, a recent
256 study has dated wolframite crystallisation (W ore) at 298.0 ± 1.2 Ma (U–Pb age) (Harlaux et al.,
257 2018). W-Sn mineralisations are hosted by subvertical quartz veins, locally with muscovite and
258 minor quartz selvages on the Vaulry area. W-Sn were deposited by a mostly aqueous fluid, with
259 moderate salinity and minor volatile component, at 315°C under hydrostatic to suprahydrostatic
260 pressures corresponding to a depth of c. 5.5 km (Vallance et al., 2001). The mineralising fluid
261 result from the mixing of different types of fluids: 1) the early high-temperature (430°C) saline
262 fluid responsible for granite muscovitisation, 2) a high-temperature ($310\text{--}350^\circ\text{C}$) and moderately
263 saline fluid with a diluted volatile component, 2) a low-temperature (230°C) and low-salinity
264 evolved meteoric fluid (Vallance et al., 2001).

265

266 **2.3.2. St-Yrieix Au district**

267 The St-Yrieix district is a major gold district located in the North-western part of the FMC (e.g.
268 Essarraj et al., 2001, Boiron et al., 2003; Vallance et al., 2004). It consists of orthogneisses and
269 micaschists of the Lower Gneiss Unit, with amphibolite or leptynite intercalations. During the
270 Neo-Variscan period, a limited granitic magmatism intruded the LGU and resulted in several
271 granite km-large stocks: Le Bourneix, Le Chalard and Chadefaine bodies consisting of the
272 biggest stocks (Stussi and La Roche 1984; Calli, 1988; Bouchot et al., 1989; Pastier 1992;
273 Chalier, 1993; Marignac and Cuney, 1999). The granite stocks and their crosscutting aplo-
274 pegmatite dykes were dated at 338.5 ± 1.5 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$ method; Alexandrov, 2000). Besides,
275 outcropping basic lamprophyre dykes were dated at 290 ± 5 Ma (Rb/Sr; Chalier et al., 1994) and
276 are probably related to the extensional tectonics during the Stephanian-Permian (Vallance et al.,
277 2004).

278 Au deposits were formed in long-lived ductile fault zones c. 310–295 Ma ago (Bouchot et
279 al., 1989, 1997; Marignac and Cuney, 1999; Le Carlier de Veslud et al., 2000). They were
280 emplaced under P–T conditions of 350–400°C and 100–200 MPa, to surface conditions and
281 recorded a pressure drop from lithostatic to hydrostatic conditions (e.g. Boiron et al., 1990;
282 Marignac and Cuney, 1999, Boiron et al., 2003). Au was deposited mainly as free gold and at a
283 lesser degree as gold-rich arsenopyrites by $\text{H}_2\text{O} - (\text{CO}_2 \pm \text{CH}_4 \text{ and } \text{N}_2)$ fluids in equilibrium with
284 the metamorphic rocks mixing with low-salinity surface water of meteoric origin percolating the
285 basement rocks at low temperatures (150–250°C) under hydrostatic pressure conditions
286 (Cathelineau et al., 1993, Essarraj et al., 2001, Boiron et al., 2003).

287 The studied granites and orthogneiss are composed of quartz, K-feldspar (including
288 microcline), plagioclase (albite), biotite and muscovite. Biotite is partially altered in chlorite in
289 sample Che10b. Accessory minerals consist of apatite and zircon inclusions in biotite crystals
290 (Table 1). Muscovite is absent from sample LAU10e (Table 1), which also contain accessory
291 pyrite and Fe-Ti oxides. K-feldspars are often altered in sericite. Hydrothermal alteration is also
292 evidenced by the formation of fine Fe hydroxides/clay in association with sericite.

293

294 **2.3.3. St-Sylvestre–Western Marche complex**

295 The St-Sylvestre–Western Marche two-mica peraluminous leucogranites host large U deposits
296 in the Western FMC (e.g. Cuney et al., 1990; Cathelineau et al., 1990; Marignac and Cuney,
297 1999). The leucogranitic complex consists of a 2–3 km thick laccolith of 6 km deep roots
298 maximum, emplaced at a depth of 10.5 ± 1.5 km (Scaillet et al., 1996a). The magmatic
299 emplacement has been dated at 324 ± 4 Ma (U–Pb ages on zircon and monazite) (Holliger et
300 al., 1986), i.e. during the Early Carboniferous syn- to late-collision exhumation (Le Carlier de
301 Veslud et al., 2013). Younger $^{40}\text{Ar}/^{39}\text{Ar}$ ages from 315 Ma to 300 Ma towards the base of the
302 complex reflect a second exhumation stage, responsible for the uplift of the St-Sylvestre
303 complex by unroofing of the overlying metamorphic and granitoidic formations during the late-
304 orogenic Variscan extension (Scaillet et al., 1996a, 1996b; Le Carlier de Veslud et al., 2013).
305 Intensive deformation associated with fluid migration during exhumation of the ductile crust was
306 dated at 294 ± 19 Ma (Rb–Sr isochron age) (Luais et al., 2009). U enrichment in the
307 peraluminous granites results from multi-stage magmatic processes (partial melting and granite
308 intrusion) followed by hydrothermal remobilisation (Cuney et al., 1990). The formation of the
309 main hydrothermal U deposits was dated at 280 Ma (U–Pb age) (Cathelineau et al., 1990).
310 Outside the mineralised zones, three types of fluids have percolated into the St-Sylvestre
311 granitic complex during the late stages of the Variscan orogeny: 1) high-T aqueous-carbonic
312 fluids, 2) lower-T aqueous fluids probably related to the granite episyenitisation, and 3) low-T
313 temperature aqueous fluids (El Jarray et al, 1994; André et al., 1999). Minor W and Sn
314 mineralisation are also hosted by the St-Sylvestre granites, and are located along the
315 granite/metamorphic rock contact (Cuney et al., 1990). The presence of both major U and minor
316 W–Sn ore deposits result from different metal fractionation stages related to the different
317 Variscan magmatic stages (Cuney et al., 1990).

318 While this study does not consider U sources, granites from St-Sylvestre and La Marche
319 were analysed to determine if they may be potential source of minor deposits. The studied
320 granite samples from La Marche (Ay.MAR) and St-Sylvestre (820A, IGAR744) are composed of
321 quartz, K-feldspar, plagioclase, biotite and muscovite (Fig. 3f). Accessory minerals consist of
322 apatite as well as zircon and monazite inclusions in biotite crystals (Table 1). K-feldspar and
323 plagioclase have undergone partial sericitic alteration.

324

325 **3. Analytical techniques**

326 **3.1. Whole rock trace element composition**

327 The whole rock trace element concentrations were analysed at the SARM (CRPG-Nancy) by
328 inductively coupled plasma mass spectrometry (ICPMS), using a ThermoFisher X7 spectrometer
329 (ThermoFisher Scientific, Germany and USA). 200 mg of powdered samples were fused with
330 600 mg LiBO₂ at 980°C, and dissolved in a mixture of HNO₃ (5%), H₂O₂ (0.5%) and glycerol
331 (10%) prior to analysis, according to the procedure described by Carignan et al. (2001).
332 Uncertainties are generally <8% at 1σ.

333

334 **3.2. Major element composition of minerals**

335 Major element compositions of minerals were analysed by wavelength-dispersive analysis using
336 a five-spectrometer CAMECA SX100 electron microprobe at the Service Commun de
337 Microscopie Electronique et Microanalyses (SCMEM, GeoRessources). The accelerating
338 voltage and the beam current were 15 kV and 12 nA, respectively. Counting times were 10 s.
339 The beam size was set to 1 μm. Si, Al, Mg and Na were analysed on a TAP crystal. Fe and Mn
340 were run on a LIF crystal, Ca and Cr on a PET crystal. K, Ti and Cl were measured with a LPET
341 crystal. Raw data were reduced using the X-PHI quantification procedure (Merlet, 1994). All iron
342 was calculated as Fe²⁺ in chlorite, olivine, phengite and feldspaths. Fe²⁺ and Fe³⁺ in amphibole
343 and serpentine, as well as water contents (chlorite, micas, amphibole, serpentine), were

344 calculated assuming ideal stoichiometry. Representative major element compositions of
345 minerals in the studied basic and ultrabasic samples and gneiss GNEISS-SY-1 are given in the
346 Supplementary Table S1.

347

348 **3.3. Trace element composition of minerals**

349 Trace element abundances in minerals were analysed by laser-ablation inductively-coupled
350 plasma mass spectrometry (LA-ICPMS) at GeoRessources. Analyses were carried out using an
351 Agilent 7500c quadrupole spectrometer interfaced to a GeoLas Pro 193 nm ArF excimer laser
352 ablation system (Lambda Physik, Germany). Operating conditions consisted of a 5 Hz repetition
353 rate, a $\sim 10 \text{ J/cm}^2$ fluence, and a 44 μm beam size. Helium was used as the carrier gas (0.8
354 L/min) and was mixed with Ar prior to its introduction in the plasma (1.5 L/min). Peak hopping
355 mode was employed, as well as dual (counting and analogue) secondary electron multiplier
356 (SEM) detector mode. Analytical list included the following isotopes: ^{27}Al , ^{29}Si , ^{42}Ca , ^{49}Ti , ^{45}Sc ,
357 ^{51}V , ^{53}Cr , ^{55}Mn , ^{57}Fe , ^{59}Co , ^{60}Ni , ^{63}Cu , ^{66}Zn , ^{69}Ga , ^{74}Ge , ^{75}As , ^{95}Mo , ^{107}Ag , ^{118}Sn , ^{121}Sb , ^{182}W ,
358 ^{197}Au , ^{208}Pb , ^{209}Bi . Dwell times were 10 ms for major elements and 20 ms for the trace elements.
359 The ThO^+/Th^+ and $\text{Ca}^{2+}/\text{Ca}^+$ ratios were carefully optimised to about 0.2–0.7 and 0.3–0.5%
360 respectively, by ablation of the National Institute of Standards and Technology standard
361 reference material SRM 610, a synthetic glass doped with trace elements. The low oxide and
362 doubly charged ion production rates prevented polyatomic and doubly charge ion interferences.
363 Acquisition times for background and ablation signals were 60 and 40 s, respectively, allowing
364 the measurement of duplicate spots per analysis.

365 The NIST 610 reference material was used for external standardisation, using the
366 preferred values of average element abundances from the GeoRem database (Max Planck
367 Institut für Chemie, Mainz, Germany; <http://georem.mpch-mainz.gwdg.de>). Typical analytical
368 series consisted of 10–16 measurements of mineral samples, bracketed at the start and end of
369 each analytical sequence by 2 measurements of the primary standard. Intensity vs. time data

370 were reduced using the ratio-of-the-mean intensities method in an Excel spreadsheet, following
371 the procedure defined in Longerich et al. (1996). Internal standards were ^{29}Si (amphibole,
372 serpentine, plagioclase, olivine, chlorite, garnet, omphacite, quartz, biotite, muscovite), ^{57}Fe (Fe
373 oxides and sulfides), ^{27}Al or ^{53}Cr (spinel) and ^{49}Ti (rutile), based on the mineral major element
374 compositions determined by microprobe analysis. During data treatment, intervals of 45 and 25 s
375 were selected for the background and the ablation signal, respectively, with an offset of 3 s from
376 the beginning of the ablation to avoid unstable parts of the signal, resulting from large particle
377 production and/or surface contamination. No correction for down-hole fractionation was applied.
378 The instrumental mass bias was corrected using a linear regression function, considering an
379 equal time interval between each analysis. The minimum detection limit, corresponding to three
380 times the standard deviation of the background measurement, is based on the 99% confidence
381 level over the Gaussian counting statistics. Limits of detection are generally 0.1–5 ppm for V,
382 Co, Ga, Ge, Mo, Ag, Sn, Sb, W, Au, Pb and Bi, 1.2–20 ppm for Cu, Zn and As, 1.5–50 ppm for
383 Cr, Ni and Ti. Uncertainties on individual measurements at 1σ are 1.1–15%, and up to 30% for
384 elements, whose abundance is <0.5 ppm. The NIST SRM 612 glass was employed as
385 secondary standard and yielded trace element abundances in agreement with the reference
386 values (see Supplementary Table S2).

387

388 **4. Results**

389 **4.1. Serpentinites**

390 **4.1.1. Bulk rocks**

391 Serpentinites display high Cr, Co and Ni abundances (2436–3427, 87–137 and 1326–1992 ppm,
392 respectively), and to a lesser extent, high V and Zn concentrations (36–64 and 44–81 ppm,
393 respectively) (Table 2; Figs. 4 and 5). The compatible and incompatible trace element
394 concentrations in serpentinites are generally typical of the Primitive Mantle (PM; "pyrolite" in
395 McDonough and Sun, 2005) (Table 2). The studied serpentinites harzburgites have a $(\text{Ce}/\text{Yb})_{\text{PM}}$

396 ratio of 0.40 to 0.70, while the light REE (LREE) abundances are below detection limit in
397 serpentinites from St-Laurent LAU1 and LAU2. In a PM-normalised multi-element diagram,
398 serpentinites show a flat pattern for medium rare earth elements (MREE) and heavy REE
399 (HREE). Serpentinites LAU1 and LAU2 are depleted in MREE compared to HREE. Th, U, High
400 Field Strength Element abundances (HFSE) are relatively variable: Th, U, Nb and Ta
401 abundances are below detection limit in all serpentinites, while Zr, Hf and Ti are only measured
402 in the serpentinites from La Flotte and Le Cluzeau (Fig. 4a). Ti and Cu are abundant in
403 serpentinites FLOT2a, FLOT2b and CLUZ6 (378–731 and 119–326 ppm, respectively), while
404 they are generally below detection limit in serpentinites LAU1 and LAU2. However, Ti is depleted
405 and Cu is strongly enriched in the serpentinites compared to the Primitive Mantle (Table 2; Fig.
406 4b). Ga and Ge abundances in serpentinites FLOT2a, FLOT2b and CLUZ6 (3.7–4.3 and 0.89–
407 1.2 ppm, respectively) are low but similar to the Primitive Mantle (4.0 ppm Ga and 1.1 ppm Ge).
408 Lower Ga and Ge are measured in the serpentinites LAU1 and LAU2 (1.6–1.8 and 0.82–0.88
409 ppm, respectively). Mo, Ag, Sn, Au, Pb and Bi concentrations are generally below detection limit
410 (Table 2). When detected, As, Sb, W contents in serpentinites are enriched compared to the
411 Primitive Mantle.

412

413 **4.1.2. Minerals**

414 In serpentinites LAU1 and LAU2, serpentine is mostly magnesian, but its Mg# ratio decreases in
415 altered zones (see also El Korh et al., in press). Serpentine in the serpentinites FLOT2a,
416 FLOT2b and CLUZ6 is generally zoned in Fe₂O₃, FeO and MgO, and show a rimward increase
417 of the Mg# (from 0.7 to 1.0). Relicts of olivine are dominated by the forsterite component (Mg# of
418 0.86–0.91). When present, chlorite and amphibole are magnesian. Spinel shows a large range
419 of composition, with variable Al₂O₃, Cr₂O₃, Fe₂O₃, FeO, MgO contents.

420 Spinel is the main host for compatible trace metals (up to 0.1–1% Ti, V, Co, Ni, Cu and
421 Zn; > 1% Cr). Cr, V and Ti abundances in chlorite and amphibole (Cr: 546–12280 ppm; V: 8.7–

422 188 ppm; Ti: 143–1808 ppm) are generally higher than in serpentine and olivine (Cr: 3.6–100
423 ppm; V: 0.53–8.6 ppm; Ti: 7.8–107 ppm) (Fig. 5), except for samples FLOT2a and LAU2 (Fe-rich
424 generation), whose high concentrations in Cr, V and Ti (up to 221 ppm, 10209 ppm and 7360
425 ppm of Cr, V and Ti, respectively) may be related to small inclusions of spinel (Table 3). By
426 contrast, serpentine and olivine display higher Co, Ni and Cu abundances (Co: 13–905 ppm; Ni:
427 479–6764 ppm; Cu: 1.3–1360 ppm) than amphibole and chlorite (Co: 26–108 ppm; Ni: 494–
428 1481 ppm; Cu: 1.25–176 ppm). Zn, Ge and As concentrations are within the same range for
429 serpentine, chlorite, amphibole and olivine (3.7–321, 0.33–5.1 and 1.3–31 ppm, respectively),
430 while Ga is mainly hosted by amphibole (10–30 ppm), chlorite (1.2–20 ppm) and spinel (5.2–125
431 ppm) (Fig. 5; Table 3).

432 Serpentine generally shows low abundances of Sn, Sb, W and Pb (Sn and Sb < 1 ppm;
433 Pb and W < 4 ppm), with the highest values being measured in the Fe-rich generation (up to 7.9
434 ppm W). Au, Mo, Ag and Bi are generally below detection limit, even if low concentrations of Mo,
435 Ag and Bi may be detected in chlorite, serpentine or spinel.

436 Accessory Fe oxides and sulphides (pyrite) may have high and heterogeneous contents
437 in compatible metals (up to 12263 ppm Cr, 2127 ppm Co, 79693 ppm Ni and 5508 ppm Cu),
438 which may be related to small-size inclusions. However, their concentration in incompatible
439 metals is generally low.

440

441 **4.2. Basic rocks**

442 **4.2.1. Bulk rocks**

443 Amphibolites show high bulk abundances of compatible metals (30 to 2000 ppm), with higher
444 Sc, Ti and V contents, and lower Cr, Co, Ni and Zn contents than ultrabasic rocks. Cu
445 abundances are lower than those of serpentinites from La Flotte and Le Cluzeau, but higher
446 than those of serpentinites from St-Laurent. Amphibolites display REE abundances similar to the
447 Primitive Mantle. In a PM-normalised multi-element diagram, amphibolites show a flat MREE

448 and HREE pattern and LREE depletion, with a $(Ce/Yb)_{PM}$ ratio of 0.60 to 0.66, typical of normal
449 mid-ocean ridge basalt (N-MORB) (Fig. 4a). Large Ion Lithophile Elements (LILE: Cs, Rb, Ba)
450 and Sr abundances are slightly higher than N-MORB (Sun and Mc Donough, 1989) and
451 Primitive Mantle. Two amphibolite samples show a positive Pb anomaly. The compatible and
452 incompatible metal concentrations as well as the HFSE abundances are generally typical of N-
453 MORB and ocean floor basalts (OFB; Jenner and O'Neill, 2002). Only Ti and Zn are depleted
454 compared to mean OFB, while Cr shows a higher abundance (Table 2; Fig. 4b). Ga and Ge
455 contents of amphibolite facies rocks (12–14 and 1.2–1.4 ppm, respectively) are higher than
456 those of serpentinites. Ge content is typical of OFB (1.1–1.5 ppm; Jenner and O'Neill, 2012), but
457 Ga content is slightly lower than in OFB (14–23 ppm; Jenner and O'Neill, 2012). Sn, Sb, W, Mo,
458 Ag, Au and Bi abundances are generally below detection limit, though higher than mean OFB
459 when measured (Table 2).

460

461 **4.2.2. Minerals**

462 In amphibolites, green-brown amphibole is mainly Mg-hornblende (see also El Korh et al., in
463 press). Plagioclase composition varies from anorthite to labradorite in samples CLUZ1 and
464 CLUZ4, and is mostly labradorite in samples CLUZ3 and CLUZ5. In amphibolite facies
465 metagabbros and dykes, amphibole and accessory titanite are the main hosts for the compatible
466 metals (24–1775 ppm of V, Cr, Ni; 1.4–93 ppm of Sc, Co, Cu, Zn). In amphibole, Ga, Ge and As
467 abundances are relatively variable (2.3–17 ppm, 1.2–2.7 ppm and 1.3–8.8 ppm, respectively),
468 while Sn and Sb contents are generally < 1 ppm (Table 4; Fig. 6). Titanite is the main host for Sn
469 (4.7–5.6 ppm), W (16–18 ppm) and Ge (3.5–5.4 ppm). Accessory sulphides and oxides show
470 high and heterogeneous concentrations in compatible and incompatible elements (up to 7180
471 ppm of Cr, 22078 ppm of Cr, 89963 ppm of Cu, and 99 ppm Ag). Plagioclase is the main host for
472 Ga (10–21 ppm) and Pb (0.46–15 ppm), while it displays low abundances of other compatible
473 and incompatible metals (Table 4).

474

475 **4.3. UHP eclogite**

476 **4.3.1. Bulk rock**

477 The zoisite-bearing eclogite have higher TiO₂, Al₂O₃, FeO contents and lower SiO₂, MgO and
478 CaO contents than amphibolites. Eclogite L04-143 shows the highest Sc, Ti, V, Zn and Pb
479 contents, although similar to those of OFB, and the lowest Cr, Co, Ni contents of the studied
480 basic and ultrabasic rocks. Sn, Sb and Bi abundances are low but well measurable (≤ 0.5 ppm),
481 while the W content is high (14 ppm) (Table 2). Ga and Ge are slightly higher than in amphibolite
482 facies dykes (17 and 2.3 ppm, respectively) and metagabbros, although within the range of
483 basic rocks (Ga: 8.7–23 ppm, Ge: 0.7–2.3 ppm; see El Korh et al., 2017 and references therein
484 for a review).

485

486 **4.3.2. Minerals**

487 In the zoisite-eclogite, rutile and its ilmenite rims are the main host for Ti, V and Cr. Significant V
488 and Cr amounts (56–986 ppm and 33–498 ppm, respectively) are also measured in the other
489 minerals of the assemblage (omphacite, amphibole, zoisite, and, to a lesser extent, garnet)
490 (Table 5; Fig. 7). Concentrations in Sc are relatively similar in all minerals (15–69 ppm). Sc, Co,
491 Ni and Zn are abundant in all minerals. The highest Ni concentrations are measured in
492 omphacite and retrograde amphibole, while Co is mainly enriched in amphibole and garnet (Fig.
493 7). Cu abundances are below 21 ppm in rutile, zoisite and amphibole. One measurement of
494 omphacite has yielded a concentration of 89 ppm. Zoisite is the main host for Ga and Ge (111–
495 186 and 8.8–19 ppm, respectively). Omphacite, garnet and retrograde amphibole are other
496 important hosts for Ga and Ge (21–41 and 3.0–21 ppm, respectively). The incompatible metals
497 are mainly concentrated into zoisite, rutile and retrograde amphibole (generally 1–20 ppm As,
498 Mo, Bi, Sn, Sb and W). Rutile contains significant W abundances (up to 164 ppm). Pb
499 abundances are also high in zoisite (75–98 ppm) and, to a lesser extent, in omphacite and

500 amphibole (6.1–39 ppm) (Table 4).

501

502

503 **4.4. Granites and gneisses**

504 **4.4.1. Bulk rock**

505 The bulk rock major and trace element concentrations were measured in the gneiss GNEISS-
506 SY-1. It is a high-SiO₂ orthogneiss, with high Al₂O₃ and K₂O contents and low FeO, Fe₂O₃ and
507 CaO contents (Table 2). In a PM-normalised multi-element diagram (Fig. 4a), sample SY-
508 GNEISS-1 display a pattern typical of the Upper Continental Crust (UCC), with Ti, Nb and Ta
509 negative anomalies, Pb positive anomaly, and LREE > HREE [(Ce/Yb)_{PM} = 4.0]. Its Cs, V, Cr, Co
510 and Ni contents are low (V: 17 ppm; Sc, Cr, Co, Ni < 5.6 ppm) compared to the UCC, as well as
511 compared to the basic and ultrabasic rocks. Its Ti content is comparable to those of basic
512 amphibolites. Ga and Ge contents (20 and 1.6 ppm, respectively) are comparable to those of
513 UCC and granites (see El Korh et al., 2017 and references therein for a comparison) and slightly
514 higher than those of basic amphibolites (Table 2). The gneiss GNEISS-SY-1 also shows higher
515 Mo, Sn, Sb, W and Pb abundances than those of basic and ultrabasic rocks, but comparable to
516 those of the UCC (Table 2; Fig. 4b).

517

518 **4.4.2. Minerals**

519 In granite and gneisses, the compatible and incompatible metals are mainly hosted by biotite
520 and muscovite. Biotite (V: 3.1–447 ppm; Cr: 5.0–449 ppm; Co: 2.5–45 ppm; Ni: 3.0–98 ppm; Zn:
521 352–2902 ppm; Ti: 7976–21870 ppm) generally shows higher compatible metal abundances
522 than muscovite (V: 1.4–153 ppm; Cr: 8.4–127 ppm; Co: 1.2–4.3 ppm; Ni: below detection limit;
523 Zn: 12–466 ppm; Ti: 417–5036 ppm) (Fig. 8). The micas have variable contents in Sc (5.5–255
524 ppm) and Ga (61–282 ppm) and low contents in Ge, As, Sb, Pb and Bi (< 17 ppm). Cu (1.5–42
525 ppm) is mainly measured in biotite from Vaulry granites. Besides, micas display significant Sn

526 and W contents (11–610 and 1.9–193 ppm, respectively), W being enriched in muscovite (2.0–
527 193 ppm) compared to biotite (1.9–77 ppm) (Table 6).

528 Sn and W abundances significantly vary in biotite and muscovite between the different
529 localities (Fig. 9). The highest Sn and W abundances are measured in Vaulry granites (VAU2
530 and VAU6), and decrease in the granites from La Marche (Ay.MAR), St-Sylvestre (820A and
531 IGAR744) St-Yrieix (St-Yrieix: GNEISS-SY-1, Lauri ras: LAU10e, Cheni: Che10b) districts,
532 which show the lowest Sn and W content. In the St-Yrieix district, biotite from gneiss GNEISS-
533 SY-1 has higher W contents than biotite in granites Che10b and LAU10e, while biotite in sample
534 Che10b shows higher Sn abundances. Zn and Ge abundances decrease from Vaulry to St-
535 Yrieix granites in biotite and muscovite (Fig. 9; Table 6). On the other hand, Vaulry and St-Yrieix
536 granites and gneisses have higher Sc contents than St-Sylvestre and La Marche granites. In the
537 St-Yrieix district, Ni, Cr and V contents of biotite decrease from Lauri ras to St-Yrieix and Cheni
538 (Fig. 9; Table 6). Samples from La Marche also show high V and Cr concentrations. By contrast,
539 samples from the Vaulry district variable Cr and V contents from one sample to the other.

540 Au and Ag abundances are generally below detection limit in all minerals. Quartz,
541 feldspars and apatite have low contents in compatible metals, and only show minor amounts of
542 Sc, Ga, Ge, Sn and Pb, when detected. High Ti and W concentrations are also measured in
543 accessory Fe-Ti oxides (sample LAU10e) (Table 6; Fig. 8).

544

545 **5. Discussion**

546 **5.1. Trace element behaviour and mobilisation during high-T hydrothermal alteration in** 547 **the Limousin ophiolite-derived rocks**

548 **5.1.1. *Serpentinites***

549 The major and trace element (including compatible elements, Ga, Ge, HFSE and REE)
550 composition of serpentinites are generally typical of ultrabasic rocks and Primitive Mantle (Fig.
551 4a). During serpentinisation of peridotites and dunites, Sc, Ti, V, Cr, Co, Ni, Cu and Zn

552 contained in olivine and spinel were mainly incorporated in chlorite, Fe-rich serpentine and Fe-
553 poorer serpentine formed with decreasing temperature of alteration, as well as in the less
554 abundant amphibole and accessory oxides and sulphides.

555 As, Ag, Sn, Sb, W, Pb and Bi are enriched in the serpentinites compared to the Primitive
556 Mantle and mean OFB (Fig. 4b). It has been shown that As and Sb can be mobile and enriched
557 in the oceanic crust during hydrothermal alteration (Spandler et al., 2004). Similarly, precipitation
558 of oxides and sulphides during the early stages of hydrothermal alteration may be responsible
559 for metal enrichment (e.g. Rouxel et al., 2003). As, Sn, Sb, Sn, Pb and Bi are distributed
560 between all mineral phases, i.e. in olivine relicts, as well as in the newly-formed serpentine,
561 chlorite, amphibole and oxides. Within each sample, higher metal concentrations are measured
562 in the earlier Fe-serpentine generation, and decrease in the next Fe-poorer generation due to
563 their lesser availability as hydrothermal alteration went along.

564 Two serpentinites (FLOT1, FLOT2a) contain olivine relicts. In both samples, Ag, Bi and
565 Mo are only detected in serpentine and accessory sulphides and oxides. In sample FLOT1, Sn
566 abundances are slightly higher in chlorite and amphibole than in olivine relicts and serpentine,
567 while W is enriched in serpentine and amphibole compared to olivine. Sb abundances are
568 similar in olivine relicts and in serpentine. Besides, As is enriched in olivine relicts compared to
569 newly-formed minerals. In sample FLOT2a, both olivine relicts and serpentine contain As and
570 Sn, with higher Sn concentrations measured in serpentine. W is mainly hosted by serpentine.
571 The high As concentrations measured in olivine relicts and similar Sn abundances in olivine and
572 serpentine may reflect: 1) Enrichment of the initial magmatic rock or, 2) Contamination of olivine
573 through fracturation during serpentinisation. In other serpentinites (LAU2, FLOT2b, CLUZ6),
574 serpentine is the main host for As, Ag, Mo, Sn, Sb, W and Bi, indicating that enrichment in
575 incompatible metals results from the high-temperature hydrothermal alteration that have followed
576 the formation of the oceanic crust (Berger et al., 2005, 2006).

577

578 **5.1.2. Amphibolites**

579 Amphibolite facies metagabbros and dykes display a major and trace element (including
580 compatible elements, HFSE and REE) composition typical of N-MORB and OFB (Fig. 4a).
581 Despite lower than those of OFB, Ga abundances are consistent with those of basaltic and
582 gabbroic rocks (8.8–41; see El Korh et al., 2017 and references therein). Besides, enrichment in
583 mobile LILE, Sr and Pb compared to N-MORB and PM results from seafloor hydrothermal
584 alteration.

585 During amphibolitisation of gabbros and basic dykes, the transition metals (MgO, MnO,
586 FeO, Sc, Ti, V, Cr, Co, Ni, Cu and Zn) were mainly incorporated into amphibole, titanite and
587 accessory oxides and sulphides. Because of the strong affinity of Ga with Al (De Argollo and
588 Schilling, 1978), Ga was preferentially incorporated into Al-rich plagioclase. Ca-plagioclase is
589 also the main host for Pb. Titanite, a nesosilicate, is the main host for Ge, confirming the
590 preferential incorporation of Ge in the less polymerised silicates (Bernstein, 1985; El Korh et al.,
591 2017; Rouxel and Luais, 2017). During amphibolite facies metamorphism, Sn, Sb and W were
592 preferentially into titanite and, to a lesser extent, into amphibole. Amphibole is the main host for
593 As, while Ag and Bi are mostly hosted by sulphides and oxides. No relicts of primary
594 assemblages being preserved, it is difficult to determine whether As, Ag, Sn, Sb, W and Bi
595 enrichment is of magmatic or hydrothermal origin. However, by analogy with serpentinites,
596 enrichment in As, Ag, Sn, Sb, W and Bi compared to OFB and PM is probably related to the
597 high-temperature hydrothermal processes on seafloor.

598

599 **5.1.3. UHP eclogite**

600 The studied zoisite-eclogite L04-143, derives from an iron-rich plagioclase cumulate, which has
601 probably been strongly metasomatised under UHP conditions (Berger et al., 2010). Its
602 composition present characteristics of both UCC and ocean island basalts (OIB), despite lower
603 LILE, Th and U contents (Fig. 4a). For this reason, the initial setting of the magmatic protolith is

604 difficult to be determined based on the trace element composition (continental intrusions or Fe–
605 Ti oceanic gabbros) (Berger et al., 2010). While the major element composition of eclogite L04-
606 143 – characterised by a low SiO₂ content and high TiO₂, Al₂O₃ and Fe contents – is different
607 from those of amphibolites, its composition in compatible transition trace metals, Ga and Ge is
608 similar to amphibolites and OFB, suggesting an oceanic affinity (Fig. 4b). By contrast, eclogite is
609 enriched in As, Sb, W, Pb and Bi compared to PM and OFB, which may reflect either the primary
610 magmatic composition, or enrichment during hydrothermal processes.

611 Most of the transition trace metals (Sc, Ti, V, Cr, Co, Zn) are distributed between all
612 mineral phases, contrary to Cu, which is generally not measured in garnet and omphacite. Main
613 UHP hosting minerals for Cu are zoisite and rutile. The high Cu abundance (89 ppm) measured
614 in one omphacite crystal is probably related to a Cu-rich inclusion, such as rutile. Similarly, Ni is
615 not detected in garnet and rutile, and was mostly hosted by omphacite and zoisite during the
616 UHP stage. Partial retrogression of eclogite L04-143 was responsible for the transfer of a part of
617 incompatible metals released by zoisite and omphacite to retrograde amphibole.

618 During UHP metamorphism, Ge was mainly incorporated into garnet and zoisite (an
619 epidote-group mineral), as observed by El Korh et al. (2017) for HP rocks, while main hosts for
620 Ga were Al-rich zoisite, omphacite and garnet. The incompatible metals As, Ag, Mo, Sn, Sb, W
621 Pb and Bi were mainly distributed between rutile, zoisite, and, to a lesser extent, omphacite.
622 During retrogression, Sn, Sb and W released by rutile and zoisite were incorporated into
623 amphibole and ilmenite, while Ga, Ge, As, Mo, Pb and Bi were only incorporated into retrograde
624 amphibole.

625

626 **5.2. Trace element partitioning in gneiss and granites**

627 Orthogneiss GNEISS-SY-1 from the St-Yrieix district belongs to the Lower Gneiss Unit (Lower
628 Allochthon). Its major element, LILE, HFSE and REE compositions are typical of granitic rocks
629 from the Upper Continental Crust. Trace element compositions of the mineral assemblage show

630 that micas (biotite and muscovite) are the main hosts for the transition metals, as well as for Ga,
631 Ge, As, Ag, Sn, Sb, W and Pb. Because of the importance of micas in the incompatible metal
632 budget for gneiss GNEISS-SY-1, focus has been placed on biotite and muscovite in the granite
633 samples to discuss the provenance of the metallic pre-concentration.

634 In a previous study, Wolf et al. (2018) have highlighted the importance of muscovite and
635 biotite in trace element partitioning in migmatites during melting. In particular, Sn remains into
636 the restite and becomes depleted in the melt as long as micas are stable in the restite during
637 partial melting. Feldspaths (microcline and albite) are the main hosts for Pb and contain high Ga
638 abundances, as well as low Sc, Ge and Sn contents. However, it is difficult to attribute measured
639 concentrations to pristine magmatic contribution, or if they were modified by hydrothermal
640 processes related to metamorphism. Trace element compositions of micas within each sample
641 are relatively homogeneous, while they vary significantly according to the locality. The
642 geographical variation is particularly evident for Sn and W as well as for Zn, Ge and Cu, whose
643 abundance generally decrease from Vaulry to La Marche, St-Sylvestre and St-Yrieix. The V and
644 Cr abundances in biotite also vary from one locality to the other. In the St-Yrieix district, the V
645 and Cr abundances increase as the distance to the ophiolite-derived rocks decreases,
646 suggesting a local source for Cr and V.

647

648 **5.3. Tracing the sources of metallic pre-concentrations in North-western Limousin**

649 **5.3.1. Gold**

650 Mineralised Au-bearing quartz veins in the Limousin area are mostly hosted in the Lower
651 Allochthon (Lower Gneiss Unit). However, the sources of Au in the Variscan belt are not known
652 precisely and were suggested to derive from the structurally overlying Upper Gneiss Unit
653 (Marignac and Cuney, 1999). Different sources were considered for the Limousin Au deposits:
654 Proterozoic volcano-sedimentary rocks (Braux et al., 1991); Cambro-ordovician clastic rocks
655 (Leblanc, 1989) or amphibolites and ultrabasic rocks from the LAGs (Floc'h et al., 1984). Based

656 on Pb isotope analyses, Marcoux and Bonnemaïson (1988) have suggested a mantle origin for
657 Au. During this study, Au abundances were below detection limit (0.1–1 ppm) in minerals from
658 ultrabasic and basic rocks of the Limousin ophiolite-derived rocks (Upper Gneiss Unit). Our data
659 do not allow determining any source for Au in the ultrabasic and basic rocks of the Limousin
660 thrust sheet of ophiolite-derived rocks.

661 The Lower Gneiss Unit is considered as possible sources for Au deposits in the St-Yrieix district
662 (Cuney et al., 1990). However, the measured Au abundances were below detection limit in the
663 mineral assemblage of gneiss GNEISS-SY-1 (Lower Gneiss Unit). Further systematic studies on
664 accessory minerals from the lower gneiss unit are needed to trace the source of Au. In addition,
665 based on our results, the granitic rocks are not the direct sources for Au, but served as heat
666 source for the circulation of Au-mineralising hydrothermal fluids in faults adjacent to granites
667 (Boiron et al., 2003).

668

669

670 **5.3.2. Tungsten and tin**

671 In the North-western Limousin, the Blond leucogranite (Vaulry district) hosts major W-Sn
672 deposits, while contemporaneous leucogranites from the St-Sylvestre–Western Marche U district
673 also contains minor W-Sn (Fig. 10). Different sources were regarded for W and Sn ore deposits
674 in the FMC (Marignac and Cuney, 1999). While crustal or mantellic sources (granites and LAG)
675 have been discussed, mobilisation of W and Sn is thought to be the result of the percolation of
676 fluids of metamorphic origin (Caron, 1994; Vallance et al., 2001). In the Montagne Noire and
677 Cévennes (southern FMC), metapelites were considered as the primary source of W, which was
678 remobilised during hydrothermal processes related to subsequent LP–HT metamorphism (Gibert
679 et al., 1988, 1992). More recently, the relationship between granites and W–Sn ore deposits has
680 been pointed up: granites are thought to represent either the direct metal source or the thermal
681 source that has driven hydrothermal circulation of mineralising fluids (Vallance et al., 2001;

682 Harlaux et al., 2018). Pb isotopes has allowed the determination of different sources for W-Sn
683 deposits according to their emplacement: 1) deposits in Palaeozoic rocks derive from
684 sedimentary, granitic and metamorphic rocks from old cratonic areas, i.e. from recycled
685 Gondwanian crust; 2) deposits in Devonian rocks derive from more juvenile material (Harlaux et
686 al., 2018).

687 In the North-western Limousin, amphibolites from the St-Yrieix district (LAG; UGU)
688 display relatively high W and Sn contents (1–17 ppm and 5–24 ppm, respectively; Calli, 1988).
689 W and Sn enrichment may have been caused by the same hydrothermal processes that have
690 altered the amphibolites and triggered Au, As, Co and V enrichment (Calli, 1988). By contrast,
691 serpentinites and amphibolites from the Limousin thrust sheet of ophiolite-derived rocks are
692 characterised by low Sn and W abundances in whole rocks and minerals. Despite intensive fluid-
693 rock interaction, the high-T hydrothermal alteration responsible for serpentinisation and
694 amphibolitisation that have followed magmatic emplacement was not responsible for any
695 significant W and Sn pre-concentration. Moreover, these hydrothermal processes cannot be
696 responsible for W and Sn transport to the mineralised veins, because they occurred after the
697 formation of the magmatic precursors of serpentinites and amphibolites, i.e. c. 100 Ma earlier
698 than the mineralisation age (Berger et al., 2010; Harlaux et al., 2018). For this reason, the
699 Limousin thrust sheet of ophiolite-derived rocks, deriving from a mantle source (such as a mid-
700 ocean ridge–lherzolite-harzburgite ophiolite type; Berger et al., 2006), cannot be considered as
701 an effective source for W and Sn mineralisation.

702 Because of metamorphic hydrothermal processes that have affected the UHP zoisite-
703 eclogite (Berger et al., 2010), it is difficult to decipher whether the high W and Sn concentrations
704 (compared to the Primitive Mantle) reflect the pre-UHP signature of its magmatic protolith, or
705 metasomatic overprint. Fluid-rock interactions are thought to have occurred during the peak
706 metamorphic UHP stage, i.e. at c. 412 ± 5 Ma, by interaction with a fluid released by the
707 dehydrating sediments during subduction (Berger et al., 2010). However, W–Sn mineralisation

708 are younger, with late Variscan ages (333–298 Ma; Harlaux et al., 2018). Thus, even if W (and
709 Sn) enrichment in the UHP zoisite-eclogite results from fluid-rock interactions, it cannot be
710 related to the same hydrothermal event that was responsible for W–Sn major and minor
711 deposits. For the same reason, the dehydrating UHP eclogite cannot be regarded as a source
712 for W- and Sn-transporting fluids.

713 In gneiss and granites, variations in the trace element composition of micas show a clear
714 geographical correlation with the different districts (Figs. 9 and 10). Biotite and muscovite in
715 leucogranitic samples from the Vaulry W–Sn major district display the highest W and Sn
716 concentrations. Micas in the leucogranites from La Marche (Ay.MAR) and St-Sylvestre (820A
717 and IGAR744), which host minor W–Sn deposits have lower W and Sn concentrations than
718 those of Vaulry. The lowest Sn concentrations are measured in micas from the gneiss and
719 granites from the St-Yrieix Au district (St-Yrieix, Cheni, Laurières), which are located farther from
720 the Vaulry district (Fig. 10). Muscovite in sample GNEISS-SY-1 displays variable W and Sn
721 concentrations. However, biotite in gneiss GNEISS-SY-1 has W abundances within the same
722 range as biotite in leucogranite IGAR744 (St-Sylvestre) and higher W abundances than biotite in
723 leucogranite 820A (St-Sylvestre). Our data for biotite and muscovite confirm the geographical
724 correlation shown by previous results obtained for whole rocks: granites from Le Bourneix (St-
725 Yrieix) have lower Sn concentrations (5–18 ppm, up to 61 ppm in an altered granite; Calli, 1988)
726 than granites from La Marche (15–49 ppm, up to 79 ppm in aplites; Ayt Ougougdal, 1994), and
727 from Blond-Vaulry granites (212–709 ppm, Belksami, 1993), rhyolites (450–503 ppm; Belksami,
728 1993) and microgranite dykes (18–191 ppm; Ebougué, 1995). Similarly, low W abundances
729 were measured in granites from Le Bourneix (St-Yrieix) (1–7 ppm; Calli, 1988). W abundances
730 increase in microgranites (5–18 ppm; Ebougué, 1995), granites (11–12 ppm; Belksami, 1993)
731 and rhyolites (22–71 ppm; Belksami, 1993) from the Blond–Vaulry area.

732 According to Romer and Kroner (2015, 2016), Sn and W mineralisations in Palaeozoic
733 basements of the western and central Variscan belt are related to the closure of the Rheic

734 ocean. They propose that enrichment of Sn and W in granites results from three processes: 1)
735 weathering-related enrichment in Sn and W of Cadomian sediments, 2) redeposition and
736 accumulation of the sediments during tectonic processes and, 3) melting of the Sn–W-rich
737 sedimentary rocks. Wolf et al. (2018) have demonstrated the role of the melting temperature and
738 the protolith composition of migmatite for Sn partitioning between granitic melt and minerals from
739 the restite. The authors explain metal enrichment in granite bound Sn mineralisation as the
740 result of multiple extraction of Sn-depleted melt at low-temperature, while Sn remains trapped in
741 micas from restite parts during partial melting. While the low-temperature Sn-depleted melts are
742 expected to be released and lost, late granitic melts formed at high-temperature during the
743 breakdown of Sn-rich restite become enriched in Sn, contrary to Cr, V and Ni, which remain in
744 the restite (Wolf et al., 2018). In the same way, our results indicate that the geographical
745 variation in Sn and W composition, as well as in Cr, Ni, V, Ge, Zn and Cu observed between the
746 Limousin districts (Figs. 9 and 10) could be the result of variable source composition of
747 sedimentary protoliths, variable anatexis degrees or late- to post-magmatic processes.

748 Differences in W and Sn within the rocks from the St-Yrieix district may reflect local
749 enrichment of W and Sn resulting from variations in the nature of the rock or from variations in
750 the geodynamic setting: biotite from orthogneiss GNEISS-SY-1, i.e. from the Lower Gneiss Unit,
751 has higher W contents than biotite in the late-Variscan leucogranites Che10b and LAU10e.
752 Besides, biotite in the leucogranite Che10b shows higher Sn abundances than the gneiss. The
753 lithological and structural control on W and Sn concentrations within the St-Yrieix district is also
754 evidenced by previous data obtained from a series of granites, paragneiss,
755 cataclastic/brecciated quartzo-feldspathic rocks, argilo-siliceous cataclasites and auriferous
756 quartz breccias from the Cros Gallet Au mine from Le Bourneix (Calli, 1988). W abundances are
757 low in granites (1–3 ppm) and increase in the quartzo-feldspathic rocks (2–7 ppm), which
758 correspond to altered and deformed granites (Calli, 1988). Besides, Sn abundances of 5–18
759 ppm were measured in paragneiss, granites, cataclastic/brecciated quartzo-feldspathic rocks

760 and argilo-siliceous cataclasites. Higher values were obtained in altered granites and quartz
761 breccias (up to 118 ppm; Calli, 1988). Our results for the different districts argue for a local
762 source of W and Sn pre-concentrations, in agreement with Vallance et al. (2001), who has
763 considered the Blond granite as the main source for W–Sn mineralisation in the Vaulry major
764 district. Subsequent percolation of hot pseudo-metamorphic fluids in the Blond granite has
765 probably leached W and Sn, and allowed precipitation of W- and Sn-rich minerals (wolframite,
766 scheelite, cassiterite) in hydrothermal veins (Vallance et al., 2001). By analogy, the St-Sylvestre
767 and La Marche leucogranites may be the local source for the minor W–Sn occurrences hosted in
768 the two districts.

769

770 **6. Conclusions**

771 LA-ICPMS analyses of the main rock-forming minerals of granites, amphibolites and
772 serpentinites have allowed acquiring new in-situ data to trace metal sources within the different
773 lithologies of the Limousin. Serpentinites from the Limousin ophiolite-derived rocks have
774 compatible and incompatible trace element compositions typical of the Primitive Mantle, with
775 high concentrations in compatible transition metals (Sc, V, Cr, Co, Ni, Zn, Cu) and low
776 abundances of incompatible metals (Ge, Ga, Mo, Sn, Sb, W). Associated amphibolites deriving
777 from gabbros and basic dykes display trace element compositions typical of N-MORB and OFB,
778 with high abundances in compatible metals and low contents in incompatible metals. Both
779 ultrabasic and basic rocks show enrichments in As, Ag, Sn, Sb, W, Pb and Bi compared to PM
780 and OFB, which may reflect either a protolith enriched in incompatible metals, or the fingerprints
781 of high-temperature to low-temperature hydrothermal alteration that have followed the oceanic
782 crust emplacement. Besides, the UHP zoisite-eclogite is generally richer in compatible and
783 incompatible trace elements than other ultrabasic and basic rocks, with the exception of Cr, Co
784 and Ni. The origin of incompatible metals (including As, Sb, W, Pb and Bi) is uncertain, and may
785 reflect enrichment of the magmatic protolith (i.e. a Fe-rich plagioclase cumulate; Berger et al.,

786 2010) or enrichment during metasomatic processes related to the UHP metamorphic stage.
787 However, the different hydrothermal processes that have affected the Limousin ophiolite-derived
788 rocks and UHP eclogite are not contemporaneous with the formation of the mineralised veins,
789 such as Au or W–Sn deposits. Thus, despite their proximity with the Au deposits, serpentinites,
790 amphibolites and UHP eclogites do not appear to have played a role in the formation of Au
791 deposits.

792 In all the studied granites and gneiss, the metallic trace elements are mainly hosted by
793 accessory sulphides and oxides, as well as by biotite and muscovite, which contain significant
794 Sn and W contents, low amounts of Ge, As, Sb, Bi and variable amounts of Sc, Cu and Ga.
795 Quartz and feldspars only show minor amounts of Sc, Ga, Ge, Sn and Pb. Variations in W and
796 Sn abundances in micas allow tracing the pre-mineralisation W and Sn concentrations in the
797 different granite-hosted ore districts. Leucogranites from the Vaulry major W–Sn ore deposit
798 display the highest W and Sn concentrations. W and Sn abundances decrease in the
799 leucogranites from La Marche and St-Sylvestre U districts, which host minor W–Sn deposits.
800 The smallest concentrations are generally measured in micas from the orthogneiss and
801 leucogranites from the St-Yrieix Au district (St-Yrieix, Cheni, Laurières). In agreement with
802 previous studies, our results argue for a local source of W and Sn and highlight a geographical
803 correlation between W and Sn concentrations in the mineral assemblage of granites and W–Sn
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805

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816

817 **8. Electronic Appendix:** Supplementary Tables are associated with this article.

818

819 **9. References**

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1151

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1157 Chalier (1994), Scaillet et al. (1996a, 1996b), Marignac and Cuney (1999), Alexandrov (2000),
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1159 Harlaux et al. (2018).

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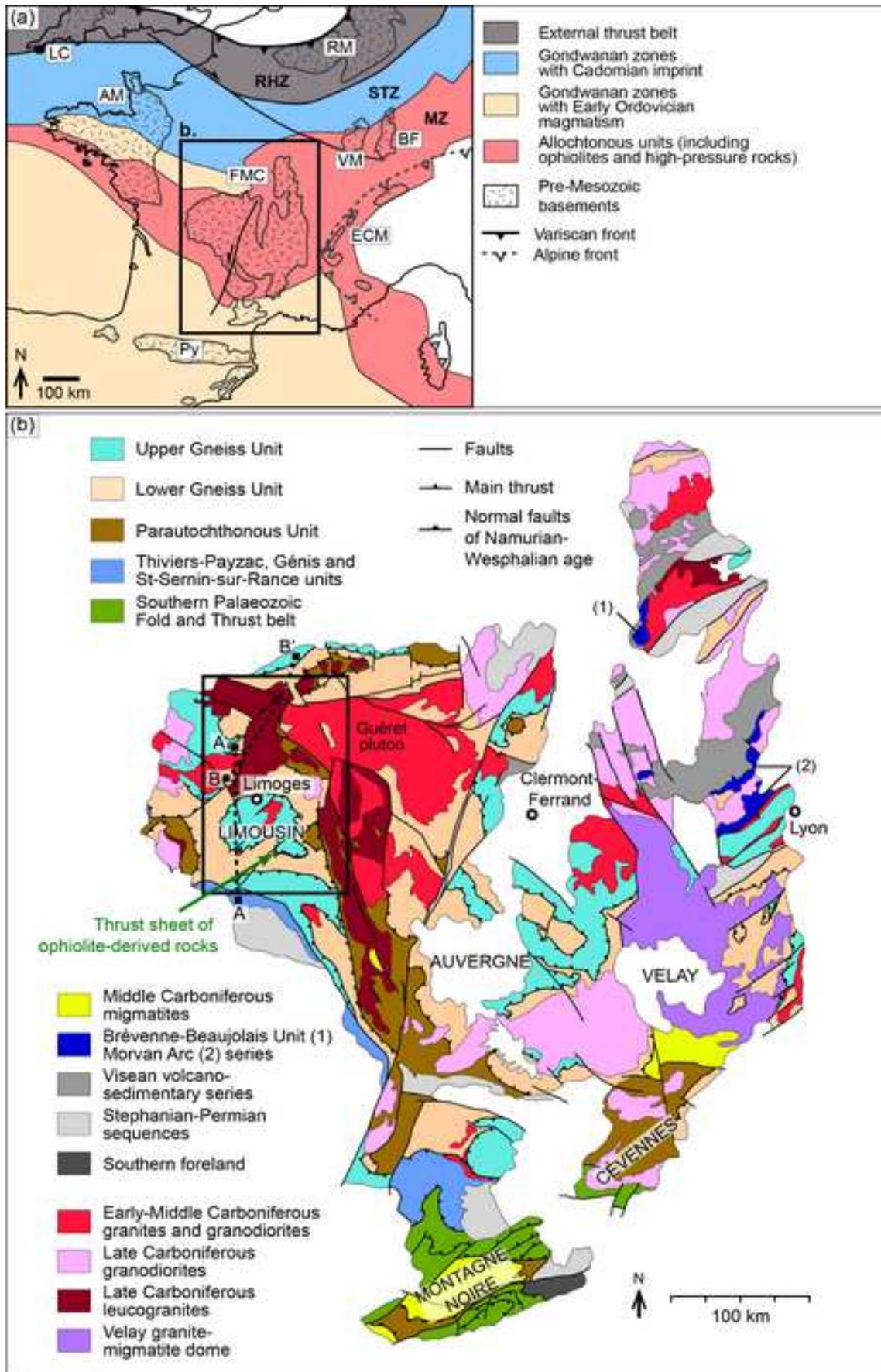


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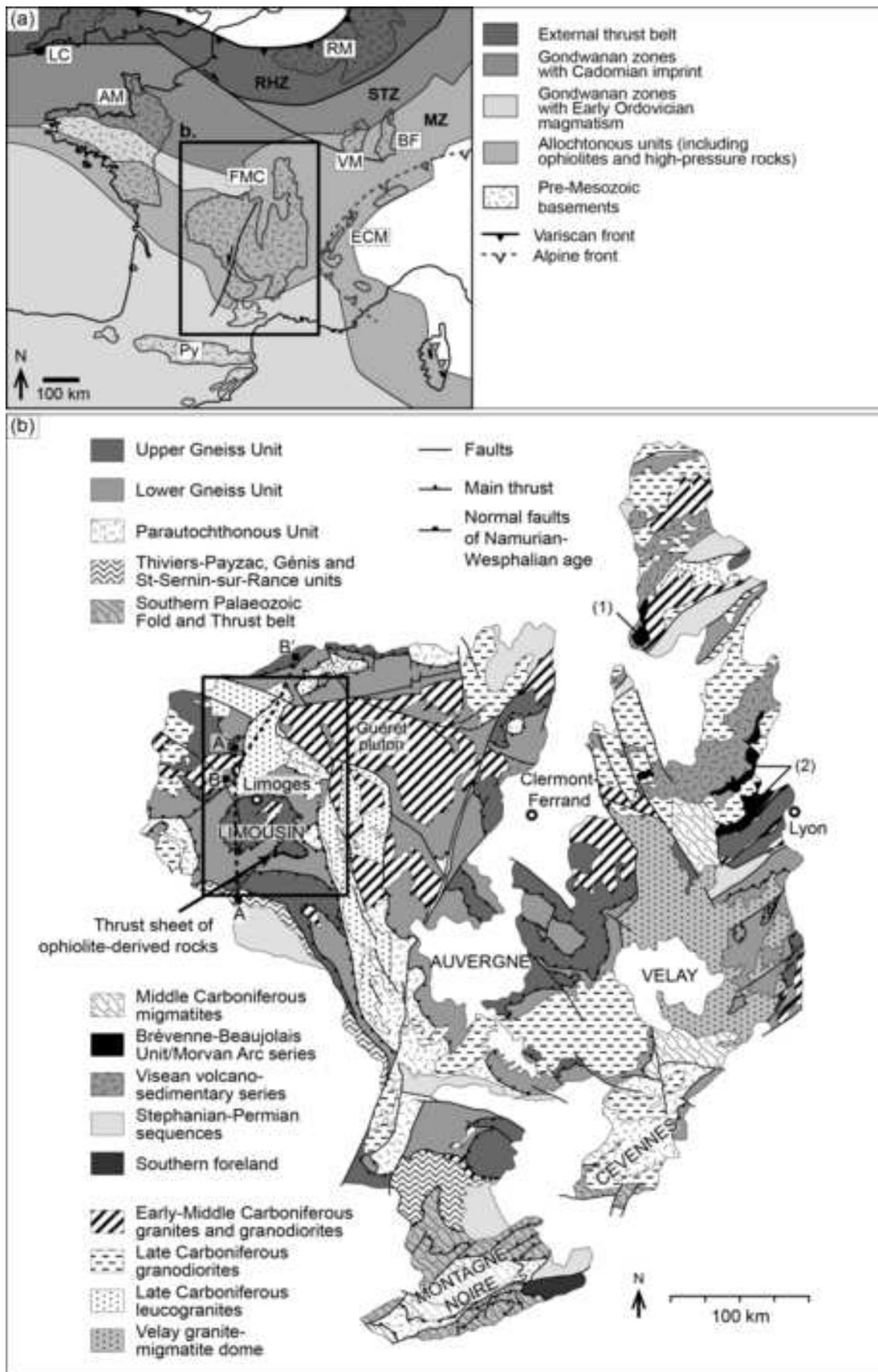


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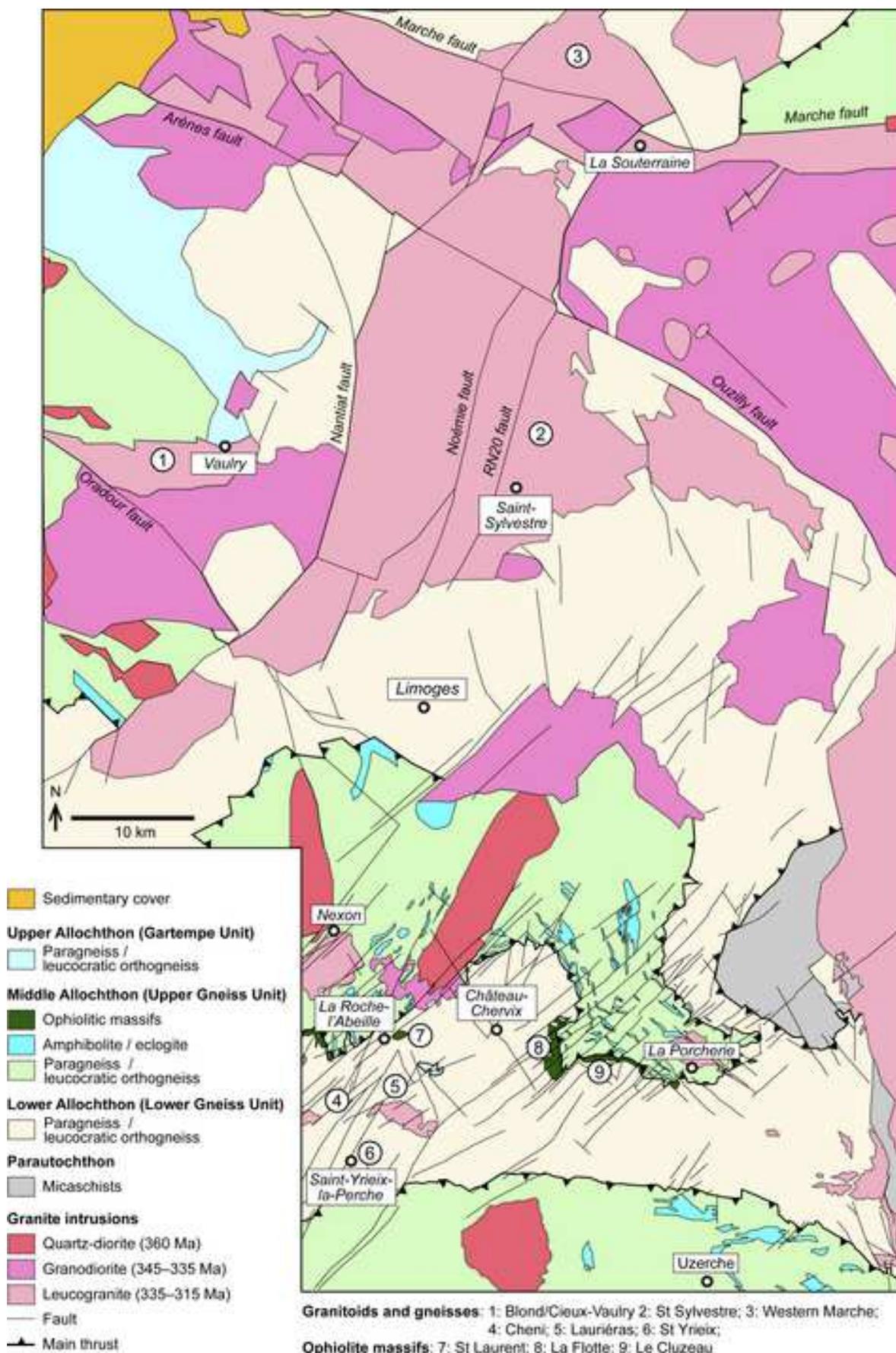


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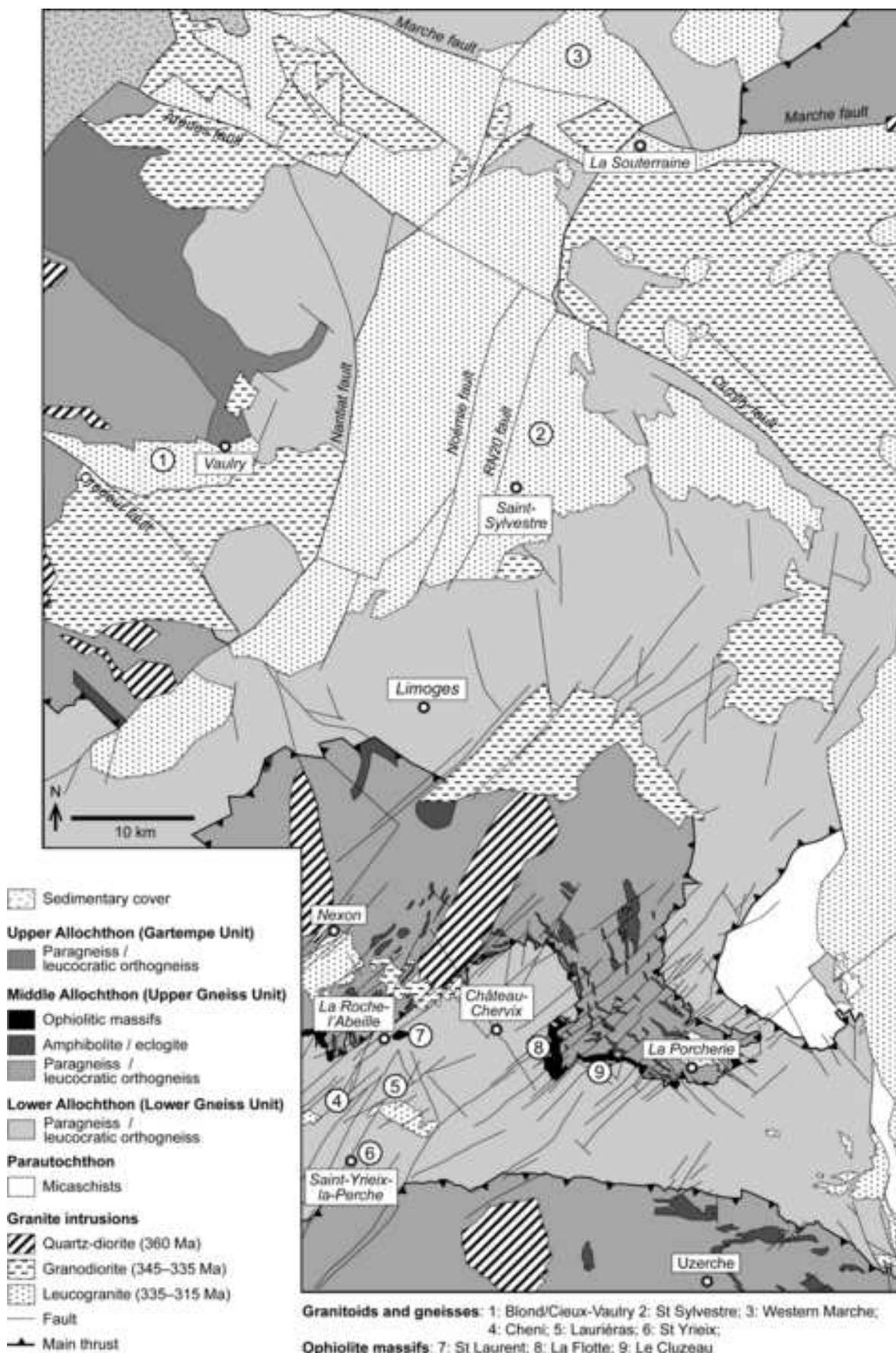


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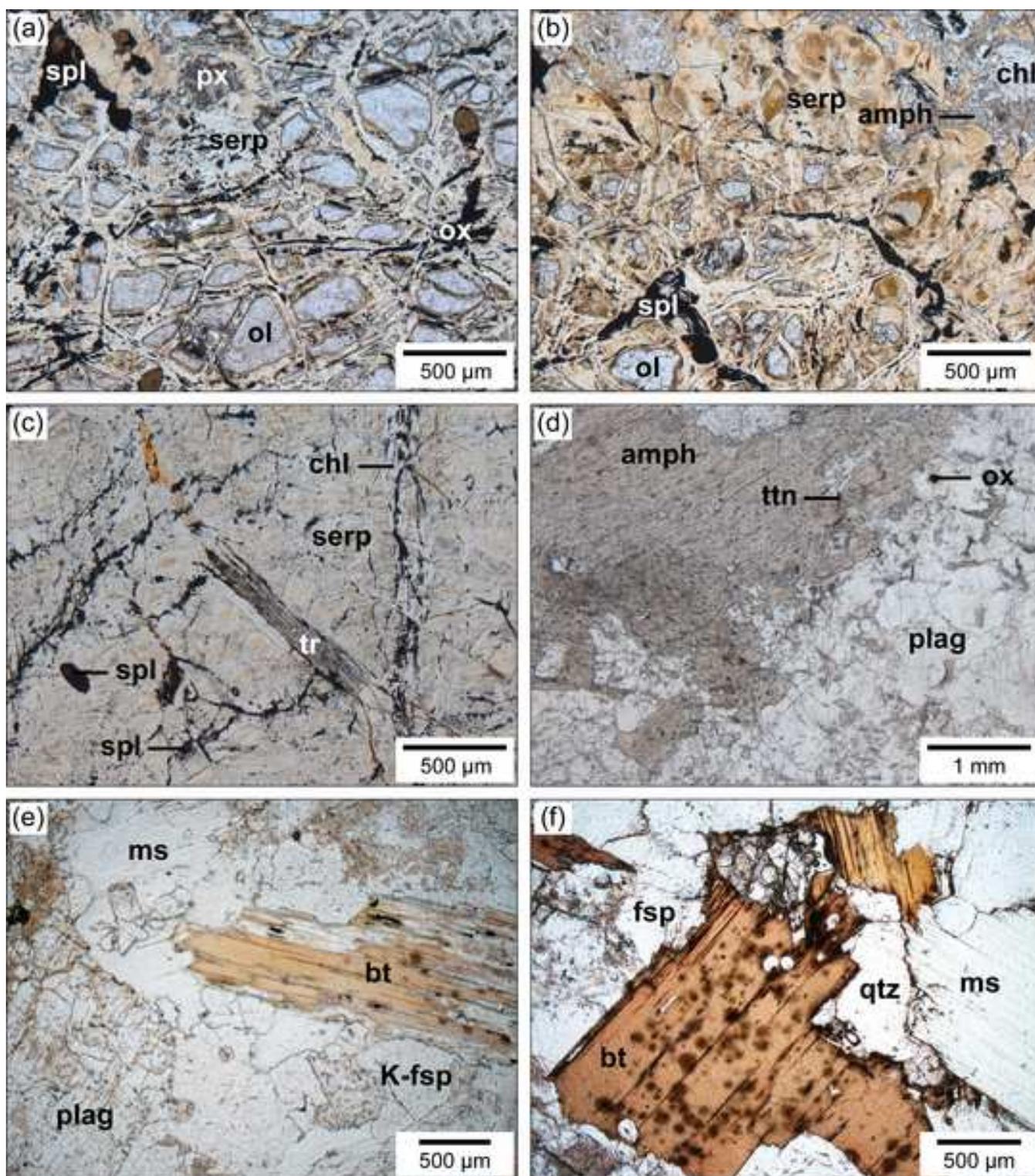


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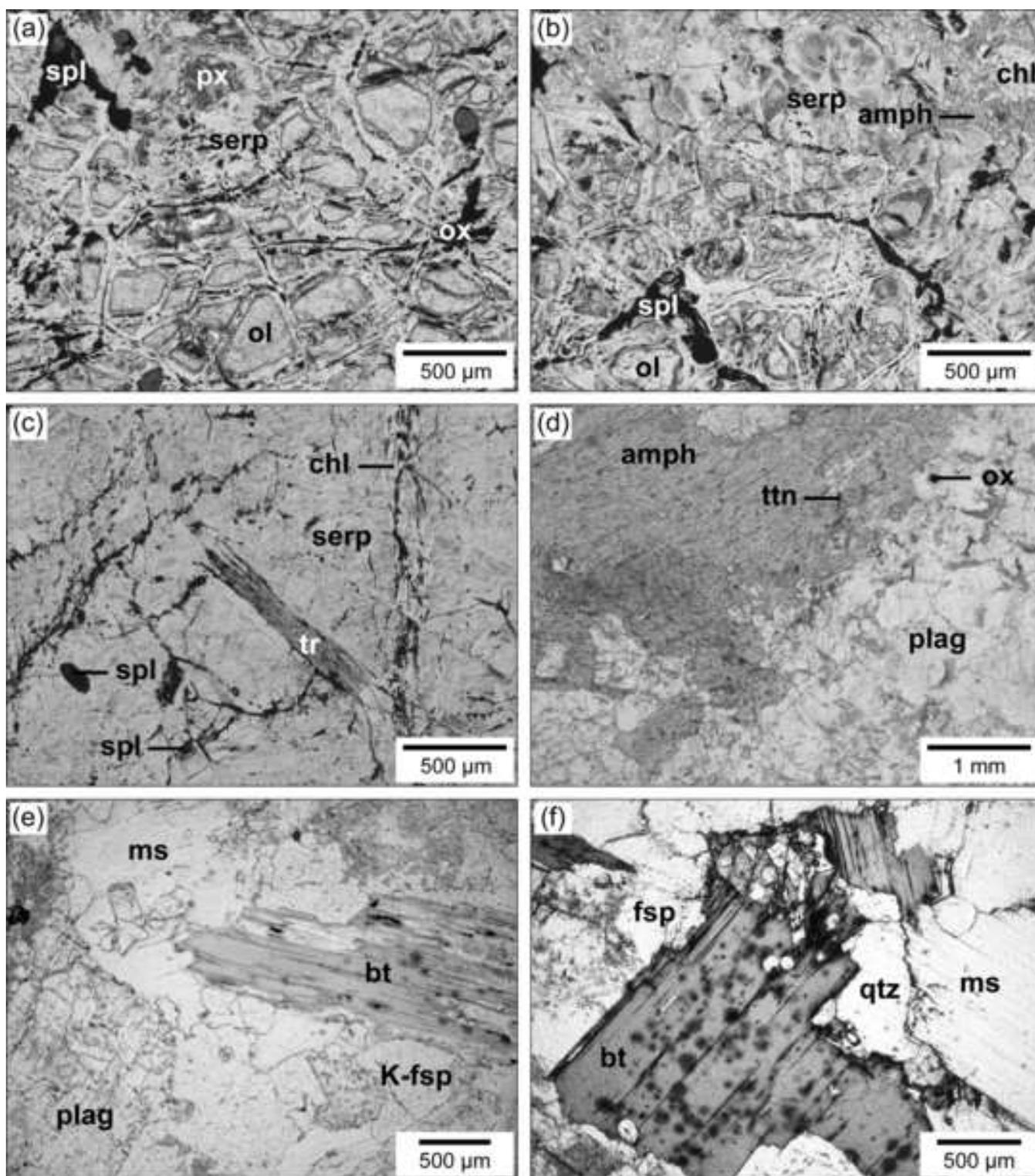


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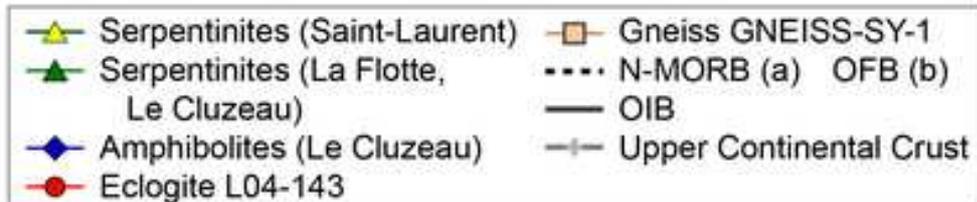
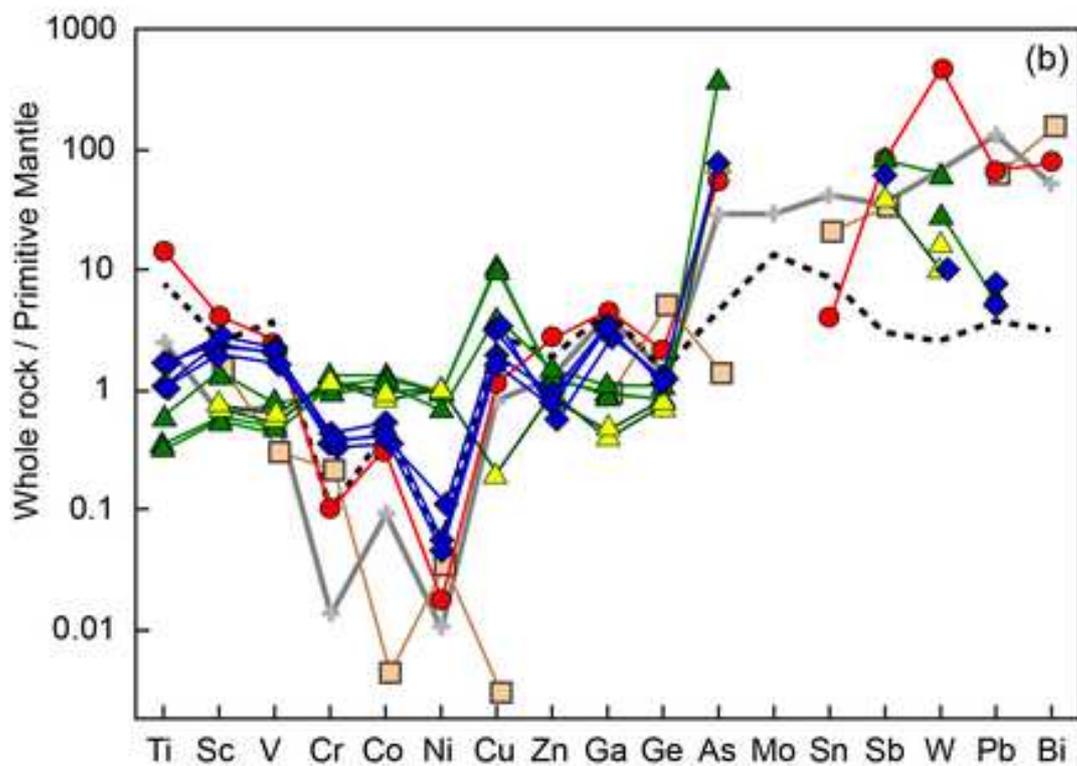
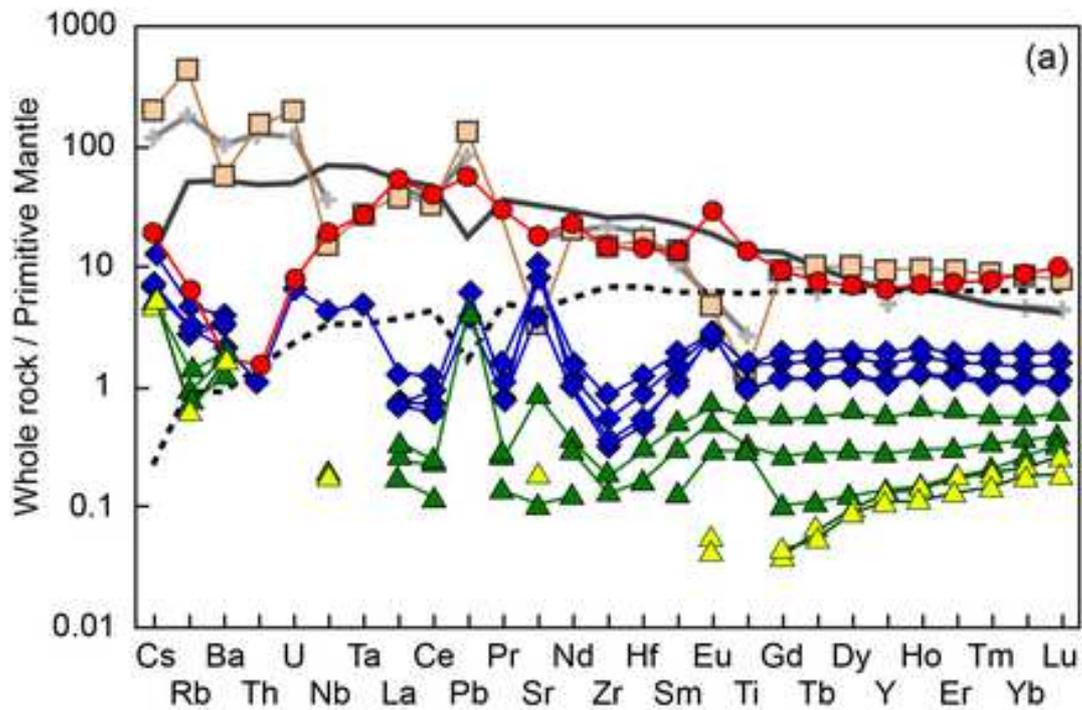


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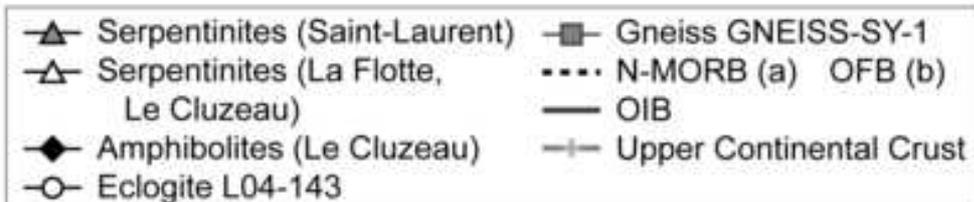
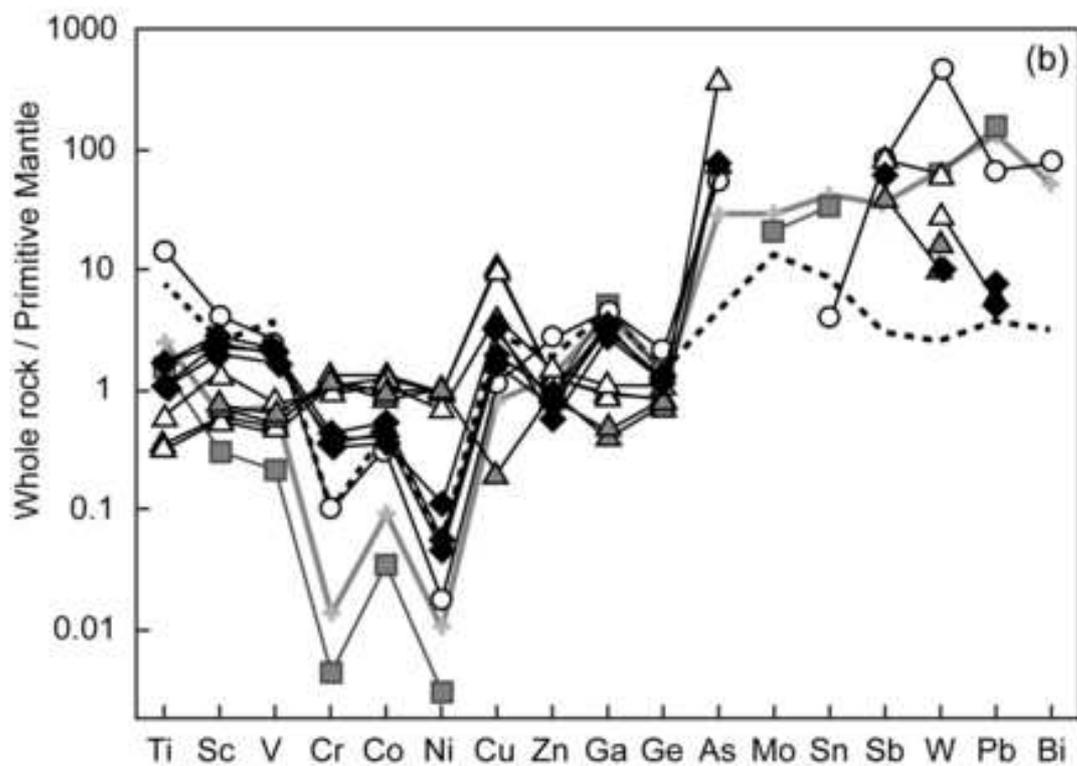
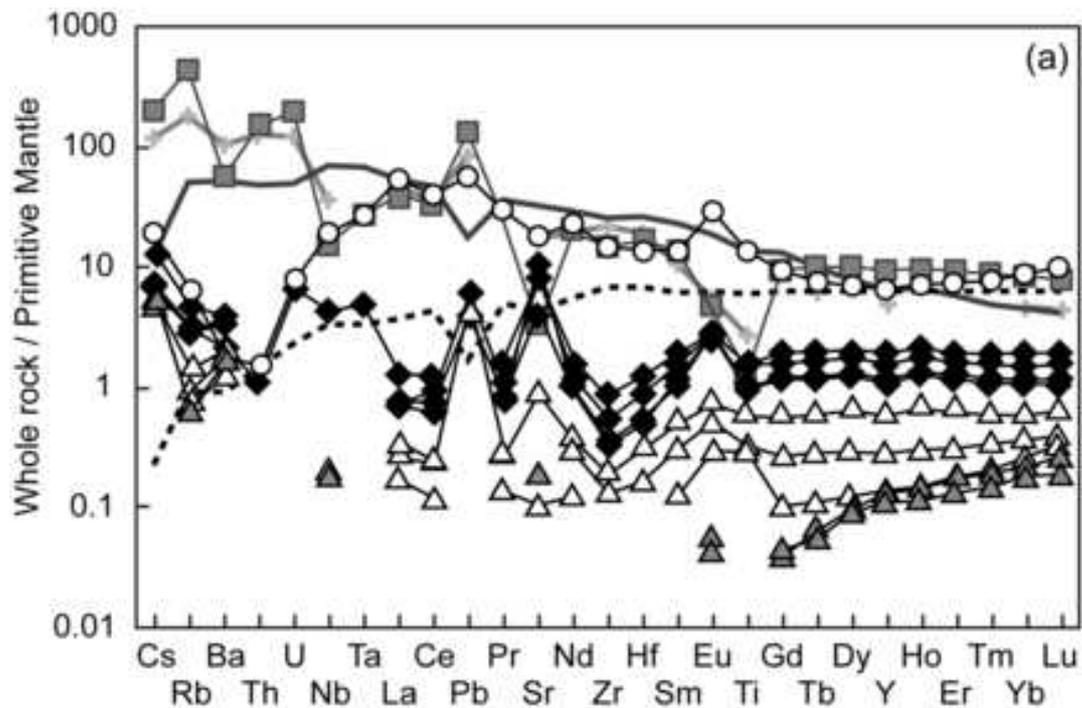


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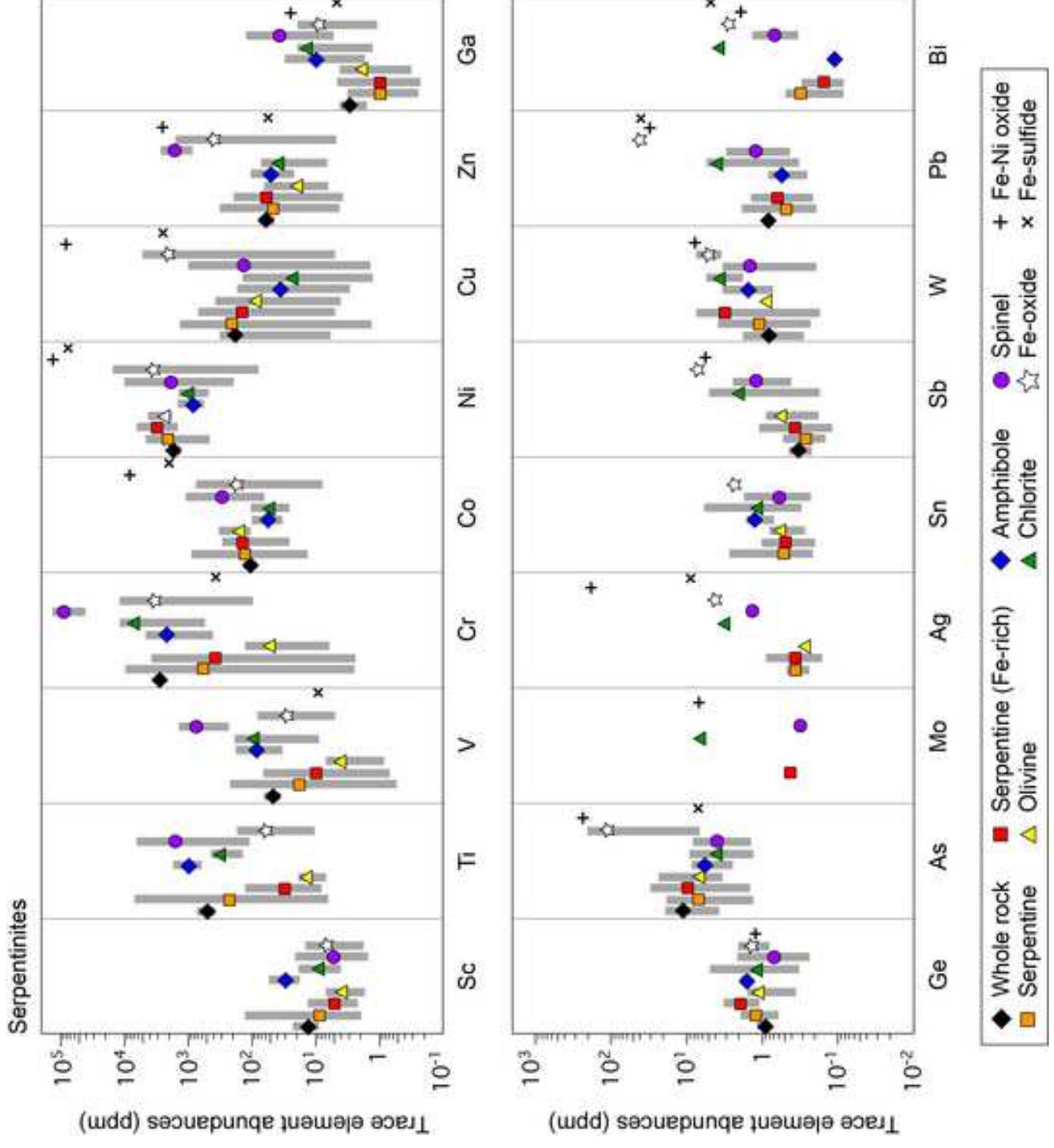


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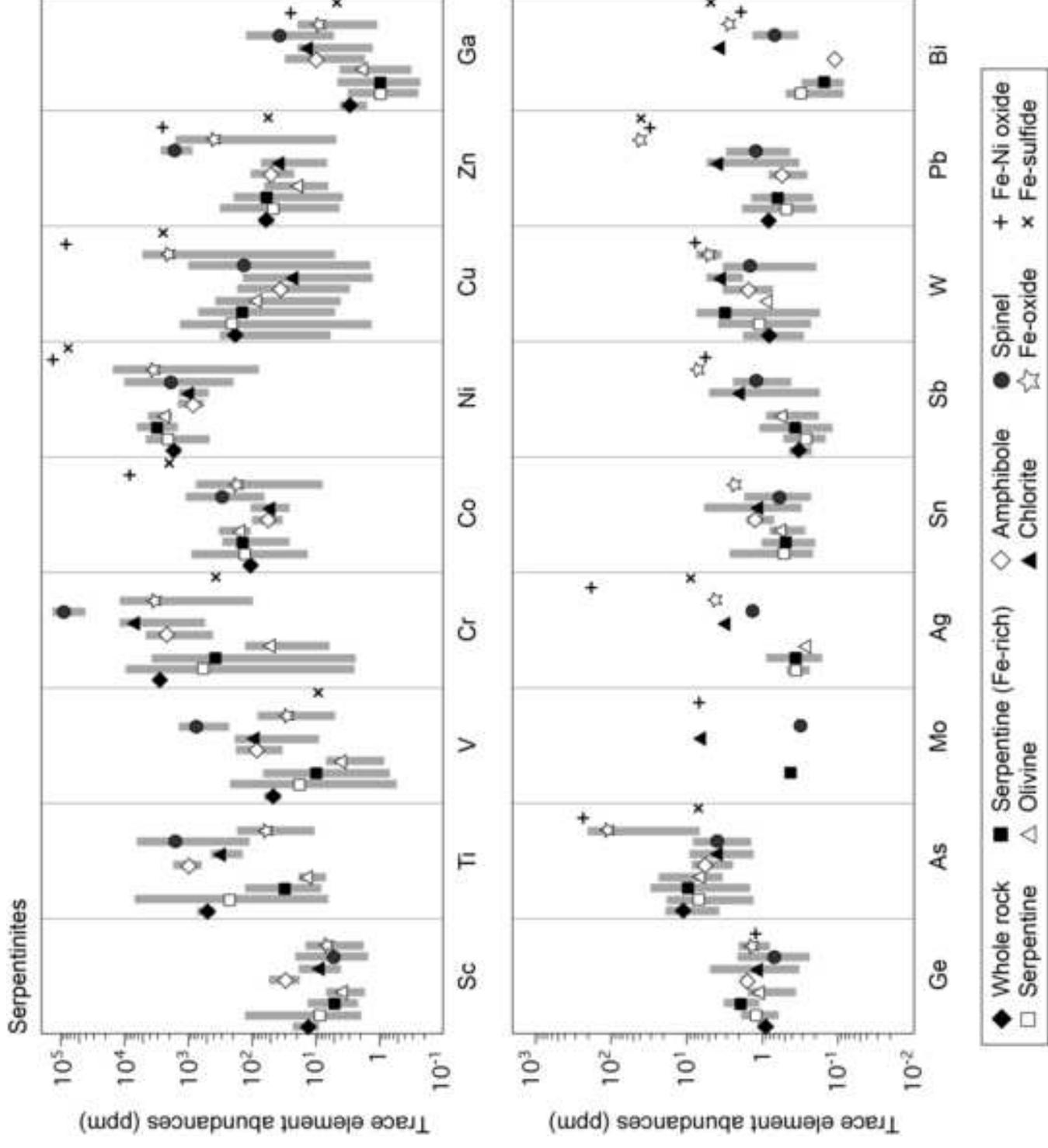


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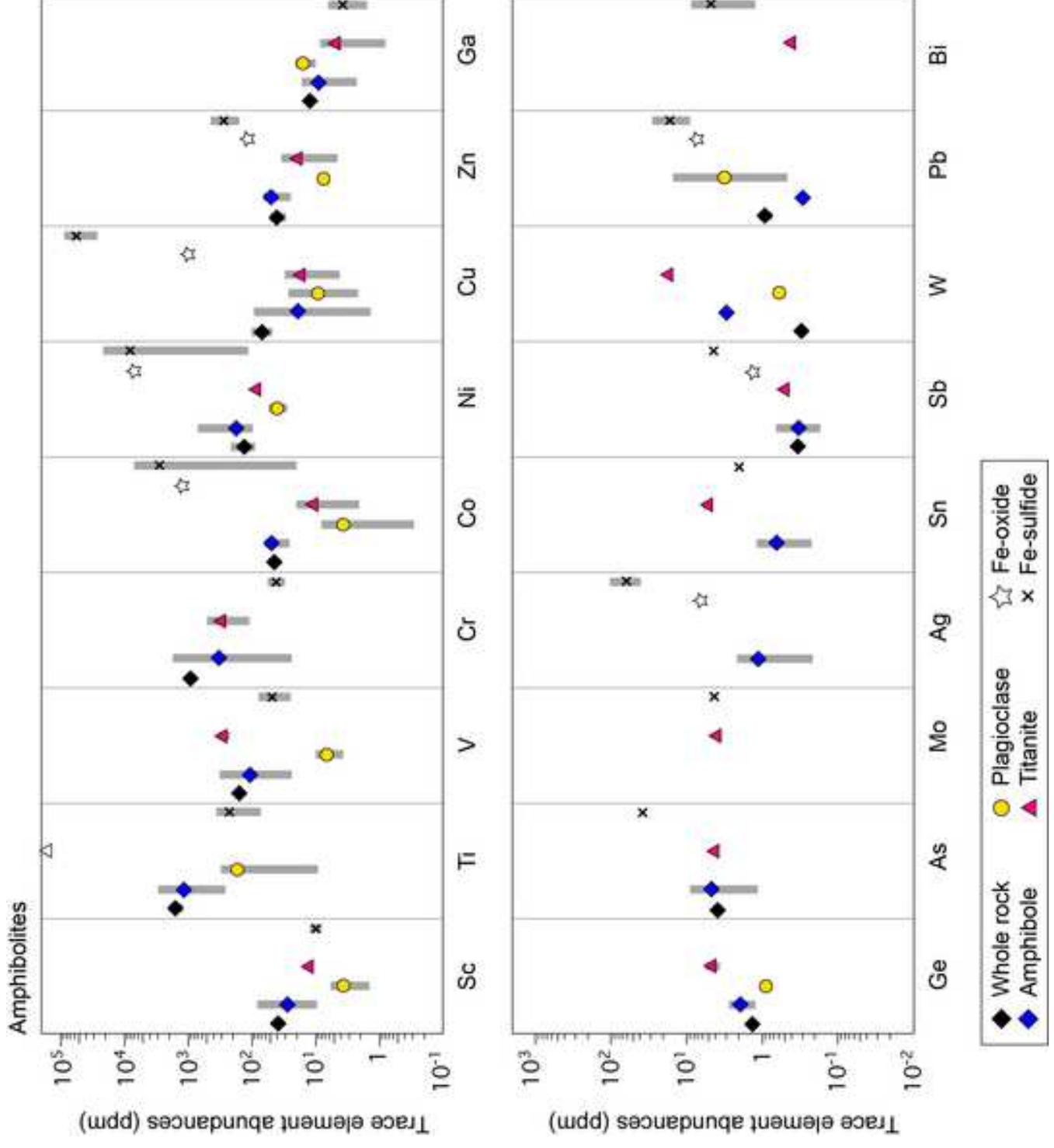


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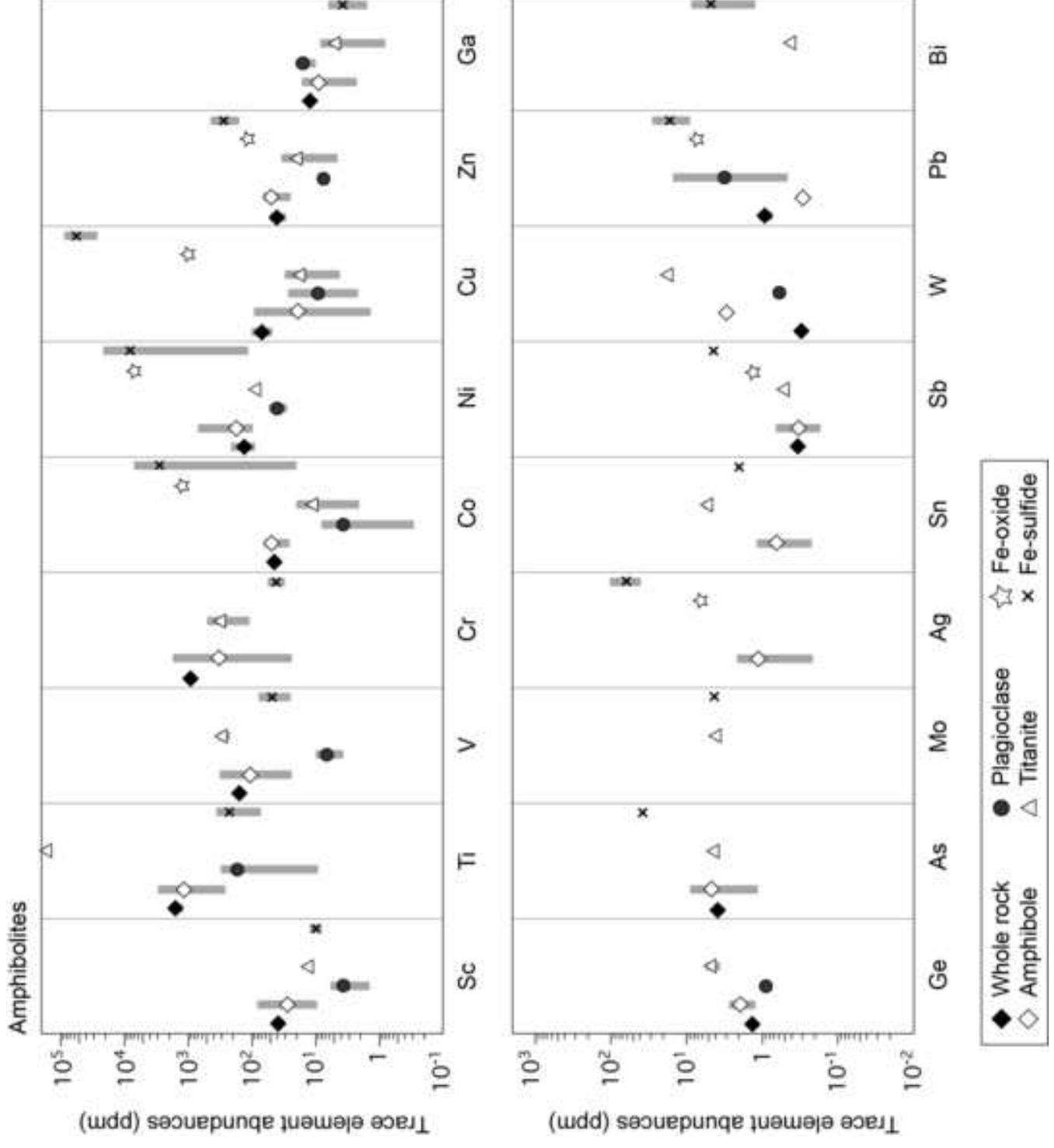
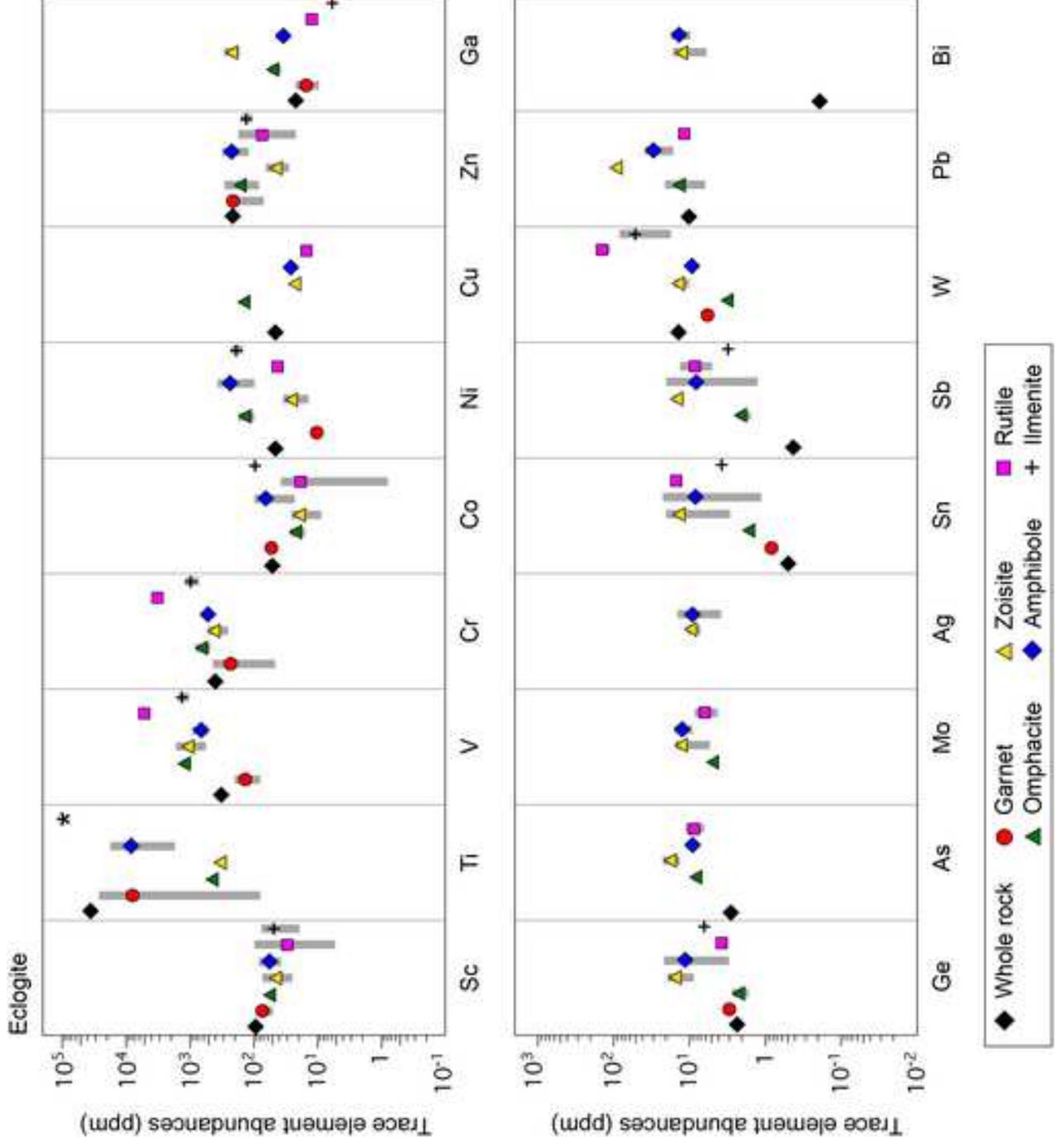
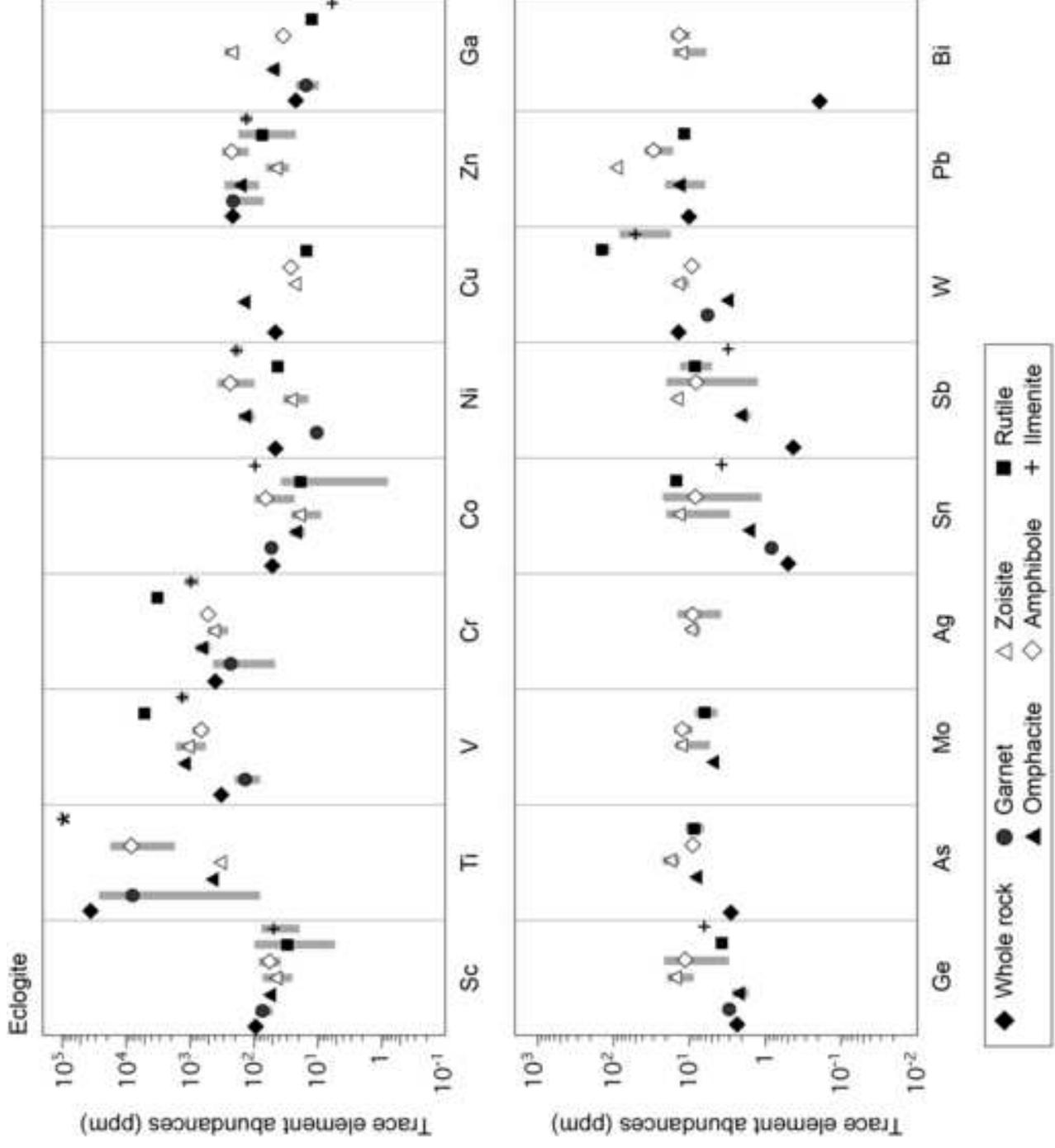


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* Ti is present as major element (> 4 × 10⁵ ppm) in rutile and ilmenite.

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* Ti is present as major element (> 4 × 10⁵ ppm) in rutile and ilmenite.

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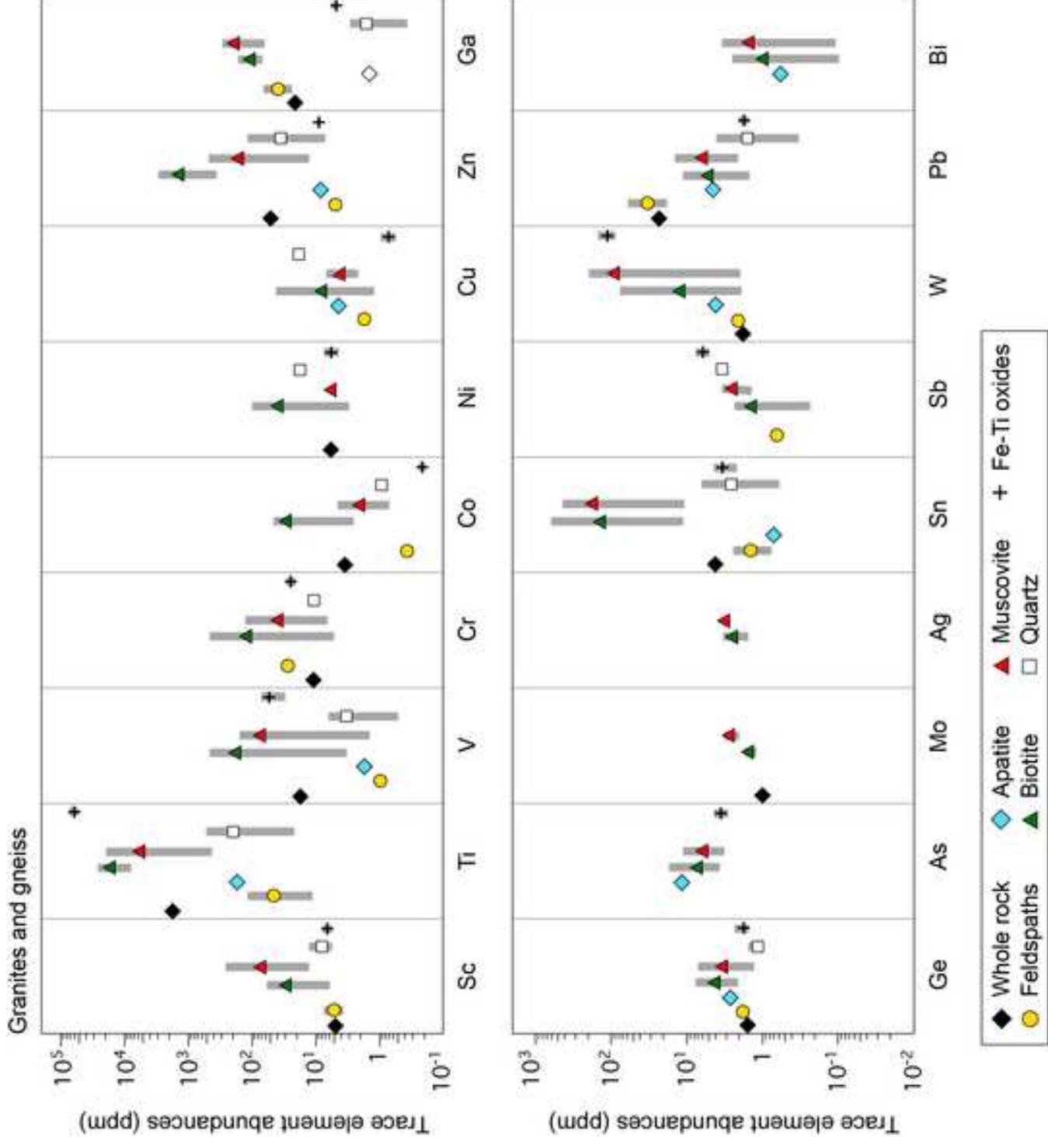


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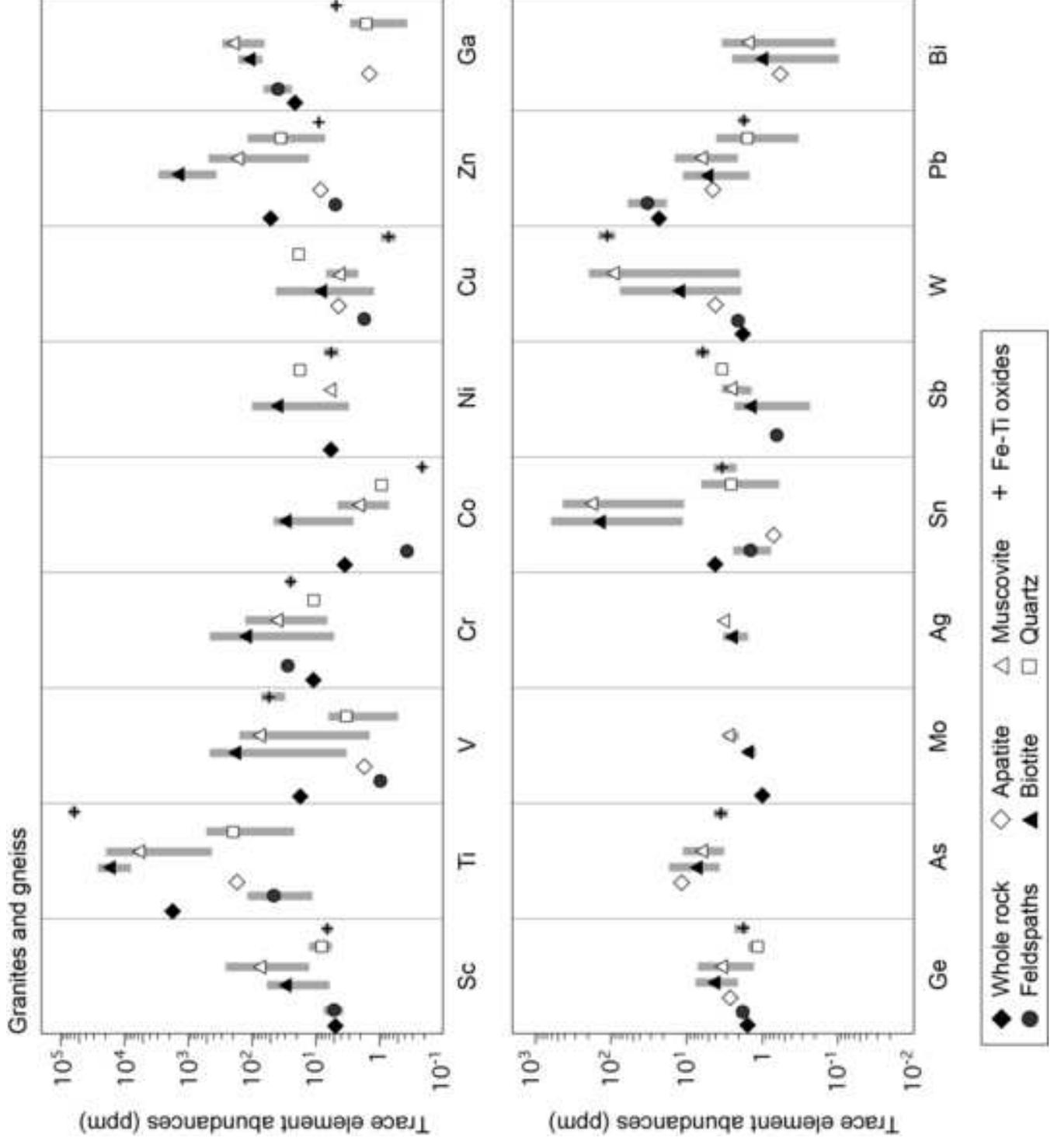


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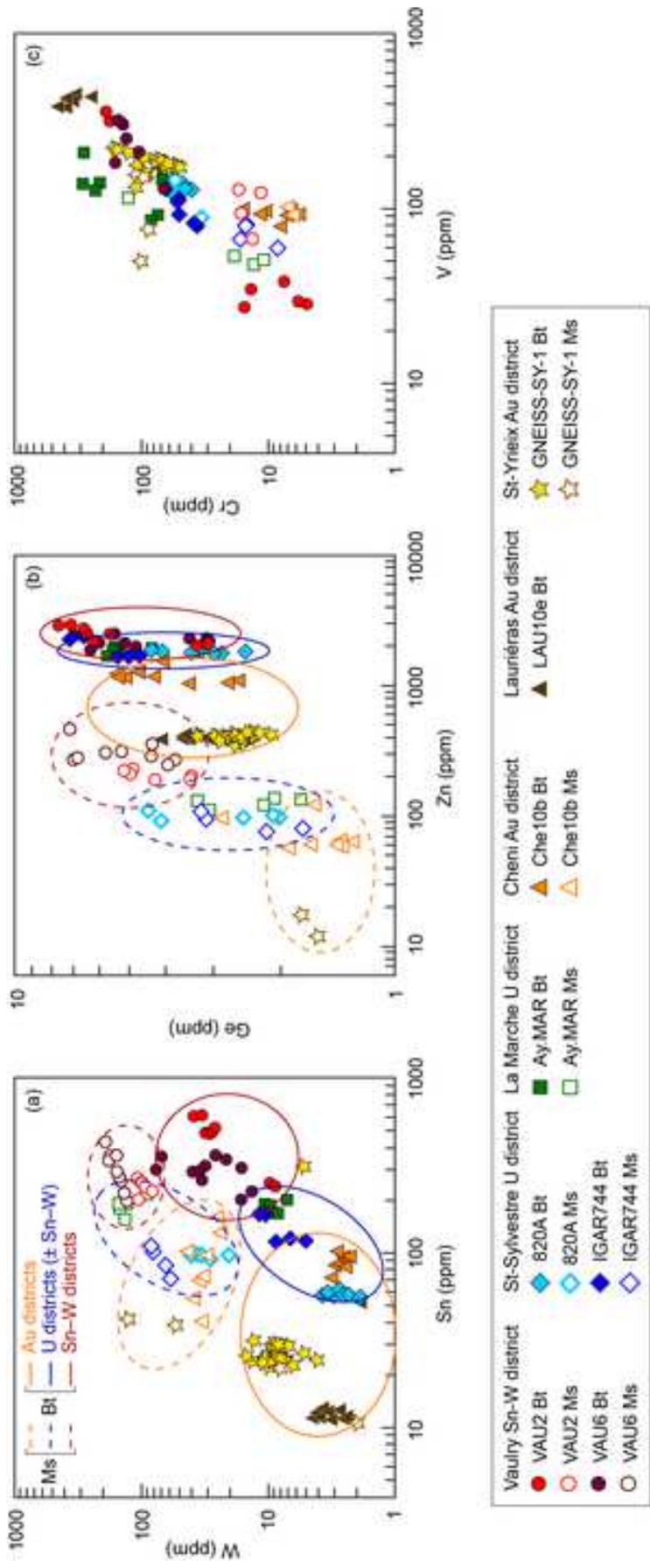


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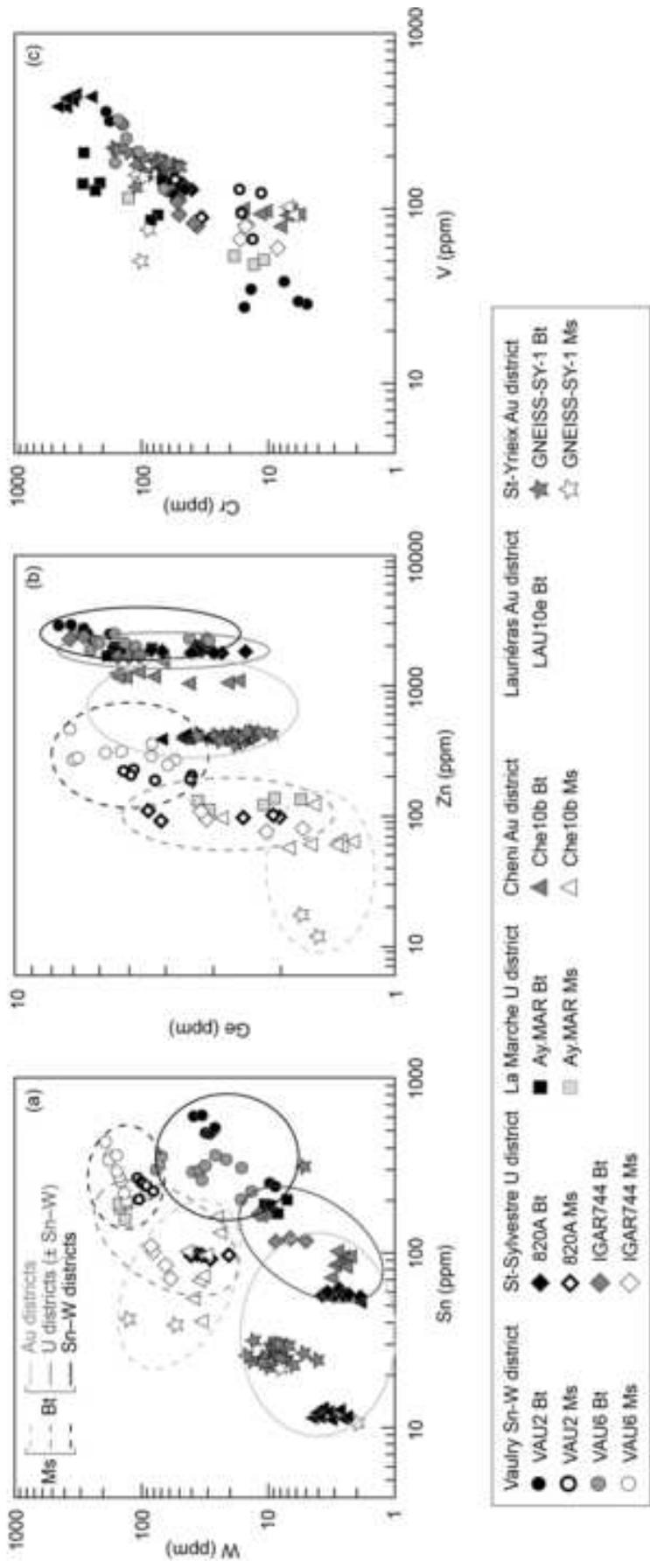


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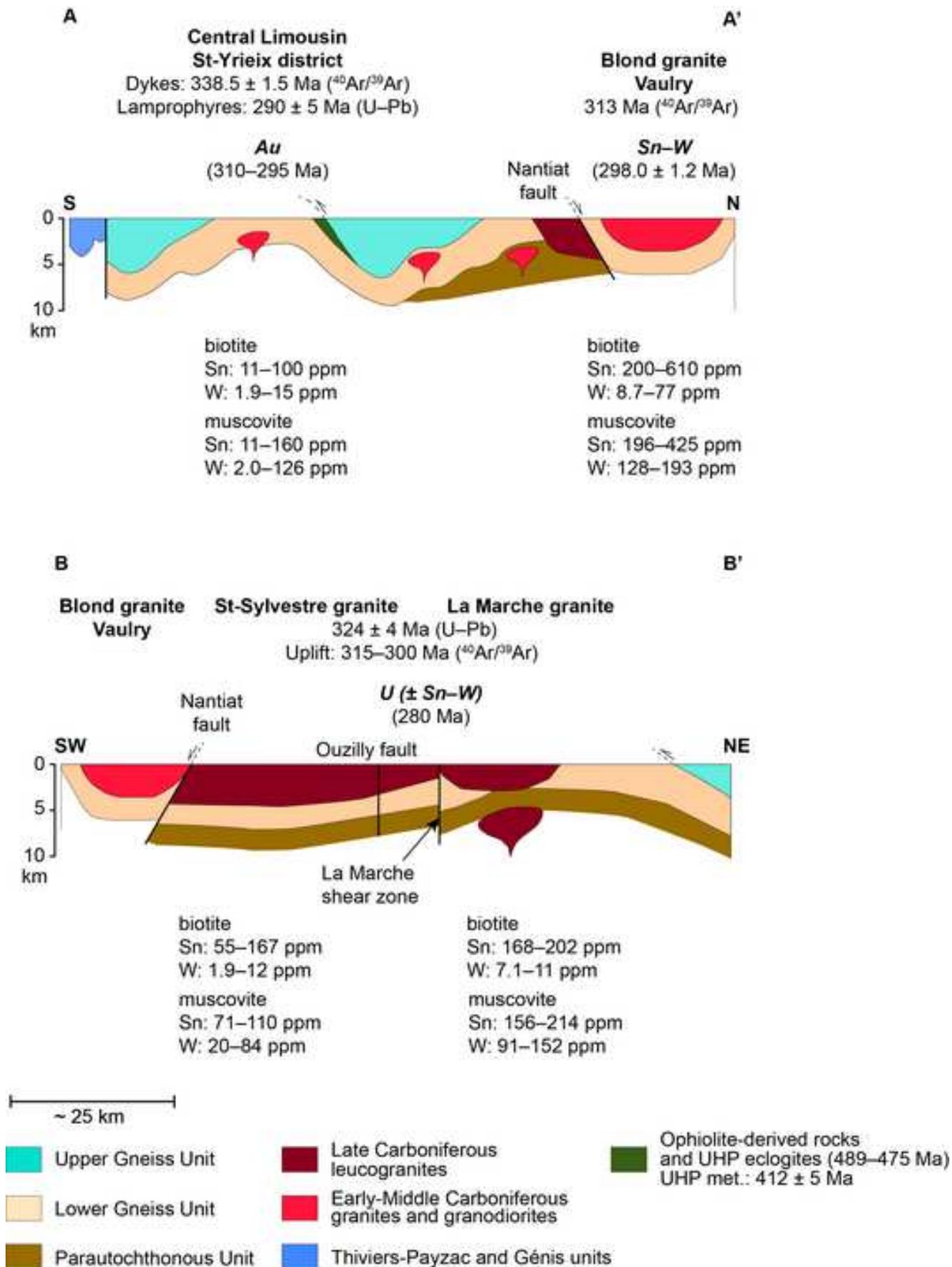


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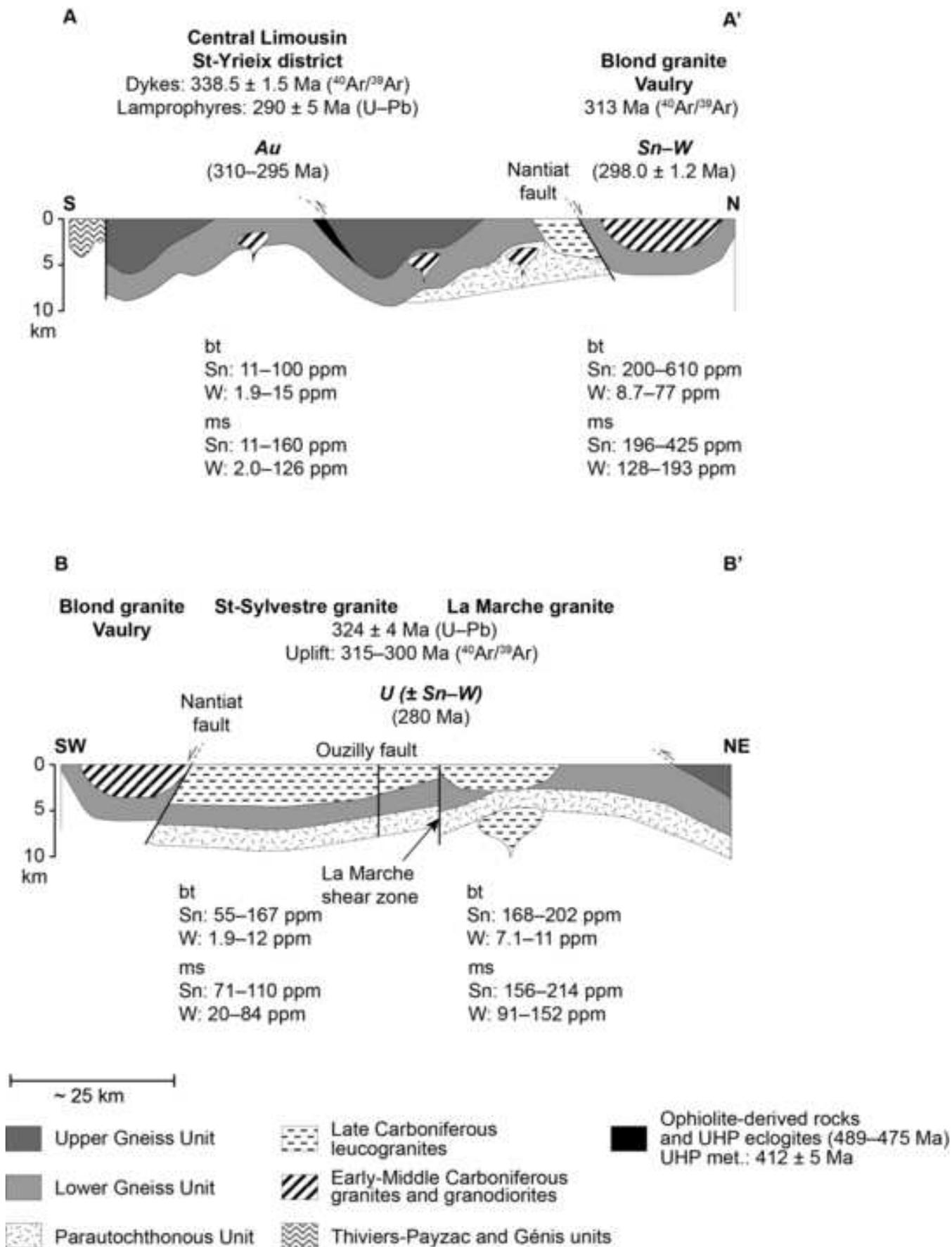


Table 1

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Table 1. Description and mineral assemblage of the studied samples

Sample	Rock type	Locality*	Mineralogical assemblage**	Ref.***
LAU-1	Serpentinite	Saint-Laurent	Serp-chl-spl-± ol-± cpx	[1]
LAU-2	Serpentinite	Saint-Laurent	Serp-chl-spl-Fe ox	[1]
CLUZ-6	Serpentinite	Le Cluzeau	Serp-amph-chl-spl-Fe ox- Fe sulph	[1]
FLOT-1	Serpentinite	La Flotte	Serp-amph-chl-± ol-Fe ox	[1]
FLOT-2a	Serpentinite	La Flotte	Serp-spl-ol-Fe ox	[1]
FLOT-2b	Serpentinite	La Flotte	Serp-chl-Fe ox	[1]
CLUZ-1	Amphibolite (metamorphosed dyke)	Le Cluzeau	Amph-plag-chl-Fe ox	[1]
CLUZ-1a	Amphibolite (metamorphosed dyke)	Le Cluzeau	Amph-plag-chl-Fe ox	[1]
CLUZ-4	Amphibolite (metagabbro)	Le Cluzeau	Amph-plag-chl-ttn-Fe ox- Fe sulph	[1]
CLUZ-5	Amphibolite (metagabbro)	Le Cluzeau	Amph-plag-chl-Fe ox	[1]
L04-143	Eclogite	500 m E of Saint-Laurent	Grt-omph-zo-amph-rt-ilim	[2]
VAU2	Leucogranite	Vaulry	Fsp-plag-qtz-bt-ms-zrc	[3]
VAU6	Leucogranite	Vaulry	Fsp-plag-qtz-bt-ms-zrc	[3]
GENEISS-SY-1	Orthogneiss	St-Yrieix	Fsp (mi)-plag (ab)-qtz-bt-ms-ap-zrc	[1]
Lau-10e	Leucogranite	Laurieras	Fsp (mi)-plag (ab)-qtz-bt-py-Fe-Ti ox-zrc	[4]
Che10b	Leucogranite	Cheni	Fsp (mi)-plag (ab)-qtz-bt-ms-chl-zrc	[5]
IGAR744	Leucogranite	St-Sylvestre	Fsp-plag-qtz-bt-ms-zrc-mz	[6]
820A	Leucogranite	St-Sylvestre	Fsp-plag-qtz-bt-ms-ap-zrc-mz	[6]
Ay.MAR	Leucogranite	La Marche	Fsp-plag-qtz-bt-ms-zrc-mz	[7]

* see Figure 1

** The mineral abbreviations are from Kretz, R. (1983); Fe sulph: Fe-sulphides (pyrite)

*** [1] This study; [2] Berger et al., 2010; [3] Vallance et al., 2001; [4] Essarraj et al., 2004; [5] Vallance et al., 2004; [6] El Jarray, 1993; [7] Ayt Ougougdaï, 1993

Table 2
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Table 2. Major and trace element composition of the samples from the Limousin ophiolite.

	Serpentinites			Amphibolites					Orthogneiss			Eclogite		OFB Mean (\pm 2sd) [3]
	LAU1	LAU2	CLUZ6	FLOT2a	FLOT2b	CLUZ1	CLUZ1a	CLUZ4	CLUZ5	Orthogneiss GNEISS-SY-1	Eclogite L04-143	[1]	[2]	
SiO ₂	38.92	38.62	39.13	35.53	35.39	48.06	46.82	49.23	48.26	74.44	39.50		40.00	
TiO ₂	bdl	bdl	0.12	0.06	0.07	0.21	0.22	0.34	0.33	0.30	2.88		2.90	
Al ₂ O ₃	1.89	1.69	4.19	3.30	4.30	18.41	14.83	16.33	16.13	13.10	20.46		20.48	
Fe ₂ O ₃	4.76	5.31	7.30	6.34	6.96	0.85	1.19	0.59	1.02	0.20	0.90		-	
FeO	2.55	1.95	3.61	4.09	3.03	5.18	6.65	4.22	5.99	1.95	16.77		17.57	
MnO	0.10	0.07	0.13	0.15	0.13	0.11	0.15	0.09	0.14	0.03	0.30		0.30	
MgO	36.33	36.06	30.41	37.52	34.74	10.24	13.54	11.11	10.94	0.49	8.32		8.47	
CaO	0.58	0.03	2.86	0.43	0.07	12.43	11.33	13.85	11.74	0.91	8.86		9.06	
Na ₂ O	0.03	bdl	0.27	bdl	bdl	1.84	1.54	1.42	2.07	2.65	0.72		0.74	
K ₂ O	bdl	bdl	0.02	bdl	bdl	0.13	0.11	0.12	0.17	0.16	0.06		0.05	
P ₂ O ₅	0.05	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.07	bdl		0.07	
LOI	13.60	14.88	11.28	11.62	14.21	1.61	2.39	1.68	2.14	0.62	-1.07		-0.86	
Total	98.81	98.61	99.32	99.04	98.91	99.07	98.75	98.97	98.93	99.83	98.77		100.74	
Ti	bdl	bdl	731	420	378	1247	1289	2014	1978	1774	17285		17381	
Sc	12	11	23	9.6	8.9	32	39	46	38	4.7	66		-	
V	51	40	64	39	36	137	161	172	167	17	213		309 (132)	
Cr	2793	3042	2436	3427	2837	958	1120	859	906	11	284		280	
Co	96	87	114	137	129	42	55	36	47	3.5	37		36	
Ni	1826	1992	1326	1883	1850	90	108	216	122	5.6	35		35	
Cu	bdl	6.0	119	326	309	48	59	101	78	bdl	34		86 (53)	
Zn	44	51	81	66	67	39	51	30	43	50	145		104 (39)	
Ga	1.8	1.6	4.3	3.8	3.7	14	12	12	12	20	17		18 (3)	
Ge	0.88	0.82	1.2	0.89	0.91	1.2	1.4	1.4	1.4	1.6	2.3		1.5 (0.2)	
As	bdl	3.7	bdl	bdl	19	bdl	bdl	bdl	3.8	bdl	1.02		-	
Mo	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl		-	
Sn	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	4.2	bdl		-	
Sb	bdl	0.22	bdl	bdl	0.45	bdl	bdl	bdl	0.33	bdl	0.51		-	
W	0.50	0.29	0.83	bdl	1.8	bdl	bdl	bdl	bdl	1.8	0.43		-	
Pb	bdl	bdl	0.84	bdl	bdl	bdl	0.72	bdl	1.1	23	14		-	
Bi	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	10		-	
Cs	0.14	0.17	0.17	bdl	bdl	0.21	bdl	0.21	0.43	6.2	0.59		0.60	
Rb	bdl	0.40	0.88	0.45	0.59	1.9	1.7	1.9	2.9	267	4.0		3.7	
Ba	9.4	13	14	8.7	12	23	23	14	27	381	12		12	
Th	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.10	bdl	12	0.13		0.18	
U	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.13	bdl	4.0	0.17		0.19	
Nb	0.12	bdl	0.13	bdl	bdl	bdl	bdl	3.1	bdl	10	13		14	
Ta	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.20	bdl	1.1	1.1		1.1	
La	bdl	bdl	0.18	0.11	0.24	0.47	0.51	0.92	0.50	26	36		38	
Ce	bdl	bdl	0.41	0.21	0.44	1.1	1.3	2.2	1.6	57	68		68	
Pr	bdl	bdl	0.07	0.04	0.08	0.21	0.23	0.40	0.27	6.8	7.7		7.6	
Sr	3.7	bdl	18	2.2	bdl	212	90	153	167	65	371		355	

Nd	bdl	0.47	0.17	0.38	1.2	1.3	2.1	1.6	25	30	29
Zr	bdl	2.1	bdl	1.5	3.9	3.6	9.7	6.1	158	157	150
Hf	bdl	0.09	bdl	0.05	0.16	0.15	0.38	0.27	4.8	4.2	3.8
Sm	bdl	0.22	0.05	0.13	0.47	0.51	0.83	0.67	5.8	5.8	5.6
Eu	0.009	0.12	0.05	0.09	0.45	0.45	0.45	0.46	0.70	4.7	4.6
Gd	0.02	0.33	0.06	0.16	0.66	0.72	1.1	0.93	5.6	5.4	5.2
Tb	0.007	0.06	0.01	0.03	0.12	0.13	0.22	0.18	1.0	0.79	0.78
Dy	0.07	0.47	0.09	0.21	0.88	0.96	1.5	1.3	7.0	5.0	4.8
Y	0.60	2.6	0.61	1.3	4.9	5.3	8.7	7.0	42	29	28
Ho	0.02	0.11	0.02	0.05	0.20	0.22	0.34	0.29	1.5	1.2	1.0
Er	0.07	0.30	0.08	0.15	0.54	0.60	0.93	0.77	4.3	3.5	3.2
Tm	0.01	0.04	0.02	0.03	0.08	0.09	0.14	0.11	0.62	0.56	0.54
Yb	0.11	0.29	0.13	0.17	0.52	0.58	0.92	0.72	4.0	4.3	4.1
Lu	0.02	0.05	0.03	0.03	0.08	0.09	0.14	0.11	0.56	0.74	0.71

Major element data are from El Korh et al. (in press)

[1] This study; [2] Berger et al. (2010); [3] Jenner and O'Neill (2002)

2sd: 2σ standard deviation

bd!: below detection limit

Table 3

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Table 3. Trace element abundances in the mineral assemblage of serpentinites

	serp LAU2	Fe-serp LAU2	Fe-serp FLOT1	serp FLOT1	serp FLOT2a	Fe-serp FLOT2b	serp FLOT2b	serp FLOT2b	Fe-serp CLUZ6	serp CLUZ6	serp CLUZ6	ol FLOT1	ol FLOT2a	spl LAU2	spl FLOT2a
Sc	3.0-15	4.4-13	2.2-4.3	1.9-6.5	2.3-133	2.9-4.0	2.7-4.0	2.2-5.9	2.2-5.1	2.2-5.1	3.4-4.7	1.7-6.9	1.5-21	4.53-14	
Ti	8.0-27	14-133	12-38	27-107	9.1-7360	21-62	47-98	8.3-42	6.5-79	6.5-79	7.8-19	6.9-8.0	1709-2103	112-410	
V	1.6-3.7	3.4-66	0.69-1.0	0.58-1.8	9.7-221	2.4-4.0	2.6-3.9	2.2-6.9	0.53-7.4	0.53-7.4	bdl	0.85-6.9	1110-1459	388-834	
Cr	9.1-100	16-3950	3.9-8.6	3.6	3.6-10209	2.5-8.6	2.5-6.9	2.3-22	5.7-24	5.7-24	8.1	6.01-133	86141-87133	76700-144817	
Co	13-83	64-299	26-259	107-275	45-905	96-235	66-115	89-267	70-426	70-426	105-332	109-153	139-295	235-1140	
Ni	2079-3670	2318-6764	1775-3251	1545-2252	479-3444	2373-4140	2121-2715	1612-3839	1612-4763	1612-4763	2055-4467	1870-2527	1154-8395	949-10686	
Cu	4.5-55	4.9-424	8-699	1.3-56	14-1360	78-381	82-333	36-285	21-270	21-270	24-386	4.1-157	1.4-76	12-1023	
Zn	22-90	25-125	3.7-7.7	4.2-10	5.8-126	22-90	18-45	62-198	17-321	17-321	6.5-65	7.5-20	891-1499	1812-2758	
Ga	0.31-0.50	0.25-4.6	0.22-0.53	bdl	0.47-3.2	0.44-1.0	0.35-0.86	0.48-1.5	0.24-1.2	0.24-1.2	0.84-4.3	0.31	5.2-10.60	34-127	
Ge	1.1-2.0	1.3-3.1	1.4-2.3	0.62-1.3	1.0-1.6	1.1-2.9	0.86-1.4	1.2-3.3	1.0-1.8	1.0-1.8	0.36-1.4	0.76-1.6	0.24-2.2	1.4	
As	2.7-5.8	4.4-9.3	5.1-12	1.4-6.9	1.3-5.2	19-31	14-19	1.5-5.3	1.4-3.4	1.4-3.4	3.9-24	3.4-5.6	1.4-8.4	bdl	
Mo	bdl	bdl	0.51	bdl	bdl	bdl	bdl	0.35-0.44	bdl	bdl	bdl	bdl	bdl	bdl	
Ag	bdl	bdl	0.23-0.89	0.48	bdl	0.21-0.24	bdl	0.16-0.20	0.24	0.24	0.26	bdl	1.3	bdl	
Sn	0.45-0.70	0.40-0.85	0.38-1.1	0.22-0.79	0.36-2.8	0.20-0.37	0.27-0.37	0.22-0.64	0.26-0.48	0.26-0.48	0.47-0.81	0.28-0.74	0.36-1.8	0.68	
Sb	bdl	0.14-0.37	0.13-1.1	bdl	bdl	0.26-0.73	0.22-0.54	0.12-0.19	0.15-0.17	0.15-0.17	0.18-0.91	bdl	0.42-2.5	1.0	
W	0.32-0.68	0.29-0.72	4.3-7.6	1.2-3.9	0.29-1.7	0.49-3.7	0.41-0.72	0.18-7.3	0.23-0.44	0.23-0.44	0.83	bdl	1.2-2.8	3.4	
Au	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	
Pb	0.49-0.54	0.32-1.4	0.32-0.85	0.27-1.9	0.23-0.26	0.22-0.47	0.25-0.47	0.44-0.93	0.19-1.2	0.19-1.2	bdl	bdl	0.44-3.0	bdl	
Bi	bdl	0.09-0.31	bdl	bdl	0.09-0.51	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.34-1.4	bdl	

	spl CLUZ6	chl LAU2	chl FLOT1	chl FLOT2b	chl CLUZ6	amph FLOT1	amph FLOT1	amph CLUZ6	Fe ox LAU2	Fe ox FLOT1	Fe-Ni ox FLOT1	Fe ox FLOT2a	Fe ox FLOT2b	Fe sulph FLOT2a
Sc	1.6-3.7	10-17	4.0-7.7	4.4-8.6	4.88-18	18-26	22-54	1.8-14	8.2	bdl	bdl	8.2	2.1	bdl
Ti	1350-6603	143-168	169-415	155-403	196-449	620-678	808-1808	14-60	46-175	46-175	bdl	bdl	11	bdl
V	234-747	149-188	8.7-73	34-96	51-106	33-60	123-182	15-23	5.1-23	5.1-23	bdl	83	13	9.7
Cr	43357-113080	10044-12280	546-11902	1421-11214	1408-9289	415-4665	1623-4740	855	663-4916	663-4916	bdl	12263	99	381
Co	64-159	38-46	44-108	26-57	34-64	33-72	52-101	9.7-145	7.7-23	7.7-23	8516	213	764	2127
Ni	199-442	1025-1270	852-1443	494-1208	573-946	586-902	737-1481	238-3004	81-323	81-323	140957	1906	15878	79693
Cu	1.4-47	1.2-7.6	7.0-23	3.49-70	4.6-141	3.0-24	14-176	5.1-25	1692-5508	1692-5508	85822	94	4747	2622
Zn	1178-2361	19-35	6.7-46	13-46	39-74	22-41	61-108	4.7-70	145	145	2606	90	1624	55
Ga	6.5-15	12-16	1.2-13	5.3-15	9.8-20	1.7-6.6	10-30	1.1	3.1	3.1	27	20	11	4.7
Ge	0.26-0.39	1.3-5.1	0.55-1.3	0.33-0.80	0.40-1.1	1.5-1.9	1.5-1.9	bdl	1.1-1.3	1.1-1.3	1.2	2.1	0.81	bdl
As	bdl	9.2	1.3	2.1-4.7	bdl	4.1-5.3	2.5-8.7	bdl	6.7	6.7	232	bdl	206	7.0
Mo	0.34	6.2	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	6.6	bdl	bdl	bdl
Ag	bdl	2.9-3.1	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	177	bdl	4.1	9.2
Sn	0.24-0.73	0.37-5.8	0.31-1.5	bdl	0.31-1.2	1.1-1.6	0.71-1.3	2.3	bdl	bdl	bdl	bdl	bdl	bdl
Sb	bdl	2.7-5.1	0.27	0.18	bdl	bdl	bdl	bdl	bdl	bdl	5.6	bdl	7.1	bdl
W	0.2	5.4	1.8	bdl	bdl	0.74-1.3	0.90-3.4	bdl	3.5-4.0	3.5-4.0	7.5	bdl	7.5	bdl
Au	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	31	bdl	bdl	bdl
Pb	bdl	5.4-5.6	0.33	bdl	bdl	0.26-0.58	0.86	bdl	bdl	bdl	40	bdl	40	40
Bi	bdl	3.3-3.6	bdl	bdl	bdl	0.12	0.12	bdl	bdl	bdl	2.0	bdl	2.8	4.7

bdl: below detection limit

Table 4

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Table 4. Trace element abundances in the mineral assemblage of amphibolites

	amph		amph (± oxidised)		plag		ttn		Fe ox		Fe sulph	
	CLUZ1	CLUZ4	CLUZ4	CLUZ4	CLUZ1	CLUZ4	CLUZ4	CLUZ4	CLUZ1	CLUZ4	CLUZ1	CLUZ4
Sc	9.8-37	12-82	12-35	2.6-5.8	2.6-5.8	1.4-4.7	13		bdl			8.0-12
Ti	264-1536	396-3000	345-1069	9.4-239	9.4-239	221-306	150697-184188	14	bdl			72-361
V	24-149	28-320	36-116	bdl	bdl	3.8-10	225-341	bdl	bdl			25-79
Cr	113-736	28-1775	24-335	bdl	bdl	bdl	113-522	bdl	bdl			30-57
Co	48-65	27-53	40-46	0.29-5.6	0.29-5.6	0.4-8.3	2.1--20	1285	1285			21-7180
Ni	97-148	97-288	224-709	bdl	bdl	28-54	86	7202	7202			114-22078
Cu	1.4-93	6.5-53	9.5-71	2.8-4.7	2.8-4.7	2.2-28	4.1-30	1017	1017			27763-89963
Zn	49-66	25-51	40-59	bdl	bdl	7.6	4.7-34	122	122			160-449
Ga	8.3-13	2.3-17	3.0-6.6	10-21	10-21	13-17	0.79-8.7	bdl	bdl			1.6-6.3
Ge	1.2-2.1	1.5-2.7	1.2-2.4	0.79	0.79	0.98	3.5-5.4	bdl	bdl			bdl
As	bdl	1.3-8.8	1.1-7.2	bdl	bdl	bdl	4.2	9.83	9.83			36
Mo	bdl	bdl	bdl	bdl	bdl	bdl	3.7	bdl	bdl			4.2
Ag	bdl	2.1	0.21	bdl	bdl	bdl	bdl	6.27	6.27			39-99
Sn	0.28-0.97	0.37-1.0	0.22-1.2	bdl	bdl	bdl	4.7-5.6	bdl	bdl			2.0
Sb	0.37-0.39	0.19-0.65	0.17-0.22	bdl	bdl	bdl	0.48	1.08	1.08			4.3
W	2.9	bdl	bdl	0.58	0.58	bdl	16-18	bdl	bdl			bdl
Au	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl			bdl
Pb	bdl	bdl	0.29	0.67-15	0.67-15	0.46-0.70	bdl	7.09	7.09			8.67-28
Bi	bdl	bdl	bdl	bdl	bdl	bdl	0.41	bdl	bdl			1.2-8.5

bdl: below detection limit

Table 5

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Table 5. Trace element abundances in the mineral assemblage of the UHP zoisite-eclogite

	garnet L04-143	omphacite L04-143	zoisite L04-143	rutile L04-143	amphibole L04-143	ilmenite L04-143
Sc	35-65	37-42	18-51	4.3-69	27-59	15-54
Ti	56-13793	255-287	172-230	599343	1020-9465	377586
V	56-132	585-811	353-986	2625-2969	333-556	648-903
Cr	33-281	305-498	166-329	1586-2006	286-394	450-720
Co	34-42	12-20	6.9-19	0.70-27	17-67	62-67
Ni	8.0	67-115	10-25	30	68-240	102-147
Cu	bdl	89	16-17	9.9-13	16-21	bdl
Zn	50-176	57-192	20-46	17-119	81-200	74-106
Ga	7.4-16	28-41	111-186	9.5	21-32	4.8
Ge	2.7-3.4	1.6-2.6	8.8-19	3.6	3-21	6.3
As	bdl	7.5-8.7	13-22	6.3-11	7.8-11	bdl
Mo	bdl	4.7	5.3-15	4.2-8.3	9.0-16	bdl
Ag	bdl	bdl	7.1-11	bdl	3.7-14	bdl
Sn	0.82	1.5-1.6	2.8-20	13-17	1.1-23	3.8
Sb	bdl	1.5-2.3	12-16	5.0-13	1.2-20	3.1
W	5.7	3.1	10-16	111-164	8.6-10	17-85
Au	bdl	bdl	bdl	bdl	bdl	bdl
Pb	bdl	6.1-21	75-98	12	16-39	bdl
Bi	bdl	bdl	5.9-16	bdl	9.6-17	bdl

bdl: below detection limit

Table 6

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Table 6. Trace element abundances in the mineral assemblage of granites and gneiss

	bt	VAU2	bt	VAU6	bt	820A	bt	IGAR744	bt	Ay.MAR	bt	GNEISS-SY-1	bt	LAU10e	bt	Che10b	ms	VAU2	ms	VAU6	ms	820A
Sc	5.8-51	9.3-55	15-19	9.5-16	15-25	26-58	50-58	17-31	55-255	14-24	35-41											
Ti	7976-21870	10293-19763	20053-21382	18693-26499	16176-17570	13261-18489	9905-13156	12883-20178	3873-5036	1744-19323	5771-8918											
V	27-356	3.1-317	123-152	79-127	84-206	132-216	374-447	78-105	10-127	1.4-21	38-146											
Cr	5.0-189	66-160	41-62	37-65	68-285	49-161	246-449	5.7-16	11-17	bdl	33-54											
Co	2.5-35	7.3-19	34-39	20-43	16-21	32-37	29-45	24-36	1.4-2.7	2.0-3.0	1.8-4.3											
Ni	3.0-39	11-20	27-34	15-24	28-39	32-69	84-98	7.6-14	bdl	bdl	bdl											
Cu	1.5-9.3	1.6-42	bdl	bdl	bdl	1.2	bdl	1.4	4.5	2.2-6.6	bdl											
Zn	1956-2902	1883-2527	1784-1914	1667-2375	1699-1985	352-441	389-434	1038-1541	191-234	252-466	92-112											
Ga	67-147	80-139	93-106	89-133	76-93	68-88	141-164	97-141	221-282	157-205	179-192											
Ge	3.1-7.7	3.1-7.0	2.5-4.4	4.7-7.1	4.4-5.7	2.1-3.3	3.1-4.1	2.6-5.3	3.4-5.2	3.8-7.2	2.0-4.5											
As	3.7-5.8	4.2-5.3	4.9	bdl	bdl	17	bdl	9.0-10	bdl	11	bdl											
Mo	bdl	1.2	bdl	bdl	bdl	bdl	bdl	1.7	bdl	2	3.2											
Ag	bdl	bdl	bdl	bdl	bdl	1.5-3.3	bdl	bdl	bdl	bdl	3.1											
Sn	238-610	200-359	55-62	115-167	168-202	22-32	11-13	52-100	202-267	196-425	92-98											
Sb	bdl	bdl	bdl	bdl	bdl	0.24-0.57	bdl	1.2-2.3	bdl	bdl	1.4-3.4											
W	8.7-38	14-77	1.9-3.7	5.1-12	7.1-11	4.0-15	2.4-4.4	1.9-3.2	80-107	128-193	20-40											
Au	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl											
Pb	2.5-8.6	2.6-6.8	3.4-5.5	4.5-8.4	4.3-5.9	1.5-8.6	6.6-9.6	4.4-11	2.2-4.2	2.1-4.7	5.7-14											
Bi	bdl	0.53-2.5	bdl	bdl	0.44-0.84	0.10-0.11	bdl	0.9	bdl	2	3.4											

	ms	IGAR744	ms	Ay.MAR	ms	GNEISS-SY-1	ms	Che10b	ms	GNEISS-SY-1	ms	ab	ms	GNEISS-SY-1	ms	ap	ms	GNEISS-SY-1	ms	qtz	ms	Che10b	ms	Fe-Ti ox	ms	LAU10e
Sc	34-52	41-87	12-86	52-72	6.1-7.1	3.6-4.5	bdl	5.5-8.8	12	5.5-8.8	5.5-7.5															
Ti	5129-7281	3914-5082	417-6312	5243-6509	29	11-117	170	21-512	395	21-512	53470-72894															
V	59-80	34-113	50-153	82-115	bdl	0.96	bdl	0.49-4.2	6.1	0.49-4.2	30-70															
Cr	8.4-16	11-127	87-109	6.3-6.9	bdl	27	bdl	bdl	11	bdl	20-28															
Co	1.2-2.0	1.1-1.5	1.1-1.6	0.69-2.4	bdl	0.33-0.37	bdl	bdl	0.91	bdl	0.19-0.23															
Ni	bdl	bdl	5.3	bdl	bdl	bdl	bdl	bdl	18	bdl	4.1-7.1															
Cu	bdl	bdl	2.1	bdl	bdl	1.7	4.2	bdl	18	bdl	0.53-0.92															
Zn	78-110	112-137	12-23	58-123	bdl	4.7	8.1	7.0-8.4	114	7.0-8.4	8.2-9.1															
Ga	135-188	157-184	61-138	189-221	62-64	23-25	1.4	0.36-1.9	2.8	0.36-1.9	4.4-5.4															
Ge	1.8-3.2	1.8-3.3	1.6-1.8	1.3-2.9	1.5-1.7	1.7-2.2	2.6	0.96-1.5	bdl	0.96-1.5	1.5-2.3															
As	bdl	bdl	3.1-3.2	bdl	bdl	bdl	12	bdl	bdl	bdl	2.8-4.4															
Mo	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl															
Ag	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl															
Sn	71-110	156-214	11-42	41-160	2.1-2.4	0.76-1.3	0.7	0.59-0.90	6.3	0.59-0.90	2.1-4.3															
Sb	bdl	bdl	bdl	bdl	bdl	0.64	4.2	bdl	3.4	bdl	5.0-7.6															
W	59-84	91-152	2.0-126	24-43	bdl	bdl	4.2	bdl	bdl	bdl	87-147															
Au	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl															
Pb	3.1-6.8	2.2-4.4	2.6-3.5	11-14	51-58	18-22	4.4	0.33-0.35	4	0.33-0.35	1.5-2.0															
Bi	bdl	bdl	0.11-0.20	bdl	bdl	bdl	0.58	bdl	bdl	bdl	bdl															

bdl: below detection limit